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Final Report: Heat and Mass Transfer in Unsaturated Porous Media

February 1982

Prepared for Pacific Northwest Laboratory
Seasonal Thermal Energy Storage Program
under Special Agreement B-97620-A-O

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FINAL REPORT: HEAT AND MASS TRANSFER
IN UNSATURATED POROUS MEDIA

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SUMMARY

This document constitutes the final report of a preliminary study of heat and water transport in unsaturated porous media. The project provides background information regarding the feasibility of seasonal thermal energy storage in unconfined aquifers. This document contains two major items: (1) a parametric analysis of the factors of importance, and (2) an annotated bibliography of research findings pertinent to unconfined aquifer thermal energy storage (ATES).

In order to establish the parameters appropriate for study, a comprehensive literature review was made. It was found that flow systems in the temperature range between 50 and 90°C have not been studied; the analysis presented here is therefore in the position of extrapolating existing data to the temperature range of interest for unconfined ATES. This analysis shows that heat and mass transfer of water vapor assume dominant importance in unsaturated porous media at elevated temperature. Although water vapor fluxes are seldom as large as saturated medium liquid water fluxes, they are important under unsaturated conditions. The major heat transport mechanism for unsaturated porous media at temperatures from 50 to 90°C is latent heat flux. The mechanism is nonexistent under saturated conditions but may well control design of unconfined aquifer storage systems.

The parametric analysis treats detailed physical phenomena which occur in the flow systems study and demonstrates the temperature and moisture dependence of the transport coefficients of importance. The question of design of an unconfined ATEs site is also addressed by considering the effects of aquifer temperature, depth to water table, porous medium flow properties, and surface boundary conditions. Recommendations are made for continuation of this project in its second phase. Both scientific and engineering goals are considered and alternatives are presented.

INTRODUCTION

The processes of heat and mass transport in unsaturated porous media are important in unconfined ATEs. Since the unsaturated soil, sediment and rock zone above an aquifer constitutes the upper boundary for flow, the transport properties of that layer must be known. In the case of unsaturated soils and sediments, these fluxes can vary extensively with temperature and moisture. This variation can directly affect both the storage efficiency of an aquifer and the impact of the aquifer on the earth's surface above. Changes in heat and water transport from underlying unconfined aquifer systems may be of environmental concern in the location of ATEs sites.

Shallow, unconfined aquifers are attractive as seasonal energy storage media for a variety of reasons. First, near surface reservoirs do not require extensive pumping systems. Storage at shallow depth is therefore less expensive. An additional benefit of shallow storage is that aquifer monitoring equipment is also less expensive. It is also true that shallow, unconfined aquifers could be created in areas where hydrologic conditions are proper. Aquifer creation is already in practice (e.g., Kemper et al.) and is particularly suited to small scale application.

Several field tests of unconfined aquifer storage of heated water have shown some promise (Cormary et al., 1978; Mathey, 1977; Werner and Kley, 1977). Given this evidence, this project was initiated to provide

background information on the feasibility of unconfined ATES. The specific background presented here is the physics of heat and water transport through unsaturated porous media. In addition to the basic principles, a discussion of other considerations of aquifer heat storage design is included. These specific design features can be broken down into four factors: aquifer temperature, aquifer depth, surface boundary conditions for heat and water transfer, and porous medium transport properties. At a specific location, any of these factors may be adjusted in order to optimize storage efficiency for a given situation. Each factor has a range over which it might be adjusted in order to achieve efficient storage but the range and feasibility of adjusting the factors will vary by site.

The analysis presented here treats aquifer temperatures in the range 5 to 95°C and specifically treats flow in unsaturated regions. This complicates the flow system by introducing two phase mass transport and introducing the question of temperature and moisture dependent properties. While liquid water flow in porous media is known to be strongly moisture sensitive in the unsaturated zone, vapor flux is known to be quite temperature sensitive. Since the importance of vapor flux is greatest at high temperatures, the complexity of two phase flow in a varying temperature and moisture field cannot be ignored as temperature is elevated. In fact, temperature effects make storage of high temperature water more difficult than cool water storage; this report therefore dwells on temperatures greater than 45°C.

Selection and Organization of Topics. The organization of this report was determined by analysis of the important components of heat and water transport from unconfined aquifer to the atmosphere. A survey

of the literature indicates that the flow system should include heat and water fluxes in saturated and unsaturated porous media as well as exchange with the atmosphere at the soil surface. Often, however, these system components are treated separately for various specific applications. In these specific treatments, simplifications of general transport theory are made in order to suit the needs of the problem. While these simplifications may be proper and justifiable, they do make synthesis of the component flow phenomena difficult. In order to put the components of saturated flow, unsaturated flow and soil surface-atmosphere exchange of heat and water in perspective, this report begins with a concise review of appropriate theory. The literature review section is brief since the companion document to this report is an extensive annotated review of pertinent literature. This document therefore includes only a brief review of research on temperature dependent properties of soils, transport coefficients for heat and water, atmospheric exchange and field scale unconfined ATEs experiments. The next section treats in detail the temperature and moisture dependent properties of the solid, gas, and liquid components of the flow system. After these properties are shown, calculations of transport coefficients are made for a variety of situations. Analysis concludes with calculations of gross effects on heat and water transport of aquifer temperature, aquifer depth, flow properties of porous media, and atmospheric conditions.

THEORETICAL BACKGROUND

INTRODUCTION

Transport of heat and mass through porous media has long been studied by scientists and engineers in order to serve both theoretical and applied goals and interests (e.g., Bird, Stewart and Lightfoot, 1960). While extensive study has been made of transport through fluid saturated earth materials by geologists and petroleum engineers, transport through unsaturated porous media has generally been treated by soil physicists and agricultural engineers. As a result of differing primary objectives of various studies and various groups of researchers, transport in saturated and unsaturated porous media have often been treated in different fashions. The differences are not fundamental; they arise from simplifications of or assumptions about basic theory. For cases such as thermal energy storage in near surface aquifers where both saturated and unsaturated conditions will be encountered, a relatively general approach must be taken to insure that both conditions can be treated. Fluid flow in saturated-unsaturated systems has been studied (Bear, 1972; Corey, 1977) and the status of both physical and numerical modeling is advanced. Studies of heat transport from aquifer systems has been investigated (e.g., Sophocleous, 1979) but the status of physical and conceptual modeling is not as advanced as in fluid transport.

An additional complication for modeling heat and water movement in aquifers is that heat flow strongly affects water flow and vice versa; the two are said to be coupled. Furthermore, the treatment of heat and

water movement in unsaturated porous media involves four components: liquid water, water vapor, sensible heat and latent heat. In the following sections, the mechanisms and mathematics of heat and water transport in porous media will be discussed in order to provide a framework for evaluating the importance of unsaturated heat and water transport on aquifer thermal energy storage. A necessary and complex component of this treatment will be the atmospheric boundary condition at the earth's surface.

MECHANISMS OF WATER TRANSPORT

Water moves through porous media in two ways: diffusion along a free energy gradient and by convection in another fluid. For most cases in soils, liquid water and water vapor are the phases in which flow occurs and convective transport of water vapor occurs in air.

In most studies of soil water movement, transport by convection is neglected. This simplification is justified in most instances for two reasons. First, the amount of water vapor in air is quite small (the saturated water vapor density of air at temperatures less than 50°C is less than 85 g m^{-3} (List, 1971). Second, air flow in soils is generally quite small. Scotter, Thurtell and Raats (1967) and Kimball and Lemon (1971) present data to show that air movement is negligible except near the soil surface where it may be quite important. Although convective transport of water vapor is probably negligible at depth in soils, it may well be an important component in water exchange between the surface soil and the atmosphere. Also, at elevated temperatures where the saturated vapor density of air is appreciable, convective transport will be greater than at common soil temperatures.

Liquid and vapor flow of water on a gradient of free energy or total water potential accounts for the bulk of water flow in soils. In the absence of water potential effects due to osmotic forces, the major component water potentials driving liquid water flow are the matric or pressure potential and the gravitational potential. Matric (or negative pressure) potentials are important for unsaturated conditions. This potential arises from adsorptive forces binding water to solid soil particles as well as surface tension forces acting at air-water-solid interfaces in the soil pore space.

In soils at saturation or below a water table, the matric potential is zero and pressure potential is either zero or positive. The matric and pressure potentials comprise a continuum of free energy from negative values in unsaturated soils to positive values for soils at saturation or under water. The driving force for water flow is continuous across a saturated-unsaturated boundary.

The gravitational potential is an important driving force for water flow under most saturated conditions. It is also important under unsaturated conditions where large matric potential gradients do not overshadow it. Since hydraulic conductivity decreases rapidly with decreasing saturation, gravity driven water transport is most important in soils near saturation.

Vapor transport in soils is most important under those unsaturated conditions where a significant amount of pore space is filled by a continuous air phase. It is best described as diffusion on a vapor density (i.e., concentration) gradient.

MECHANISMS OF HEAT TRANSPORT

The three mechanisms by which heat is transported in soils are conduction, convection, and radiation. Of these, radiation is generally considered to be least important and is often neglected. Radiation is, however, a very important heat transfer mechanism at the soil surface and may also be important for transport in large soil pores (deVries, 1971).

Conduction of heat in soils occurs through the solid, liquid and gas phases. Since these phases have differing heat capacities and thermal conductivities and since the spatial arrangement of these phases is very complex, calculation of average properties of a soil is difficult (deVries, 1963). The mathematics of heat conduction are well known (Simonsen, 1975) and often, heat transfer in soils is treated as conduction without regard for the transport mechanisms which actually occur.

Convective heat transport occurs in both vapor and liquid water flows. In liquid flows, convection may be separated into two components, forced and free. Forced convection is that which occurs when external forces move the carrying fluid. A common example of this in soils is irrigation (Wierenga, Hagan and Nielsen, 1970). Free convection occurs on a fluid density gradient which causes flow without any applied external force. The density gradient is established by different temperatures at various points in the fluid. Studies of this phenomenon are numerous (e.g., Masuoka, 1978; Wooding, 1963) because the problem is important for both heat storage and fluid mixing.

Convective heat transport in the vapor phase is divided into transport of sensible and latent heat. Of these, latent heat transport is of primary importance since the sensible heat capacity of soil air is

very low (Montieth, 1973) and soil air flow is minimal (Kimball and Lemon, 1971). Latent heat transport occurs when a fluid changes phase and is transported. Since the phase change requires addition or subtraction of an amount of thermal energy called the heat of vaporization or the heat of fusion, this energy is transported with the fluid. In soils, latent heat convection can occur when the gas phase is stationary but a constituent, water vapor, is moving on a vapor density gradient.

COUPLING OF HEAT AND MASS TRANSPORT

Neither heat nor mass transport in porous media can be fully treated without consideration of the inter-relations of both transport phenomena. For the medium treated in this report, moist soils, the transport mechanisms are of operational as well as theoretical importance. Heat and mass transport are directly coupled in heat transport by convection. The coupling arises because a moving water or air and water vapor phase has both sensible and latent heat capacity. If water is transported, heat must also be transported.

Indirect coupling occurs in mass transport. The driving forces for both liquid and vapor transport of water are functions of temperature. In the next section it will be shown that it is convenient, both mathematically and for measurement purposes, to separate and identify water transport on temperature gradients.

Two other very important areas of indirect coupling are the effect of soil water content on specific heat and thermal conductivity and the effect of temperature on water transport properties. These coupling

phenomena have been studied for calculation of heat and mass transport in soils and aquifers at elevated temperatures and are known to be very important considerations (Jackson, 1963; Jury and Letey, 1979; Sepaskah and Boersma, 1979). In fact, the changes in magnitude of various mechanisms of heat and water transport at temperatures above 60°C make the behavior of such flow significantly different from flow at normal earth surface temperatures.

FORMULATION OF EQUATIONS

The study of coupled heat and mass transport in soils has proceeded along both theoretical and applied paths. Procedures to calculate heat or water flow have been developed from theoretical considerations but often, calculation procedures have been developed by discovering relationships between observed fluxes and presumed gradients. Both approaches have merit and should contribute to enhanced understanding of these transport systems. It is true, however, that most approaches ultimately require experimentally determined parameters in order to give satisfactory estimates of transport (Jury and Letey, 1979).

Three scales have been used in studies of transport in porous media: molecular (Sposito, 1978), microscopic (Raats, 1975) and macroscopic (Dybbs and Schweitzer, 1973; Nakano and Miyazaki, 1979; Philip and deVries, 1957; Taylor and Cary, 1964). Of these, the macroscopic approach is most often used. This is in part due to the marked heterogeneity encountered at either the molecular or microscopic scales. It must be noted, however, that heterogeneity is also a problem at the macroscopic scale. The two major approaches which have been used for derivation of macroscopic transport equations are classical engineering

continuum mechanics (e.g., Philip and deVries, 1957) and nonequilibrium thermodynamics (Taylor and Cary, 1964). Although the approaches result in different equations and applications of both sets of equations have met with varying degrees of success (Gee, 1966; Jackson et al., 1976; Jury, 1973), a recent study (Jury and Letey, 1979) has provided a useful technique for reconciling the two theories with existing experimental data. Useful derivations and comparisons of the two major theoretical approaches have been presented by Gee (1966) and Jury (1973).

In this report, discussion of heat and mass transport in soils will be treated using the formulation of Philip and deVries (1957). Although their approach is not mathematically or physically rigorous, recent theoretical studies have provided derivation and reasoning to justify their work (Nakano and Miyazaki, 1979; Raats, 1975). For the purposes of this report, it is convenient to present the Philip and deVries theory in summary form, a technique borrowed from Sophocleous (1979). Table 1 shows this summary. In this format the direct coupling of heat and water transport is seen in Equations (9) and (11). Indirect coupling is explicit in the temperature gradient term of the water balance equation (10) and each of the five transport coefficients [(5) - (9)] has both water content dependent and temperature dependent terms. It is of note that, in Equations (10) and (11), these coefficients occur within the temperature and water content gradient terms. Often the coefficients are removed from the gradient terms. Although this is mathematically incorrect, it is done for one of two reasons: (1) the simplification may be made if the temperature or water content dependence is negligible for a specific situation and (2) the approximation is used in order to linearize and solve the non-linear equation.

Table 1. Philip and deVries (1957) theory for coupled heat and water flow in soils.

Total water flux	$q_w = q_l + q_v$	(1)
Liquid water flux	$q_l = -D_{\theta l} \nabla \theta - D_{Tl} \nabla T - g \rho_l / \mu K(\theta) i$	(2)
Vapor flux	$q_v = -D_{\theta v} \nabla \theta - D_{Tv} \nabla T$	(3)
Heat flux	$q_h = -\lambda \nabla T - \rho_l D_{\theta v} \nabla \theta$	(4)
Transport coefficients:		
Moisture vapor diffusivity	$D_{\theta v} = f D_a v \rho_{vs} / \rho_l \partial h / \partial \theta$	(5)
Thermal vapor diffusivity	$D_{Tv} = \bar{T} f D_a v h / \rho_l \partial \rho_{vs} / \partial T$	(6)
Moisture liquid diffusivity	$D_{\theta l} = K(\theta) g \rho_l / \mu (\partial \Psi / \partial \theta)$	(7)
Thermal liquid diffusivity	$D_{Tl} = \gamma \Psi K(\theta) g \rho_l / \mu$	(8)
Apparent thermal conductivity	$\lambda = \lambda_o(\theta) + L D_{Tv}$	(9)
Balance equations:		
Water	$\partial \theta / \partial t = \nabla \cdot ((D_{\theta l} + D_{\theta v}) \nabla \theta) + \nabla \cdot ((D_{Tv} + D_{Tl}) \nabla T) + g \rho_l / \mu \partial K(\theta) / \partial z$	(10)
Heat	$C \partial T / \partial t = \nabla \cdot (\lambda \nabla T) + \rho_l L \nabla \cdot (D_{\theta v} \nabla \theta)$	(11)

$C(\theta)$ = volumetric soil heat capacity, a function of θ

D = molecular diffusion coefficient for water vapor in air

$K(\theta)$ = intrinsic permeability, a function of θ

L = latent heat of vaporization

$\bar{T}$ = temperature

T = temperature gradient correction factor

f = porosity-tortuosity term

g = acceleration of gravity

h = relative humidity

i = unit vector in the direction of the gravitational field

q = flux density (subscripted for water (w), liquid (l), vapor (v) or heat (h))

γ = coefficient of surface tension

θ = volumetric soil water content

$\lambda_o(\theta)$ = true thermal conductivity, a function of θ

μ = viscosity of liquid water

v = mass flow factor

ρ = density (subscripted for liquid water (l) or saturated water vapor (vs)).

Ψ = soil matric potential

The theory presented by Philip and deVries (1957) was developed for application to soils under common agricultural soil water content and temperature conditions. The equations do not consider convective transport of water vapor or sensible heat in the soil air and heat transfer by radiation is neglected. While these assumptions are probably adequate for most situations, these omitted terms may be of greater importance under the elevated temperature regimes which may accompany aquifer thermal energy storage.

BOUNDARY CONDITIONS AND SPATIAL VARIABILITY

The boundary conditions of any flow problem are part of the specification of the flow and are critically important for many cases of heat and water transport in soils. An important boundary condition involves specification of fluxes at the soil surface. For the case of near-surface, unconfined aquifers, depth to and fluctuation of the water table must also be specified. Since the upper and lower boundary conditions are so important, flow in these systems is often treated in one dimension only. This simplification has justification in many systems but often one must then separately consider the question of horizontal spatial variability.

Spatial variations of soil properties. Variation of soil properties in the vertical direction has long been recognized and both analytical (Clothier, Scotter, and Kerr, 1977) and numerical (Seegerlind, 1976) techniques have been developed to account for this variation. Furthermore, artificially induced variations such as surface mulching are used to control heat and water transport in soils (Papendick, Lindstrom and Cochran, 1973). Horizontal variations in soil properties are also

known to exist and extensive measurement schemes have been used to quantify the variation (Nielsen et al., 1973). Although research of this type helps to assess the magnitude of the variation, relatively little work has been done to simplify calculations for these areas. A common approach is to perform calculation on spatial units over which soil properties can be assumed homogeneous. Another recent promising approach involves estimation of a dispersion or diffusion coefficient which is large enough to account for variability in soil properties over a large land area (Bresler and Dagan, 1979). Simmon (1981) and Gelhar, Gutjahr and Naff (1979) also offer innovative treatments of the problem of spatial variability.

Boundary conditions at the water table. The boundary condition for water flow is complicated by the fact that the water table height may be controlled by phenomena which are external to the unsaturated zone. Thus, aquifer characteristics must be known in order to specify this condition. The heat flow boundary condition may also be controlled by aquifer characteristics. This boundary condition is further complicated by the fact that under saturated or near saturated conditions, the importance of heat convection in liquid water increases sharply. In fact, heat flow in saturated porous media is generally described as heat conduction and forced and free convection in the liquid phase (e.g., Masuoka, 1972).

Boundary conditions at the soil surface. The mechanisms of both heat and water movement change in relative importance at the boundary between the atmosphere and the soil. Liquid phase water flow decreases in importance from saturated conditions to the unsaturated conditions which generally obtain at the soil surface of sites where aquifer thermal

storage would be considered. At this boundary, the liquid water phase disappears and water flow from soil to atmosphere is entirely in the vapor phase. Although liquid phase inflow from the atmosphere to the soil as rainfall or dewfall is common, evaporation or plant transpiration are the most common mechanisms of water loss.

At the soil surface, heat transfer by conduction is negligible and radiation and convection are the important components of heat transfer. Because the nature of heat transfer at this boundary changes markedly, the soil surface usually shows the largest fluctuation in temperature of any point in the continuum from water table to atmosphere. The boundary condition can be specified using an energy budget of the soil surface (Montieth, 1973):

$$q_s = LE + R_L + R_s - H \quad (12)$$

where q_s is net heat flux to soil, LE is the latent heat of vaporization times the vapor flux, R_L is net long wave irradiance, R_s is net shortwave irradiance and H is sensible heat flux by convection. Note that coupling of heat and water transport is shown in this equation.

The components of the surface energy budget can be calculated from environmental measurements. Surface and air temperatures are required to calculate the net long wave radiation term from the Stefan-Boltzman Law:

$$R_L = E_s S T_s^4 - E_a S T_a^4 \quad (13)$$

where E is the emissivity of the surface (s) or air (a), S is the Stefan-Boltzman constant and T_s and T_a are surface and air temperatures. Net shortwave radiation can be calculated as:

$$R_s = (1-A)S_I \quad (14)$$

where A is surface albedo and S_I is incoming shortwave radiation. Sensible heat transfer and evaporation can be calculated from (Montieth, 1973):

$$H = \rho c_p (T_s - T_a) / r_h \quad (15)$$

and

$$E = (\rho_{vs} - \rho_{va}) / r_v \quad (16)$$

where ρc_p is volumetric heat capacity of air, ρ_v is water vapor density of the surface (s) or air (a) and r_h and r_v are resistance to heat (h) or vapor (v) transport. Equations (15) and (16) are written in resistance form rather than as differential equations in order to express the fluxes in terms of measureable, macroscopic differences in temperature and vapor density. Use of this form of expression does, however, mean that the resistance terms must be integrated values for the distance between the surface and the measurement position for the air values. In practice these resistances must be experimentally determined. Classical wind tunnel studies have shown the resistances to be of the form (Campbell, 1977):

$$r = C(d/n)^{0.5} \quad (17)$$

where C is a constant, d is a characteristic dimension of the flow geometry and n is windspeed. Resistances calculated from this form of equation are adequate for forced convection situations in which wind shear is the dominant force causing fluid movement. In the free convection situations which occur when the soil surface warms above air temperature due to large radiation loads, other procedures for calculating resistance must be adopted (Campbell, 1977).

The water flow boundary condition at the soil surface is given by Equation (16) and can be specified by the atmospheric vapor density and

the transport resistance. Often, the boundary condition is treated by specifying the actual vapor flux, E .

Since the vapor density at the soil surface is strongly dependent on temperature and atmospheric conditions, it is often eliminated from the flux equation by substituting the Penman transformation and the energy balance equation in order to obtain the Penman equation (Campbell, 1977). The resultant equation is satisfying insofar as it qualitatively shows the effects of windspeed, net radiation, temperature and atmospheric vapor density on evaporation. It is limited, however, in that it assumes an unlimited water supply to the soil surface. In situations where this is not the case, the surface boundary condition must be specified using one of the methods outlined in the previous paragraph.

ADVECTION

As a consequence of this one dimensional treatment of heat and water flow in soils described in the preceding paragraphs, spatial variability of the surface boundary conditions must be considered. Enhanced heat or water transport from local warm or moist areas may be considerably larger than that which would be predicted for the area based on energy balance measurements for a larger area (Philip, 1959; Rosenberg, 1969). The effect can be predicted using theory which treats two or three dimensions (Itier and Perrier, 1978) and may well be an important consideration in establishing aquifer thermal energy storage sites in areas with a hot, dry climate.

REVIEW OF PERTINENT LITERATURE

There is a voluminous body of literature which pertains to the transport of heat and water in soils. Although most reports treat only, certain aspects of the transport processes, the information is quite important both for understanding the transport processes and for determining the relative magnitudes of various system components. The review of literature presented here is quite brief and assumes that the reader already has some understanding of the principles involved. In addition to this review section, an annotated bibliography of other pertinent work is included with this report as an appendix.

The Effect of Temperature on Bulk Soil Properties

The temperature dependence of the change of water potential with water content. This curve, the soil water release curve, is a basic soil property of fundamental importance in determining water movement and storage in soils. The temperature dependence of this curve is often calculated as the product of the coefficient of surface tension and the water potential (Philip and deVries, 1957) but most researchers have found this to underestimate the measured temperature dependence (e.g. Chahal, 1965; Wilkinson and Klute, 1962). Sophocleous (1979) calculated temperature dependence from the equation relating water potential and relative humidity (Campbell, 1977) and obtained better but not perfect

results. Although the theoretical validity of his calculations has recently been challenged (Herkelrath, 1981), it can certainly be said that Sophocleous' scaling of temperature dependence with Ψ/T is a useful numerical approximation. Chahal (1965) and Peck (1960) have presented theoretical arguments and some experimental data to support the hypothesis that the change in entrapped air in the soil solution varies with temperature and therefore should be part of the temperature correction.

Two data sets showing temperature effects on the soil water release curve are reproduced in Figure 1. They show that upon elevation of temperature, less water is retained in a soil at a given soil water potential. In addition, actual versus predicted temperature coefficients are shown for one data set (Chahal, 1965). The data presented in this figure indicate that the effect of temperature on the soil water release curve is appreciable. It is also clear that more work must be done in order to develop a means to predict this relationship.

Temperature effects on soil hydraulic conductivity and diffusivity.

These soil properties and the soil water release curve are the common data used to characterize soil hydraulic properties. The soil hydraulic conductivity is often calculated as:

$$KH = k(\theta)g\rho_1/\mu \quad (18)$$

where KH is unsaturated hydraulic conductivity, $k(\theta)$ is intrinsic permeability (a soil property), μ is the viscosity of water, ρ_1 is the density of water and g is gravitational acceleration. Viscosity and density are commonly used to calculate the temperature dependence of soil hydraulic conductivity. Philip and deVries (1957) used these terms with the surface tension coefficient to develop a temperature

FIGURE 1. Matric Water Potential Versus Soil Water Content for Four Soils and Temperature Coefficient Versus Matric Water Potential for One Soil. Data are from Wilkinson and Klute (1962) (graphs on left) and Chahal (1965) (graphs on right).

dependence for diffusivity. Jackson (1963) has used this theory successfully to calculate temperature dependence of soil water diffusivity (Fig. 2). His data show that the temperature effect is greatest in soil with generally high diffusivities. In other words, temperature effects on soil water diffusivity are greatest on coarse textured soils.

The data of Flocker et al. (1968) are shown in Figure 3 in order to show the magnitude of the differences in each of the commonly measured soil hydraulic properties. The original data of Flocker et al. show a great deal of overlap for the three temperatures. In fact, it may well be that the only significant difference shown is that between the diffusivity values at 10 and 20°C and those at 30°C. The work of Jensen et al. (1970) also shows that temperature effects on diffusivity in soils near saturation may be negligible. For dry soils, Scotter (1975) shows substantial changes in diffusivity due to temperature (Figure 4).

It should be noted that the data for soil water diffusivity discussed above combine both liquid and vapor phase transport. A distinction must be made between these in order to properly account for coupled flow of heat and water. Figure 4 shows data from Scotter (1975) in which vapor and liquid diffusivities are measured separately. These data were used with the theories of temperature dependence described previously to predict liquid plus vapor diffusivity at other temperatures. The calculations demonstrate the validity of the model for soils at low water contents.

Temperature effects on apparent thermal conductivity. According to heat transfer theory, both components of the apparent thermal conductivity

FIGURE 2. Soil Water Diffusivity Versus Temperature for Three Soils. Data from Jackson (1963).

FIGURE 3. Soil Matrix Potential, Hydraulic Conductivity, and Soil Water Diffusivity Versus Soil Water Content at Three Temperatures. Data from Flocker et al. (1968).

FIGURE 4. Soil Water Diffusivity Versus Soil Water Content for Two Soils. The upper two graphs show component diffusivities; the lower two graphs show the effect of temperature on soil water diffusivity. Data from Scotter (1975).

FIGURE 5. Thermal Conductivity Versus Soil Water Content for Three Soils at Two Temperatures and Thermal Conductivity Ratio Versus Water Filled Pore Space for Three Soils. Data from Sepaskhah and Boersma (1979).

for a soil (Eq. (9)) should be temperature dependent (Philip and deVries, 1975). This theoretical conclusion is verified by the work of Sepaskhah and Boersma (1979). Figure 5 shows their measurements for three soils at 25 and 45°C. Their data show a substantial effect: thermal conductivity at 45°C. can be as much as three times the value at 25°C. For completely dry soil and those which were quite wet, they found that increase in thermal conductivity was less than 20 percent. At these water contents, it can be presumed that vapor transport is negligible; thus the apparent and true thermal conductivities are equal. If this is the case, the major factor affecting apparent thermal conductivity is the vapor flow component.

Detailed Studies in Unsaturated Soil Systems

SEPARATION OF COMPONENT FLUXES

Since Philip and de Vries (1957) published the approximate theory for heat and mass transfer in nonisothermal soils, numerous studies have been conducted to test their assumptions and predictions (e.g. Cassel et al., 1969; Fritton et. al., 1970; Gee, 1966; Hadas, 1977; Jackson et al., 1974; Jury, 1973; Kimball et al., 1976). The conclusions of these authors vary but several points have been clarified by their studies:

1. In soils near saturation, liquid water flow dominates transport.
2. In dry soils, vapor flux is at least comparable in magnitude to liquid flux.
3. Calculation of vapor flux from macroscopic theory does not adequately account for the observed magnitudes of these fluxes.
4. Heat transfer, in particular, is strongly affected by vapor flux.

For the study of aquifer thermal energy storage, the above conclusions must be used as guidelines only. At temperatures between 50 and 100°C, it is likely that the magnitude of vapor flux will be considerably greater than those for the temperatures of the studies cited above.

Component diffusivities for mass transfer. One example of component diffusivities for a silt loam soil at 25°C is given in Figure 6 (Gee, 1966). The figure shows the relative sizes of the diffusion coefficients defined in Equations (5)-(8). Observations 1 and 2 listed above are supported by these data. In addition, the figure shows several other important points. Although thermally driven liquid diffusivity increases markedly with soil water content, it is always overshadowed by liquid diffusivity on a water content gradient. Vapor flux diffusivity due to water content differences is shown to be much larger than thermally driven vapor diffusivity in the water content range in which they both occur. This relative importance may be mitigated by two factors. First, thermally driven vapor diffusivity has appreciable magnitude over a wider range of water contents and second, it is likely that this coefficient will be greatly increased at elevated temperatures. A final point to note is that although the vapor and liquid diffusivities for water content gradients are generally considered to be related in similar fashion to those in Figure 6, there is some disagreement as to the approximate water content at which the sum of thermally driven liquid and vapor diffusivity is largest. This optimum is near saturation in Figure 6 but other researchers have found this optimum to be at much lower water contents (Sutor and Novak, 1968).

Apparent thermal conductivity. Studies of heat transfer have seldom attempted to separate true conductivity from thermally driven

FIGURE 6. Four Components of Soil Water Diffusivity Versus Soil Water Content. D_l is for liquid flow on a water content gradient, D_v is for vapor flow on a water content gradient, D_{Tl} is for liquid water flow on a temperature gradient and D_{Tv} is for vapor flow on a temperature gradient. Data from Gee (1966).

vapor transport. For this reason, apparent thermal conductivity values show a marked dependence on soil water content. Measurements also show a strong dependence on temperature (Sepashkah and Boersma, 1979). The inclusion of a vapor transfer dependent term in apparent thermal conductivity necessitates an enhancement factor to make predicted values agree with measure values (Hadas, 1977; Sepaskhah and Boersma, 1979).

Enhancement factors for vapor flow. Philip and de Vries (1957) found that classical diffusion theory was inadequate to account for observed vapor fluxes in soils. They accounted for the discrepancy by using theoretical arguments related to microscopic soil particle, void, and water arrangement to add an effective temperature gradient term ($\bar{T}$ in Eq. (6)). Other workers have found additional correction is necessary to reconcile measured and predicted fluxes. Hadas (1977) and Sepaskhah and Boersma (1979) have attempted to justify the inclusion of additional correction factors on the basis of turbulent air flow in pores. Other microscopic justifications (Jury and Letey, 1979) have been suggested to account for the observed discrepancies but it is becoming apparent that measurements for specific soils are required for exact studies. The work of Jury and Letey (1979) indicates that existing theory is probably adequate for many studies which do not require high resolution.

EFFECT OF SOIL SURFACE CONDITIONS ON TRANSPORT

The relationships between environmental conditions and evaporation rate have been studied extensively and there are a number of theoretical and empirical techniques available for prediction of water fluxes. A common and useful approach is the use of an energy budget of the soil

surface. This technique was presented as Equations (12)-(17). In practice, this technique is often simplified further in order to decrease requirements for collection of meteorological data. The Penman equation (Campbell, 1977) is an example of such a simplification. For situations in which the dominant component of the energy balance is radiation and water supply to the soil surface is not limiting, the Penman equation will give acceptable predictions. In fact, some researchers have found that simplifications of the Penman equation will also yield adequate prediction. Others, however, have found that other components of the energy balance must be included in order to achieve useful predictive ability (e.g. Barton, 1979; Jury and Tanner, 1975). Simplification is especially difficult in cases where water flux in the vapor phase is important. The work of Hanks et al. (1967) demonstrate this well. They studied evaporation from three soils under conditions of constant potential evaporation rate which were applied by either wind or radiation. The wind caused evaporation and surface cooling, while the radiation treatment caused evaporation with surface heating. The wind treatment caused greater evaporation from the soils than the radiation treatment and the authors showed that the reason for the decreased evaporation in the radiation treatment was thermally driven vapor transport downwards into the soil. The water content profiles the authors' measured matched well with those predicted using the theory of Philip and de Vries (1957) and they concluded that, for detailed studies, consideration of both vapor and liquid water fluxes was essential. Fritton et al. (1970) reached the same conclusion in a similar study and Hammel (1979) has developed a surface energy balance approach which allows prediction of heat, liquid water and water vapor movements in dry soils.

It has been widely recognized that, for situations in which considerable mass or heat transport is due to movement of water vapor, the detailed energy balance of the soil surface must be considered. The elevated temperatures and likelihood of enhanced vapor transport which would accompany aquifer thermal energy storage will necessitate careful consideration of the soil-atmosphere interface.

Near surface resistance to heat and water transport. Once it is concluded that a detailed treatment of the surface energy balance is required to assess the effect of environmental conditions on heat and water transport in unsaturated soils, it remains to gather both meteorological variables and the required constants. Two parameters of primary importance are the resistances to heat and mass transport (Eqs. (15) and (16)). These are often difficult to determine but considerable research has been done in this area. Often the resistances are separated into near-surface boundary layer resistance and soil resistance. Boundary layer resistances for forced convection and laminar flow can be calculated from theory relating resistance to windspeed in the manner shown in Eq. (17). This simple theory has recently been validated for soil column studies of water flow under temperature gradients (Aragaki and Shirai, 1975), but most researchers have found that more sophisticated calculations are required for the unstable, turbulent flow conditions which commonly occur at the soil surface (Bailey et al., 1975; Hammel, 1979).

A resistance calculation for stable, turbulent flow conditions can be derived by assuming that convective transport takes place across a turbulent boundary layer in which horizontal windspeed increases logarithmically with height (Campbell, 1977; van Bavel, 1966):

$$r_h = \frac{\ln[(z - d + z_h)/z_h] \ln[(z - d + z_m)/z_m]}{k^2 u} \quad (19)$$

where r_h is heat transport resistance, z is height, k is the von Karman constant, u is average wind speed, d is the zero plane displacement, and z_h and z_m are surface roughness heights for heat transport and momentum transport respectively. Hammel (1979) altered Equation (18) to account for free convection from tilled bare soil surfaces. He assumed a zero plane displacement of zero and set the roughness heights for momentum and heat transfer equal. Next, he included a stability correction suggested by Businger (1975) to obtain:

$$r_h = [\ln(z/z_m) + K]^2 / k^2 u \quad (20)$$

K is the stability correction and can be calculated from measurements of temperature and windspeed (Businger, 1975). Equations (18) and (19) have both been used successfully in calculations of evaporation and sensible heat flux (Hammel, 1979; van Bavel, 1966).

Resistance to heat and mass transport in soils is commonly calculated in the form of diffusion coefficients as previously discussed. A primary reason for treating this resistance as part of the surface boundary condition is that considerable work on heat and water conservation by surface mulching has been done in the past (e.g. Greb, 1966; Hanks and Woodruff, 1958; Papendick et al., 1973) and those studies have shown that the heat, liquid and vapor transport properties of mulch layers and well aggregated surface soils may be substantially different from those of the underlying soils. These surface layers are often dry enough so that vapor phase movement is often the dominant mechanism of water flux. Diffusion of water vapor on either temperature or water

content gradients can be calculated using the theory outlined in Equations (3), (5) and (6), but mass flow of water vapor in air may also be important near the soil surface. Hanks and Woodruff (1958) have shown this to be the case and have also indicated that mulches are more efficient in conserving moisture if the voids in the mulch are small. Papendick et al. (1973) have corroborated this point and stated that conservation of moisture requires balancing the decreases of liquid water flow in a mulch with large voids against the enhancement of vapor transport in those large pores. Theoretical studies have also shown the importance of mass flow near the soil surface (Kimball and Lemon, 1971; Scotter et al., 1967) and studies of heat transfer have cited turbulent transport of water vapor in large voids to explain anomalous values of apparent thermal conductivity (Hadas 1977; Sepaskhah and Boersma, 1979).

It is clear that the calculation of surface resistances to heat and mass transport is a difficult task. It is, however, an essential one since both liquid and vapor fluxes are strongly affected by both the soil and boundary layer resistances to transport.

ADVECTION OF HEAT AND WATER

The horizontal distribution of sources and sinks of water and heat gives rise to the phenomenon of heat and water advection to any given point on a surface. In a one-dimensional (vertical) treatment of heat and mass transfer, net lateral fluxes must be accounted for if they are sizeable. Although little work has been done in this area, it may well be an important consideration for aquifer thermal energy storage because surface losses of heat or water may be considerably enhanced by advected dry or cool air. A recent study of advection effects on the surface

energy budget of a small plot irrigated with warm water has shown sizeable sensible and latent heat losses (Cook and Norman, 1977). Their calculations showed that evaporative flux density for a small plot would approximately equal that from a large area if the atmospheric boundary layer was unstable (i.e. soil surface warmer than air). Under stable conditions where forced convection is the dominant transport mechanism, small plot evaporative fluxes will overestimate flux from a large area. This effect of advection would influence the efficiency of an aquifer thermal energy storage facility and must be considered to properly estimate surface boundary conditions as well as to interpret scaled model predictions.

Field Scale Experiments in Heat and Mass Transfer

Studies of coupled heat and mass transfer which are pertinent to aquifer thermal energy storage can be grouped in four categories: aquifer thermal energy storage, application of heat and water to soils; evaporation from shallow water tables, and sensing of aquifers from surface temperature measurements. Three of these have only limited applicability to aquifer thermal energy storage but some useful data regarding heat and mass transport have been developed in these areas. Consideration of these studies may well aid in identifying the important soil characteristics which affect heat and mass transport.

EVAPORATION FROM SOILS WITH SHALLOW WATER TABLES

Although evaporation from shallow water tables has been studied using theoretical models or laboratory experiments, relatively few field

studies have been conducted. Laboratory and theoretical experiments have shown that evaporation is a function of environmental conditions only when a water table is within a critical distance of the surface. The critical distance is a function of soil hydraulic properties and layering (Hadas and Hillel, 1972; Saxena et al., 1971). When a water table occurs below the critical distance, evaporation may reach a steady soil-limited rate, but often, a surface dry layer will gradually develop and evaporation rate may decrease with time (Hadas and Hillel, 1972).

Results from a long term study of evaporation from sands with shallow water tables (Hellwig, 1978) have shown more detailed diurnal trends. Hellwig found that there are two relative maxima in evaporation rate for each diurnal cycle. One occurs near midday in response to environmental demand while the other occurs shortly after sunrise and the coolest soil surface temperatures of the day. The midday maximum is greater than the early morning peak in situations where the water table is shallow enough to keep the soil surface perpetually moist. In situations where the soil surface dries, the early morning maximum is larger than that at midday. The author attributes the early morning evaporation peak to water vapor flow on a temperature gradient. The effect noted by Hellwig may well be enhanced in the case of near-surface aquifer thermal energy storage.

SENSING OF SHALLOW AQUIFERS USING SOIL SURFACE TEMPERATURE

Recognition of patterns of heat transfer characteristic of shallow aquifers has been used to map water table location and fluctuations (Hunter, 1978). The technique requires considerable analysis since there are numerous other factors affecting soil surface temperature

changes. Nevertheless, the fact that detection of water tables is possible indicates that heat and mass transfer from shallow water tables is sufficient to affect surface temperatures. As yet, no studies have calculated either heat or water flow from surface temperature variations.

APPLICATION OF HEAT AND WATER TO SOILS

These studies are significantly different from most soil warming studies because both water and heat are added. In studies where heat pipes are placed in soil without frequent irrigation (e.g. Sepaskah et al., 1973) the soil near the heat pipes dries and apparent thermal conductivity becomes very low. When water is also added to soil, coupled heat and mass transfer occur in much the same manner as above a water table. Two very useful studies have recently contributed to the body of knowledge about heat and mass transfer in elevated temperature regimes. Elwell et al. (1978) studied irrigation and warming of greenhouse beds for several soil materials and found that suitable plant temperature and moisture environments could be maintained with 40°C. buried heat pipes and a shallow water table in the beds. They showed that 30 percent of greenhouse heat requirements could be supplied with the soil warming system and also provided useful data comparing transport properties of three soil types. The surface heat and water fluxes they observed indicate that surface losses from shallow, warm water tables may be appreciable.

Another comparable study performed on small field plots (Cook and Norman, 1977) showed large heat and water losses from well irrigated, heated soils. Surface and system energy balance calculations showed that 90 percent of the heat supplied to the soil was lost through the

surface. Sixty-five percent of the heat was lost through evaporation and accounted for a 40 percent increase in evaporative water loss.

UNCONFINED AQUIFER THERMAL ENERGY STORAGE STUDIES

Several field tests of heat storage in unconfined aquifers are documented in recent reports. Three of these are of particular interest because they give estimates of surface heat losses and storage efficiency (Cormary et al., 1978; Mathey, 1977; Werner and Kley, 1977). All three experiments stored water within 3 meters of the soil surface and thus were subject to effects of surface environmental conditions. Unsaturated soil heat and water transport properties are not given in any of the experiments so detailed knowledge of heat and water transport through the unsaturated zone cannot be quantified. Cormary et al. (1978) do estimate that 30 percent of the stored heat was lost through the upper boundary of the aquifer. It should be noted that the water table in their experiment rose from 3 meters depth to 1 meter depth during the course of their experiment. As mentioned previously, water transport is generally governed by external environmental conditions when a water table is this close to the soil surface. This is an undesirable condition for efficient storage of heat and it is aggravated by the fact that aquifer temperatures are highest near the water table.

Mathey (1977) presents temperature fields for a thermal energy storage and recovery system which was operated through the summer months. His calculations of heat storage show a large loss of heat from the zone above the water table during the storage period and he concludes that proximity of the water table to the soil surface (3 meters for his study) is a major factor in the low storage and recovery efficiency he

observed. Data to calculate the potential heat loss to the soil surface are not given.

Werner and Kley (1977) present data for an aquifer thermal energy storage experiment in which they conclude that surface heat loss is relatively unimportant. Their aquifer lay within 0.75 meters of the soil surface and the aquifer maintained soil surface temperatures near 15°C. in an area initially covered with snow. The gradient in temperature from the water table to the soil surface was approximately equal to the lateral gradients away from the heated water injection point so the authors concluded that the proximity of the soil surface to the water table did not substantially affect heat flux away from the heat storage area.

It is important to note that aquifer thermal energy storage experiments are making use of sites with very shallow water tables. For these cases, heat and mass transport will be closely coupled and proper analysis of the thermal energy balance will require treatment of convective transport of both sensible and latent heat. Storage efficiency is relatively low in these unconfined aquifers but the reason for excessive heat loss is not known. Some authors (Kemper et al.) suggest that efficiency will increase if the surface to volume ratio of the storage volume is decreased as much as possible but the problem in geometry is further confounded by the fact that a large portion of the heat stored in an aquifer resides near the upper boundary of the aquifer. This makes control of heat and water transport through the unsaturated zone particularly important.

EFFECT OF TEMPERATURE AND MOISTURE ON TRANSPORT PROPERTIES OF SOILS

In a previous section, the physical mechanisms of transport of both heat and water were shown to be coupled in several ways. The equation describing heat flow (Eq. (11)) shows direct coupling of heat and water transport as convection terms. In addition, indirect coupling is evidenced by heat transfer driven by a water content gradient (Eq. (11)) and water transfer driven by a temperature gradient (Eq. (10)). A second indirect coupling of major importance is the temperature and moisture dependency of the heat and water transport coefficients described in Equations (5), (6), (7), (8) and (9). The variation of these properties with temperature may be extremely important for aquifer thermal energy storage since elevating groundwater temperatures may markedly increase water and heat flux through unsaturated boundaries of aquifers. In particular, vapor phase transport of water (and latent heat) may become appreciable at the temperatures encountered in these systems. Table 2 summarizes the dependencies which must be accounted for in order to analyze coupled flow in unsaturated soils. This listing shows that most of the soil constituents are temperature dependent. The marked moisture dependence of bulk soil properties is a result of the range in water and water vapor contents normally encountered in these porous media.

Table 2. Temperature (T) and moisture (M) dependent properties of soil constituents.

Symbol	Name	Dependencies	Equation
WATER VAPOR			
D_a	Molecular diffusion coefficient	T	(5), (6)
L	Latent heat of vaporization	T	(5), (6)
h	Relative humidity	T, M	(5), (6)
λ_v	Thermal conductivity	T, M	
c_{pv}	Specific heat	T	
ρ_{vs}	Saturated vapor density	T	(5), (6)
v	Mass flow factor	T	(5), (6)
LIQUID WATER			
γ	Coefficient of surface tension	T	(8)
ρ_l	Density	T	(7), (8)
μ	Viscosity	T	(7), (8)
λ_l	Thermal conductivity	T	
C_{pl}	Specific heat	T	
SOIL PARTICLES			
λ_s	Thermal conductivity	T	
C_{ps}	Specific heat	T	
BULK SOIL			
ψ_e	Air entry water potential	T	
f	Porosity-tortuosity term	M	(5), (6)
$\partial\psi/\partial\theta$	Soil water release curve	T	
λ_o	True thermal conductivity	T, M	(9)
C	Volumetric heat capacity	T, M	(11)
$\bar{T}$	Temperature gradient correction	M	(6)

EFFECTS ON SOIL CONSTITUENTS

Properties of water vapor. Since the vapor density of air at normal earth surface temperatures is several orders of magnitude lower than liquid water, one might expect that it is insignificant for calculations of heat or mass transfer in soils. This is not the case; a number of water vapor properties are strongly temperature dependent in the range 0-100°C. The effect of various properties on the transport coefficients for coupled heat and water flow are shown in Equations (5), (6) and (9). Of the properties listed, the greatest temperature dependencies are found for vapor density and the slope of the curve relating vapor density and temperature (Fig. 7). At 50°C, these quantities are 6 and 12 times their 0°C values but at 90°C the increases are 87 and 48 times, respectively. These increases are significant since both effect proportional increases in the moisture and thermal vapor diffusivities. The density increase also affects heat capacity and thermal conductivity markedly. Figure 7 serves to indicate that, at temperatures above 45°C vapor flow becomes increasingly important.

Figure 8 shows three other temperature dependent properties of importance. Although the changes with temperature are not as marked as those in Figure 7, a proper treatment must include them. The values for the free air diffusion coefficient are a subject of some debate. A number of sets of values exist and there is apparent disagreement about what data and what measurement techniques are appropriate for this coefficient (Philip and deVries, 1957). The differences among the data sets are, however, generally less than 10 percent and the coefficient varies 170 percent from 0 to 90°C. Furthermore, most data sets fit the equational form commonly used for diffusion coefficient calculations

FIGURE 7. Saturation Vapor Density Versus Temperature and the Slope of the Saturation Vapor Density-Temperature Curve Versus Temperature

FIGURE 8. Temperature Dependent Properties of Water Vapor

(Vargaftik, 1975). The data set used for Figure 8 and a suggested equation to calculate the free air diffusion coefficient are shown in Table 3.

Another important factor to note in Figure 8 is the temperature dependence of the mass flow factor. Although it is justifiably neglected for most agricultural studies (Philip and deVries, 1957), it is an important consideration for high temperature gas flows.

The ratio of actual to saturated vapor density, the relative humidity, is shown in Figure 9. It is both temperature and moisture dependent but its value changes little over the range of temperatures and moistures commonly encountered. It is of importance in very dry soils and is often quite low at the soil surface. A theoretical equation relating relative humidity, total water potential and temperature is given in Table 3.

The specific heat and thermal conductivity of both dry air and vapor saturated air are shown in Figures 10 and 11. Both are seen to be much smaller than the values for other soil constituents but they do exhibit notable temperature dependence. Although thermal conductivity of vapor saturated air almost doubles from 0 to 100°C, specific heat at 100°C is one hundred times that at 0°C. Much of the increase in specific heat occurs at temperatures greater than 75°C.

It must also be noted that latent heat transfer with water vapor is a very important phenomenon in unsaturated porous media. Although the energy transport phenomenon is not true conduction, it is often treated as such since it is proportional to a temperature in the same manner as heat flow by conduction. This aspect of heat transfer will be discussed in a subsequent section on bulk soil apparent thermal conductivity.

Table 3. Calculation equations for temperature (0-100°C) and moisture dependent properties. (Units: T-C, Ψ - cm head, R - 4693.5 cm head/C).

Parameter (units)	Data Source, Eq. Source	Equation, Correlation (r^2), Standard Deviation (s)
GAS		
Diffusion coefficient (cm ² /s)	1, 10	$D_a = 0.212((T+273)/273) + 1.808$
Latent heat of vaporization (J/g)	7, 11	$L = 2501.4 - 2.373E-3T$, $r^2 = .99995$, $s = 0.334$
Relative humidity	3, 3	$h = \exp[\Psi/(R(T+273))]$
Thermal conductivity of vapor (mW/cmC)	2, 11	$\lambda_v = 0.1584 + (8.154E-4 + 1.7473E-7T)T$, $r^2 = .99999$, $s = 3.44E-4$
Thermal conductivity of dry air (W/cmC)	6, 11	$\lambda_a = 2.42E-4 + 7.56E-7T$, $r^2 = .99989$, $s = 3.50E-7$
Saturated vapor density (g/m ³)	5, 4	$\rho_{vs} = 1323 \cdot \exp[17.27T/(237.3+T)]/(T+273)$
Slope of vapor density curve (g/m ³ C)	5, 4	$\partial \rho_{vs} / \partial T = \rho_{vs} / (T+273) (2.165L/(T+273) - 1)$
Vapor volumetric specific heat (J/cm ³ C)	5, 11	$\rho_c^{pv} = ((8.233E-7T - 1.594E-6)T + 8.307E-4)T + 8.402E-3$, $r^2 = .9996$, $s = 1.24E-2$
Mass flow factor	2, 11	$v = 1/((((-9.575E-9T + 3.42E-7)T - 3.63E-5)T - 1.463E-4)T + 0.993321))$, $r^2 = .99995$, $s = 4.359E-3$

LIQUID

Temperature dependence of surface tension (1/C)	8, 11	$\gamma = 1.8732E-3 + T(1.0315E-5 + T(2.4911E-8 + 1.47E-10T)),$ $r^2 = .99999, s = 0.236E-6$
Thermal conductivity of water (mW/cmC)	2, 11	$\lambda_1 = 5.70042 + .012564T, r^2 = .99999, s = 5.88E-3$
Viscosity/density (cm ² /s)	2, 11	$\mu/\rho_1 = (((3.3194E-10T - 9.1922E-8)T + 9.911E-6)T - 5.516E-4)T + .017781,$ $r^2 = .99995, s = 4.608E-5$
Water density (g/cm ³)	5, 11	$\rho_1 = .99989 + (5.5699E-5 - (7.633E-6 - (4.304E-8 - 1.407E-10T)T)T)T$
Surface tension (g/s ²)	10, 11	$\sigma = 75.65 - (0.1417 + 2.625E-4T)T, r^2 = .99999, s = 0.0245$
Volumetric specific heat of water (J/cm ³ C)	5, 11	$\rho C_{p1} = ((-1.2687E-7T + 1.844E-5)T - 2.313E-3)T + 4.2092,$ $r^2 = .99999, s = 2.35E-2$

SOLID

Thermal conductivity of quartz (W/cmC)	8, 11	$\lambda_q = .0905 - 2.65E-4T, r^2 = .9989, s = 2.91E-4$
Volumetric specific heat of soil minerals (J/cm ³ C)	8, 8	$\rho C_{ps} = 1.85 + 4.29E-3T \quad (0-60^\circ C)$
Volumetric specific heat of muscovite (J/cm ³ C)	9, 11	$\rho C_{pm} = 2.324 + 4.53E-3T, r^2 = .9979, s = 7.66E-3$

-
1. Campbell, 1977 (data for 0-50 C only).	6. Kaminski, 1980.
2. C.R.C., 1979.	7. List, 1971.
3. Edlefsen and Anderson, 1943.	8. Philip and deVries, 1957.
4. Hammel, 1979.	9. Robie et al., 1978.
5. Himmelblau, 1974.	10. Vargaftik, 1975.
	11. regression.

FIGURE 9. Relative Humidity Versus Soil Water Content and Soil Water Potential at Two Temperatures. Silt loam soil.

FIGURE 10. Thermal Conductivity Versus Temperature for Important Soil Constituents

FIGURE 11. Volumetric Specific Heat Versus Temperature for Important Soil Constituents

Two other observations about the heat transfer properties of gases should be mentioned. First, Figure 10 shows that the true thermal conductivity of water vapor is less than that of dry air and second, the specific heat of vapor saturated air (Fig. 11) is significantly larger than that of dry air ($.0013 \text{ J/cm}^3\text{C}$; deVries, 1963).

Properties of liquid water. Figure 12 shows three liquid water properties which are of importance for water transport in soils. Although water retention in soils is physically complex, its temperature dependent variation is often modeled as the variation in surface tension (Chahal, 1965; Wilkinson and Klute, 1962). The data presented in Figure 12 show that surface tension varies 22 percent from 0 to 100°C .

The temperature dependencies of viscosity and density are shown in Figure 12 because they bear directly on the temperature dependence of soil hydraulic conductivity. In Equation (18) this conductivity is separated into fluid and soil matrix properties; the fluid property is the ratio of density to viscosity. The data of Figure 12 show that density varies 4 percent and viscosity 85 percent between 0 and 100°C .

A water property of major importance in heat transfer is specific heat. Figure 11 shows a small temperature dependence (4 percent) but also reveals that water has the highest specific heat of any common soil constituent. Thermal conductivity of water is shown in Figure 10. As temperature increases, water becomes more important since the soil mineral fraction conductivity either decreases or remains constant.

Heat transfer properties of soil solids. The thermal conductivities and specific heats of soil solid constituents are shown in Figures 10 and 11. Each figure has data for soil organic matter, soil minerals and a specific soil mineral. Although there is adequate heat transfer data

FIGURE 12. Temperature Dependent Properties of Liquid Water

for rocks and crystalline minerals (e.g., Kaminski, 1980; Robie et al., 1978), data for soil minerals is much less common. DeVries (1963) reviewed the available data for soil minerals and concluded that the thermal conductivity of all soil minerals except quartz could be assumed to be a single, temperature invariant value. He also recommended that the thermal conductivity and temperature dependence of quartz be accounted for separate from other minerals. The data in Figure 10 show that quartz (at random crystal orientation) has a significantly higher conductivity than other minerals. The value decreases 30 percent from 0 to 100°C. It can be seen from the data presented here that the thermal conductivity of soil solid constituents is much larger than any other component. In a subsequent section, it will be shown that the geometry of the soil solid matrix diminishes the heat transfer effect of these high conductivity materials.

DeVries' (1963) review of specific heat data resulted in a regression equation relating soil mineral specific heat to temperature (see Table 3). Figure 11 shows this regression line as well as data for muscovite (Robie et al., 1978). Muscovite has a significantly higher specific heat than the average line for soil minerals and may be an important constituent of some soils.

There are few data available on the specific heat and thermal conductivity of soil organic matter. DeVries (1963) recommended that both be considered constants.

BULK SOIL PROPERTIES

Temperature and moisture dependent properties of the bulk soil are listed in Table 2 and Equations (5)-(9) of Table 1. Without exception, these are theoretically complex and experimentally difficult to measure.

In this section, pertinent available data have been used to calculate values for various properties and the temperature dependence of these values. As the discussion proceeds, a qualitative assessment of the integrity of the calculations will be made. Table 4 contains the equations used for the analyses presented here.

Porosity-tortuosity term. This term occurs in the vapor diffusion coefficients (Eqs. (5) and (6)) and serves to decrease diffusivity to account for the length and geometry of the flow path. Accordingly, it is generally modeled as a function of a tortuosity factor, total porosity and water content. Figure 13 shows an example of the data required to calculate the porosity-tortuosity term. The figure also includes several curves derived from theoretical considerations (see Lai et al., 1976 for a more complete discussion). It has been suggested, however, that a straight line would fit the data well (Campbell, 1980). Such a regression line intercepts the zero diffusion coefficient line at approximately 10 percent air filled porosity. This intercept would infer that there is a certain minimum porosity at which the air phase becomes continuous through a porous medium. Below this value there is no air (or vapor) flow. This supposition is adopted here and the porosity-tortuosity term is modeled as a linear function of soil water content (see Table 4, Eq. (21)).

Soil water release curve. It has been shown in a previous section that the soil water release curve is temperature dependent. It should be noted, however, that all measurements shown were in the range 0 to -1 bar. In the wet range of soil water contents, there is some temperature dependence but the dependence is too small to measure for drier soils. Chahal (1965) showed that the temperature dependence of

Table 4. Calculation equations for bulk soil properties.

Name (units)	Data Source, Eq. Source	Range	Equation	
Porosity-tortuosity term	7, 1	$0.1-\theta_s$	$f = 0.66(\theta_s - 0.1)$	(21)
Air entry water potential (cm)	6, 4	$0-100^\circ\text{C}$	$\psi_e = 2\sigma/\rho_l g r$	(22)
Change of water content with water potential (cm^{-1})	6, 2	$<\psi_e$	$\partial\theta/\partial\psi = \frac{-\theta_s}{b\psi_e} \left(\frac{\psi}{\psi_e} \right)^{-\theta/(b\psi)}$	(23)
Intrinsic permeability (cm^2)	6, 2	$<\psi_e$	$k(\theta) = \frac{\mu}{\rho_l g} \left(\frac{\psi_e}{\psi} \right)^{2+3/b}$	(24)
Temperature gradient correction factor	8, 5	$0-\theta_s$	$\bar{T} = 2 + 10\theta$	(25)
Volumetric heat capacity ($\text{J}/\text{cm}^3\text{C}$)	3, 3		$C = \rho c_{ps} x_s + \theta \rho c_{pl} + (\theta_s - \theta) \rho c_{pv}$	(26)

¹Campbell, 1980

θ_s = volumetric soil water content at saturation.

²Campbell, 1974

r = effective pore radius.

³deVries, 1963

b = empirical constant.

⁴Capillary rise eq.

x_s = volume fraction of soil solid constituents.

⁵Hammel, 1979

All other symbols are as defined in Tables 1 and 2.

⁶Laboratory measurement

⁷Lai et al., 1976

⁸Philip and deVries, 1957

FIGURE 13. Ratio of Actual to Free Air Diffusion Coefficient Versus Air Filled Porosity. Data for four soils from Lai et al. 1976. Curved lines are based on theory and the straight line is from Campbell, 1980.

surface tension did not account for all the variation in the soil water release curve with temperature but the trend and order of magnitude were correct. In the absence of a better approach, therefore, the surface tension correction will be used to estimate the temperature dependence of the soil water release curve. The correction is applied by adjusting the air entry water potential (Ψ_e) of a soil according to the capillary rise equation (Table 4, Eq. (22)). The initial value of this water potential is taken from a log-log regression of water content on water potential (Campbell, 1974; Corey, 1977). The regression technique also yields the equation for the water release curve (Table 4, Eq. (23)). The measured value of air entry water potential is used with the capillary rise equation (Eq. (22)) to calculate effective pore radius (r) and the equations of Table 3 for surface tension and density are used to calculate air entry water potential values at other temperatures.

Hydraulic conductivity. This bulk soil property is both temperature and moisture dependent. Temperature dependence is commonly calculated by separating intrinsic permeability from the fluid properties as in Equation (18). The temperature dependence is then calculated from the temperature dependence of ρ_l/μ (see Table 3). Moisture dependence can be calculated by scaling the intrinsic permeability with the shape of the water release curve; this scaling procedure infers a relationship between energy status of soil water and its ability to flow. The technique used here is that of Campbell (1974) (Table 4, Eq. (24)).

Temperature gradient correction factor. This term of Equation (6) is a matching factor required to adequately predict vapor flux on a temperature gradient. Hammel (1979) gives an equation (see Table 4) which predicts the factor from soil water content. For a specific study

the factor should be determined by comparing experimental data with calculations, but the published relationship is used here in order to account for the effect in the model systems studied.

Volumetric heat capacity. The bulk heat capacity of a soil can be calculated as the sum of the constituent specific heats of the solid, liquid and gas phases of a soil (Table 4, Eq. (26)). The data of Figure 11 show that the proportion of water in a soil will have a major effect on bulk specific heat. This is demonstrated well in Figure 14. Graph (a) shows that the heat capacity of four soils increases linearly with water content. Although, the zero water content value of specific heat differs according to the volume percentage of muscovite, organic matter and total solid phase, the most important effect is the addition of water. There is 20 percent variation in specific heat at constant water content due to soil solid composition while water content variations account for a 200 percent variation in heat capacity. If the specific heat of the solid phase of a soil were known, the soil heat capacity could be computed as a linear function of water content.

The effect of temperature is also demonstrated in Figure 14 for a muscovite rich soil. It can be seen that the water content effect overshadows the temperature effect, particularly at high water contents. This is a result of the low temperature dependence of the specific heat of water. Figure 15 shows the temperature dependence of heat capacity directly. The effect is important for dry soils (79 percent variation over 100°C) and almost negligible for saturated soils (4 percent variation over 100°C). It is apparent from the data presented here that the bulk heat capacity of soils varies importantly with temperature and water content. In order of importance, the important factors in the

FIGURE 14. Volumetric Heat Capacity Versus Water Content for Four Soils and Three Temperatures

FIGURE 15. Volumetric Heat Capacity Versus Temperature for a Silt Loam Soil at Three Water Contents

calculation of specific heat for a soil are water content, temperature and composition of the soil solid phase.

Bulk thermal conductivity. In order to calculate soil thermal conductivity, the constituent thermal conductivities shown in Figure 10 must be combined. Since heat flow depends on the pathway, the constituent conductivities cannot be simply combined as in the case of specific heat. A well tested technique (deVries, 1963) is used here to calculate bulk soil thermal conductivity from components. The treatment accounts for heat flow by conduction through liquids, gases, and solids and also includes latent heat transfer. This heat transfer mechanism is known to be important under normal soil conditions but little work has been done at elevated temperatures. The experimental work of Sepaskhah and Boersma (1979) at 45°C and deVries (1963) at 75°C are notable exceptions. These authors have found significant contributions of latent heat flux to overall heat transfer (see Fig. 5).

Bulk thermal conductivity is calculated in this report according to the technique of deVries (1963). He suggested formulae to combine solid liquid and gas conductivities and showed that the suggested technique fit experimental data well. His procedure for accounting for heat transfer as latent heat involves modification of the thermal conductivity of air with a vapor flow term:

$$\lambda_v = \lambda_a + h_v \partial \rho_{vs} / \partial T \quad (27)$$

Reference to Figures 7 and 8 shows that this term will be extremely temperature dependent.

The deVries theory used in this report has been modified in accordance with the work of Sepaskhah and Boersma (1979). These authors present values for the empirical coefficients required in the calculations which have been adopted here. In addition, the calculations in this report account for the variations in thermal conductivity of quartz, water and water vapor with temperature. Figures 16 and 17 show the data of Sepaskhah and Boersma (1979) for thermal conductivity as a function of soil water content for two soils at two temperatures. The deVries calculations for three temperatures are shown as well. It can be seen that the theoretical calculations fit the experimental results better for the loam soil but quite well in both cases. It should be noted that the theoretical curves calculated by Sepaskhah and Boersma (1979) fit the data better than those shown here. This is because they adjusted an empirical coefficient (θ_e in their work) for their 45°C calculations. The calculations shown in Figures 16 and 17 use a single value of that constant for all calculations.

Several features of note are shown in Figures 16 and 17. At 25°C, thermal conductivity increases with increasing water content but there is a range of intermediate water contents at which the increase is more rapid than for either wetter or drier soils. It is in this zone that the physical description of the porous medium changes from viewing airfilled pore space as a continuous pathway to viewing the water phase as continuous. Since the water phase thermal conductivity is greater than the gaseous phase, total thermal conductivity also increases. At 45°C, the effect of increasing water content shows slightly different results. Thermal conductivity still increases with water content but vapor phase latent heat flux at drier soil water contents allows thermal

FIGURE 16. Apparent Thermal Conductivity Versus Soil Water Content. Data for loamy sand soil from Sepaskhah and Boersma (1979).

FIGURE 17. Apparent Thermal Conductivity Versus Soil Water Content. Data for loam soil from Sepaskhah and Boersma (1979).

conductivity to increase markedly even before the liquid phase water pathway becomes of dominant importance. The result of the two mechanisms, vapor phase latent heat transport and liquid phase conduction, is a smoothly increasing thermal conductivity. At high temperatures (90°C), latent heat flux assumes such importance that its effect is much larger than the effect associated with increasing cross-sectional area for heat conduction through the liquid phase. In fact, thermal conductivity increases until the air filled pore space is reduced sufficiently to restrict latent heat flux. After this point increased water content decreases thermal conductivity. At saturation the thermal conductivity reaches a temperature adjusted value for the soil-water system which is only slightly higher than the values at 25 and 45°C .

The effect of temperature on thermal conductivity is demonstrated well in Figure 18. A silt loam soil is plotted in the figure and is used in figures which follow because its hydraulic properties are well known. Figure 18 shows that there is only a small temperature effect on this soil when there is no air phase (soil water content = .47). It is interesting to note that, at 90°C , the dry and wet soil thermal conductivities are nearly the same. This is in marked contrast to the situation at 20°C where there is almost an order of magnitude difference in their values. This increase is a direct result of temperature enhanced latent heat flux. Note that the dry soil curve has much the same slope as that of the slope of the vapor density-temperature curve shown in Figure 7. The thermal conductivity at intermediate water content (0.16) shows enhanced thermal conductivity due to vapor transport. The overall conductivity is highest for this curve because there is considerable

FIGURE 18. Apparent Thermal Conductivity Versus Temperature for a Silt Loam Soil

heat conduction through the liquid phase and the relative humidity (which affects vapor transfer) is near 1.

Figure 19 shows the effect of soil properties on the relationship between thermal conductivity, temperature and soil water content. The figure shows the effects of high quartz content, high organic matter content and soil bulk density on thermal conductivity. The reference case is the silt loam soil of Figure 17 at its field bulk density of 1.25 g/cm^3 . The curves at three temperatures for this soil show the trends discussed previously: increased importance of latent heat flux as temperature increases. In fact, the trends hold for all four soils shown. At temperatures of 50°C and below, the relative positions of the curves are the same: quartz rich > high bulk density > reference > organic matter rich. The curves at 0 and 50°C differ in that the effect of latent heat flux at lower moisture contents is enhanced at 50°C . This latent heat flux component accounts for the thermal conductivity maximum at low water contents for the 90°C curves. The maximum appears to be slightly broader for the high bulk density samples and may occur at slightly lower water content. This is expected since higher bulk density is associated with lower total porosity and a smaller water content range where both liquid phase conductivity and latent heat flux are important. Note that at 90°C the thermal conductivity of the organic matter rich soil is less than that of the dense soil under dry conditions, exceeds the dense soil conductivity at its (latent heat flux) peak value, and falls below again at the saturated value.

The predicted thermal conductivity peaks at 90°C are quite large. In fact, they are in the range of solid mineral conductivities. The curves are similar to the experimental observations of deVries (1963)

FIGURE 19. Apparent Thermal Conductivity Versus Soil Water Content for Four Soils at Three Temperatures

and they conform with theoretical expectation. When latent heat flux as water vapor is considered, the apparent thermal conductivity of soils is a marked function of temperature. Soil thermal conductivity maxima may occur at water contents below saturation at temperatures of 70°C or greater (see Fig. 18).

Water transport coefficients. In the nonisothermal water flow theory outlined in Table 1, four transport coefficients are required to completely describe flow. Equations (5)-(8) show the temperature and moisture dependencies which must be accounted for. Equation (24) (Table 4) may be used to calculate unsaturated conductivities from a water release curve and a saturated permeability value. The temperature and moisture dependence of three soils is shown in the next five figures. All temperature dependent coefficients are calculated according to the equations already given. Soil specific data required are a water release curve, a saturated permeability value, and total porosity. Table 5 shows these values for the three soils used in the calculations; the data were obtained in the Oregon State University Soil Testing Laboratory.

The coefficients plotted in the following figures are identical to those of the original Philip and deVries formulation (1957) except for the following modifications:

- 1) All temperatures and moisture dependent terms are accounted for. Even though some parameters change little over the temperature range studied, they are included for the sake of completeness.
- 2) The coefficients for water content driven transport of liquid water and water vapor are cast in the form of water potential driven flows. That is, the coefficients are calculated as

Table 5. Soil hydraulic properties and parameters for equations (23) and (24).

Soil name	b value	Intrinsic permeability, at ₂ saturation, cm ²	Air entry water potential cm	Total porosity
Fine sandy loam	2.81	1E - 7	-8.5	.43
Silt loam	5.1	5.6E - 8	-11	.53
Clay	13.1	6.1E - 10	-50	.56

conductivities and have units of cm/s. This modification removes the term $\partial\Psi/\partial\theta$ from Equation (7) and changes $\partial h/\partial\theta$ to $\partial h/\partial\Psi$ in Equation (5).

The temperature dependence of liquid water flow in saturated soils is shown in Figure 20. The two liquid flow components, water potential and temperature driven, are shown for three soils. Since both terms have temperature dependence related to viscosity, water density and surface tension, the overall dependence of each is roughly equivalent. For each soil, the temperature dependence for thermally driven liquid water flow is slightly greater but no values changes as much as an order of magnitude from 5 to 95°C. It should also be noted that the coefficient for thermally driven water flow is much smaller than that for water potential driven flow. This is magnified even more when one considers that common water potential gradients (in units of centimeters of equivalent head) may be several orders of magnitude greater than temperature gradients (in degrees celsius).

Figure 20 also serves to illustrate the differences between soils. The clay soil has a low water potential driven liquid conductivity and less difference between the two component conductivities than the other two soils. Finally, it can be noted that the difference among soils is greater for the potential driven flows than for the temperature driven ones.

The two vapor flow components show much more marked temperature dependence than do the liquid flow coefficients just discussed. In Figure 21, it can be seen that both vapor flow terms vary three orders of magnitude between 5 and 95°C. Although the rate of increase is approximately constant throughout much of this range, an increase in

FIGURE 20. Liquid Flow Transport Coefficients Versus Temperature for Three Soils. M indicates potential driven liquid conductivity, T indicates temperature driven liquid diffusivity.

FIGURE 21. Vapor Transport Coefficients Versus Temperature for Three Soils

rate of change is exhibited at the highest temperatures shown. The rate of increase is similar for both coefficients and for all three soils but the magnitudes of the components change among soils. Although the potential driven vapor conductivities are almost identical for the silt loam soil and the fine sandy loam, the temperature driven coefficients are different. Unlike the liquid flow coefficients of Figure 20, thermal vapor diffusivity is larger for the silt loam than for the fine sandy loam. Inspection of Equation (6) shows that this must be due to the porosity-tortuosity factor, f . This, in turn, is a function of the total porosity and the water content. Silt loam thermal vapor diffusivity is greatest because it has the largest air filled porosity. The fine sandy loam has the lowest total porosity and the clay soil has the highest water content at $-15,000$ cm water potential. The difference between the potential driven vapor conductivity for the clay soil and that for the other two soils is greater than the difference among the thermal vapor diffusivities because the temperature gradient correction factor does not appear in the potential driven coefficient. Since this factor is largest for the clay because of its larger water content, the difference among coefficients is diminished.

The preceeding two figures serve to illustrate the temperature dependence of the four water flow transport coefficients. The vapor flow terms are much more temperature dependent and increase markedly in overall importance at temperatures near 100°C . It was shown that the temperature dependence of the vapor flow component of heat transport also becomes increasingly important at high temperatures. The differences among soils are as commonly encountered for liquid flow coefficients. Coarse textured soils have higher saturated conductivity than do fine

textured soils. This trend is not noticed for the vapor flow terms these coefficients are largest for the soil with the most air filled pore space.

A comparison of the four transport coefficients for a silt loam soil is shown in Figure 22. It is immediately obvious that the potential driven vapor conductivity is smallest and may be of importance only in very dry soils. Gee (1966) demonstrated that this coefficient may increase markedly at near zero water contents (Fig. 6). Figure 22 also shows the rapid decrease in both liquid flow coefficients with decreasing soil matric potential. Although the thermally driven diffusivity is higher than the potential driven conductivity at $-15,000$ cm water potential water flow is probably less for thermally driven liquid flow because, in field situations, the water potential gradient is generally much larger than the temperature gradient. This is almost certainly true for unconfined aquifer thermal energy storage applications where the operating temperature range is between 0 and 100°C . A final point to note in this figure is that the soil water potential at which the thermally driven vapor diffusivity becomes equal in magnitude to the thermal liquid diffusivity is clearly shown. At water potentials less than -300 cm, the thermal vapor flow coefficient is largest.

Figures 23 and 24 show the water potential dependence of the transport coefficients for a clay and fine sandy loam soil. In both cases, the relationships among the four coefficients are similar. Potential driven vapor flow is a minor constituent for the water potential range shown in the figures and thermal vapor transport assures dominant importance as the liquid transport coefficients decrease with decreasing water potential. The point at which the thermal vapor and potential

FIGURE 22. Liquid and Vapor Transport Coefficients Versus Soil Water Potential for a Silt Loam

FIGURE 23. Liquid and Vapor Transport Coefficients Versus Water Potential for a Fine Sandy Loam

FIGURE 24. Liquid and Vapor Transport Coefficients Versus Soil Water Potential for a Clay

driven flows cross is at approximately -100 cm for the sandy loam and near -500 cm for the clay. It can also be seen in Figure 24 that the thermal vapor diffusivity increases more than is the case of the silt and sandy loam soils. In fact, this coefficient is still increasing at -15,000 cm water potential and may effect considerable water flux even in very dry soils. The water potential dependence of thermal vapor diffusivity is almost negligible for the fine sandy loam. The relationship among the thermal vapor diffusivities for the three soils is intuitively sensible. The water potential dependence for the fine sandy loam is small at water potentials below -2000 cm because air filled porosity does not increase much more with decreasing water potential. The clay soil air filled porosity does increase significantly in this water potential range and the thermal vapor diffusivity increases in direct proportion.

Figures 22-24 serve to document the known trends in the four components of water flow described in the Philip and deVries (1957) theory described in Table 1:

- 1) Temperature effects are shown to be most important for vapor flow.
- 2) Potential driven vapor flow is shown to be of minor importance in the normal soil water potential range.
- 3) Liquid flow coefficients decrease markedly with decreasing water potential.
- 4) Vapor flow assumes increasing importance as water potential decreases.
- 5) Soil type affects the specific values of the flow coefficients but not the relative relationships among them.

- 6) Both the intrinsic permeability and the water release curve must be known in order to assess the water flow properties of a particular soil.

SUMMARY AND CONCLUSIONS

In the last twelve months of this study of heat and water transport from unconfined aquifers, considerable progress has been made. An extensive review of literature resulted in an annotated bibliography of pertinent published reports. In addition, the review also established the level of general knowledge about the problems and phenomena likely to be encountered in unconfined ATES. It was with this background that a parametric analysis of heat and water fluxes was performed.

There are three published reports of unconfined ATES field experiments reviewed in this report. These studies were all conducted with storage within three meters of the surface. Heat loss was judged to be substantial in two of the studies, but calculations were not presented to support this contention. Since the operational and physical principles of unconfined ATES were not addressed, the present study was performed without field experiments to be used in comparison.

Summary remarks written here include treatment of several topics. First, an analysis of the transport phenomena involved is presented. Next, a review and assessment of available data is made. This assessment leads directly to prediction of various component heat and water flows. A presentation of gross system effect on heat storage concludes the parametric analysis of unconfined ATES surface loss of heat. This analysis, however, leads directly to a series of recommendations for future work.

UNSATURATED ZONE TRANSPORT PROPERTIES

Mathematical model of the system. Any detailed analysis of heat loss from aquifers must also treat water losses. This is certainly true for saturated flow in permeable media where convective transport of heat in liquid water can be quite large. Under unsaturated conditions, liquid water convection is not large, but vapor flow in the unsaturated pore space is appreciable. While water transport as vapor is small in magnitude, heat fluxes can be substantial. In order to account for vapor phase heat transfer, both heat and water flow must be known. Table 1 of this report summarizes the coupled heat and water flow theory commonly used by soil scientists and engineers (Philip and deVries, 1957). The eleven equations shown in the table demonstrate temperature and moisture dependencies for both heat transport and water transport. A considerable portion of the parametric analysis presented here involves the analysis of the temperature and moisture dependent factors shown in Equations (5)-(19). Tables 3 and 4 of this report present calculation equations for these factors as functions of temperature and moisture.

In addition to presenting the relationships necessary for calculation of heat and water flow in porous media, a practical theory is presented for calculation of fluxes at the soil-atmosphere boundary. These fluxes may be very important for shallow storage systems, and they can be calculated from meteorological measurements using Equations (12)-(16), (19), and (20).

Calculation of transport coefficients. Since analysis of unsaturated heat and water transport cannot be easily done by inspection of the equations of Table 1, a number of calculations of transport coefficients are presented in Figures 16-24. These figures show important temperature

and moisture dependencies over the ranges of temperature and water potential likely to be encountered for unconfined ATEs. These dependencies can be summarized with reference to Table 6. The table shows moisture dependencies at constant temperature and temperature dependencies at constant moisture. Although the complexities of these relationships are best shown in the figures previously mentioned, the following generalities can be noted from the table:

1. Apparent thermal conductivity (true thermal conductivity plus temperature driven latent heat transport) is very temperature dependent under unsaturated conditions. Under saturated conditions the temperature dependence is less than 10%. The table does not, however, show that the strongest temperature dependence is actually at intermediate soil moisture values (see Fig. 18) where water vapor flux is largest.
2. Apparent thermal conductivity is strongly moisture dependent at low temperatures only. At high temperatures, moisture dependence is masked by the temperature effect on latent heat flux.
3. Temperature driven vapor flow decreases with increasing soil moisture at either high or low temperature. Its temperature dependence is, however, much larger. As a practical matter, the temperature effect alone is important for soils with continuous air-filled pore space. For soils without air continuity, vapor flux is zero.
4. Potential driven vapor flow shows the same dependencies as temperature driven vapor flow.

Table 6. Temperature and moisture dependence of heat and water transport coefficients.

Percentage change from:						
	<u>-15 bar to</u> <u>saturation</u>		<u>-15 bar to 0.1 air</u> <u>filled porosity</u>		<u>0 to 90°C</u>	
	<u>0°</u>	<u>90°</u>	<u>0°</u>	<u>90°</u>	<u>saturated</u>	<u>air dry</u>
Apparent thermal conductivity:						
Fine sandy loam	837	-14	732	1	-3	1244
Silt loam	623	-33	550	-12	7	1373
Temperature driven vapor flow:						
Fine sandy loam			-16	-8		$10^{4.6}$
Silt loam			-54	-48		$10^{4.6}$
Clay			-86	-79		$10^{4.6}$
Temperature driven liquid flow:						
Fine sandy loam	$10^{8.7}$	$10^{8.9}$	$10^{6.5}$	$10^{6.5}$	591	371
Silt loam	$10^{6.9}$	$10^{7.1}$	$10^{5.5}$	$10^{5.4}$	593	415
Clay	10^5	$10^{5.1}$	$10^{4.7}$	$10^{4.7}$	592	452
Potential driven vapor flow:						
Fine sandy loam			-50	-46		$10^{4.8}$
Silt loam			-72	-68		$10^{4.8}$
Clay			-88	-82		$10^{4.8}$
Potential driven liquid flow:						
Fine sandy loam	10^{12}	10^{12}	$10^{8.6}$	$10^{8.7}$	401	183
Silt loam	$10^{10.1}$	$10^{10.3}$	$10^{7.6}$	$10^{7.6}$	400	210
Clay	$10^{7.5}$	$10^{7.7}$	$10^{6.8}$	$10^{6.8}$	401	232

5. Potential and temperature driven liquid flows exhibit similar dependencies. Both liquid flows increased by orders of magnitude from dry to wet soil conditions.
6. The effect of soil type on temperature and moisture dependence is also evident. Liquid flow properties show less moisture dependence and more temperature dependence in fine textured soils. Vapor flow shows more moisture dependence in fine textured soils, but temperature dependence is not affected by soil type.

In addition to the dependencies noted above, the absolute values of the various transport coefficients is of some interest. Since under normal earth surface conditions saturated liquid flow is the dominant water flux, this value is useful to assess the magnitudes of other flow components. This technique is adopted in Table 7. The four water transport coefficients are divided by the saturated hydraulic conductivity and the result is shown for a range of soil types, temperatures, and water contents. These data serve to show that liquid transport coefficients are quite water potential dependent and vapor transport coefficients are primarily temperature dependent. In addition, it is also clear that temperature driven vapor transport is an important mechanism. It is similar in magnitude to the unsaturated liquid conductivity at -100 cm and temperatures less than 45°C, and it is the largest coefficient under drier or warmer conditions. It should be noted that the fluxes resulting from these various mechanisms will depend on gradients in temperature and water potential in addition to the transport coefficients. Since temperature gradients are generally smaller than potential gradients, other transport mechanisms may have higher fluxes than the temperature

Table 7. Relative magnitudes of water transport coefficients. All values are divided by the saturated hydraulic conductivity (potential driven liquid flow).

Temperature:		0°C		45°C		95°C	
Water potential:		-100 cm	-15000 cm	-100 cm	-15000 cm	-100 cm	-15000 cm
Potential driven liquid flow:							
Fine sandy loam		10^{-3}	10^{-10}	10^{-3}	10^{-10}	10^{-3}	10^{-10}
Silt loam		10^{-2}	10^{-8}	10^{-3}	10^{-8}	10^{-3}	10^{-8}
Clay		10^{-1}	10^{-5}	10^{-1}	10^{-6}	10^{-1}	10^{-6}
Potential driven vapor flow:							
Fine sandy loam		10^{-9}	10^{-8}	10^{-8}	10^{-7}	10^{-6}	10^{-5}
Silt loam		10^{-9}	10^{-8}	10^{-8}	10^{-8}	10^{-7}	10^{-6}
Clay			10^{-7}		10^{-6}		10^{-5}
Temperature driven liquid flow:							
Fine sandy loam		10^{-4}	10^{-8}	10^{-4}	10^{-8}	10^{-4}	10^{-8}
Silt loam		10^{-3}	10^{-7}	10^{-3}	10^{-6}	10^{-2}	10^{-6}
Clay		10^{-1}	10^{-4}	10^{-1}	10^{-4}	10^{-1}	10^{-4}
Temperature driven vapor flow:							
Fine sandy loam		10^{-3}	10^{-3}	10^{-3}	10^{-3}	10^{-1}	10^{-1}
Silt loam		10^{-3}	10^{-3}	10^{-2}	10^{-2}	10^{-1}	10^{-1}
Clay			10^{-1}		$10^{-0.4}$		10

mechanisms. Still, it is of note that two-phase water flow does not necessarily decline as the soil desaturates; vapor flux increase may offset the liquid flux decreases at high temperatures.

The parametric analysis of factors affecting heat transfer demonstrated an additional point not shown in Tables 6 and 7. Unlike the water flow coefficients which are influenced by water potential, pore size distribution, total porosity, and temperature, apparent thermal conductivity is also dependent on the composition of the solid phase. The thermal conductivity calculations shown in Figure 19 show that quartz content and bulk density are important factors affecting heat transfer in soils. The effect of quartz is clear: the higher the quartz content, the higher the thermal conductivity. The temperature effect of quartz is only noticeable for saturated soil conditions. Since the thermal conductivity of quartz decreases with increasing temperature, thermal conductivity of a water saturated, quartz-rich material may decline (e.g. Table 6, fine sandy loam). The effect is not noticed for low quartz soils or unsaturated soils where latent heat flux obscures the effect.

The effect of soil bulk density is twofold. Increased bulk density increases thermal conductivity by increasing the solid fraction of the soil but decreases air-filled pore space for vapor transport. Figure 19 shows that bulk density increases thermal conductivity at low temperatures but decreases it at high temperatures.

Conclusions. The parametric analysis of transport mechanisms has established that heat and water flow in unsaturated systems is complex. Liquid water flow decreases markedly with water potential, but

vapor flow increases with decreasing water potential. Temperature effects on water flow are greatest for unsaturated soils and are very important at temperatures above 50°C. Apparent thermal conductivity is largest when there is a vapor phase present because latent heat flux enhances true conduction. This heat transport mechanism is of dominant importance at high temperatures. It must be noted, however, that the high temperature calculations are beyond the range of experimental measurements. Although porous media transport theory clearly shows that the trends presented should occur, the predictions have not been experimentally validated.

HEAT LOSS PREDICTIONS FROM UNCONFINED ATES SYSTEMS

The analysis of temperature and moisture dependence of unsaturated zone heat and water transport presented above is crucial for understanding the nature of the problems which may be encountered with unconfined ATES. The major issue of concern is heat flux, but this cannot be adequately studied without a knowledge of both liquid and vapor fluxes of water. Vapor flow in particular markedly influences heat flux since vapor flux and latent heat flux are directly related. In a system as loosely controlled as evaporation from an unconfined aquifer, it is a complex problem to calculate heat loss from a heated groundwater body. Nevertheless, this issue will ultimately be the focus of the conclusions presented here. Accordingly, this summary of major factors affecting unconfined ATES includes calculations of heat loss from model systems. The factors considered in this analysis are (1) depth to the unconfined aquifer, (2) temperature of the aquifer, (3) soil transport properties, and (4) atmospheric conditions.

The calculations presented here were made using a one dimensional finite element solution to Equations (10) and (11). Surface boundary conditions of air temperature, net radiation, relative humidity, and wind speed were fixed and the values were inserted in the model using Equations (12), (15), (16), and (20). Lower boundary conditions were aquifer temperature and the air entry water potential of the porous medium. Calculations were made for a five month period assuming constant boundary conditions. Time steps from 15 seconds to two days were used and no iteration was performed at a given time step. Since soil water potential and temperature profiles were constant or nearly so after five months of simulation, it was presumed that the unsaturated flow system was in dynamic equilibrium. These profiles do not, however, have the same shape as commonly observed profiles above shallow water tables. Table 8 shows water content and temperature for two water table heights. As expected, soil temperatures are higher for the profile with the shallower water table, and both soils show a steeper temperature gradient near the surface. The water content profiles, however, do not show a steady decrease in water content with elevation above the water table. Instead, the water content decreases only slightly until near the surface. At this point (between 60 and 30 cm in the data shown) the water content decreases markedly and the soil is dry to the surface. This pattern of water distribution is common in Eastern Washington and Oregon and is characteristic of temperature induced water flow under dry climatic conditions (Hammel, 1979). The profile remains moist throughout much of the subsoil, and surface evaporation is inhibited by the very dry surface soil layers. These moist soil conditions mean that vapor transport on the large temperature gradient is small or zero. Thus,

Table 8. Temperature and moisture profiles for a 65° water table at two depths. Air temperature is 15°C. Silt loam soil.

Depth	10 Meter Water Table		3 Meter Water Table	
	Temperature	Water content	Temperature	Water content
-cm-	°C	cm ³ /cm ³	°C	cm ³ /cm ³
Air	15		15	
8	15	.07	19	.08
16	16	.11	23	.12
30	17	.12	29	.11
60	20	.22	33	.46
90	22	.22	37	.30
120	23	.22	42	.31
150	25	.22	46	.32
200	28	.23	52	.35
250	30	.23	58	.40
300	33	.23	65	.53
600	37	.24		
800	47	.26		
900	56	.30		
1000	65	.53		

even though the soil temperature is such that vapor flow and latent heat flux transport coefficients could be large, there is little air-filled pore space for vapor flow. The configuration of temperature and moisture profiles limits heat and water flux since the dry surface soil has very small water flux and the moist subsoil has a relatively low thermal conductivity. These phenomena may indeed make unconfined ATEs feasible.

The efficiency of heat storage in unconfined aquifers can be investigated using the data presented in Table 9. The table shows heat flux from an aquifer as a function of the four factors mentioned earlier. Three effects of these factors are clearly shown. First, shallow aquifers lose more heat than deep aquifers; second, higher temperature aquifers lose more heat; and third, cooler air temperatures enhance heat loss. These observations are as expected, but the data presented also allow some additional analysis. Comparison of the 65 and 90°C aquifer temperature data at a single depth shows that heat flux is roughly proportional to the temperature gradient from the air to the water table. That is, even though the heat and water transport coefficients are temperature dependent in varying degrees, the overall effect on the flow system is not as important as the gradient in temperature.

The effect on heat flux of aquifer depth is the largest magnitude effect of the four under discussion. Aquifer temperature is the second largest factor, and the effect of air temperature and soil type are both smaller. It is interesting to note that the effect of air temperature is relatively unimportant in the range studied here. Heat loss may well be large for colder temperatures, but the temperature dependence of the transport coefficients should work against that effect. In the table it can be readily seen that storage of cool water is more

Table 9. Heat flux from unconfined ATEs. Effects of air temperature, aquifer temperature, aquifer depth, and soil type.

Aquifer depth	Aquifer Temp. °C Air Temp. °C	5			65			90		
		5	15	25	5	15	25	5	15	25
meters		-----Heat flux, w/m ² -----								
		SILT LOAM								
10		--	-3	-8	5	1	-3	9	3	--
5		--	-4	-9	12	7	3	23	17	13
3		-2	-5	-10	30	22	14	43	40	32
2		-2	-6	-12	42	33	23	59	54	45
		FINE SANDY LOAM								
10		--	-3	-6	3	--	-3	7	2	--
5		--	-3	-8	8	5	2	18	14	12
3		-1	-4	-8	16	14	10	35	33	30
2		-1	-5	-9	28	24	18	61	57	50
		CLAY								
10		2	-2	-7	9	4	--	12	7	4
5		1	1	-8	15	11	7	23	19	15
3		--	-5	-10	23	19	14	38	34	29
2		--	-5	-12	35	29	22	58	52	46

efficient than heat storage under the conditions tested. This is a reasonable conclusion since it is likely that temperature gradients for cool water storage will be less than those for heat storage. A second important point to note is that the parametric analysis presented previously showed that vapor flux (and therefore latent heat flux) becomes small at low temperatures. This would also increase storage efficiency.

Although the effects of air temperature, aquifer depth, and aquifer temperature are fairly straightforward, the effect of soil type on heat flux is complex. For storage of cool water, the three soils shown in Table 9 are quite similar. The soils are also similar for 10 meter aquifers. For other conditions, the soils behave differently. The clay soil has the largest heat flux from a 5 meter water table, but at 3 meters the soil loam flux is substantially higher. The silt loam is higher for 65°C water at 2 meters but all three soils are quite similar for 90°C water at 2 meters. It seems that under extreme conditions soil type is not an important factor governing heat loss from aquifers. Under less extreme conditions, however, soils with intermediate vapor and liquid water flow characteristics have high heat loss. The fine sandy loam studied here is quite efficient over a large range of aquifer conditions. It may well be that a more coarse textured material would be a better medium for unconfined ATEs.

RECOMMENDATIONS FOR FUTURE STUDY

Three groups of questions and problems arose from the analyses presented in this report. All have merit and deserve additional study. The areas are (1) detailed measurement of unsaturated zone properties and fluxes, (2) assessment of field techniques to enhance unconfined

ATES, and (3) coupled heat and water flow calculations for predicting storage efficiency for various sites. All three areas are interrelated and simultaneous studies in the three areas could be beneficial.

Detailed measurements. There is a definite need to develop techniques to measure unsaturated zone heat and water fluxes. Techniques are required in order to monitor test site installations, as well as to gather basic data. A pressing need for the questions of unconfined ATES is validation of heat and water flow predictions above 45°C which are given in this report. To date, experimental data in this temperature range are not available. Measurements in this range would be directly applicable to the unsaturated zone flows likely to be encountered in ATES. Development of test site monitoring equipment could be done along with the measurements described above. It is suggested that this phase of work be given high priority.

Assessment of field techniques to enhance unconfined ATES. If the principles and dynamics of unsaturated heat and water flow are kept in mind, a number of cultural practices for improving aquifer storage efficiency can be suggested. Since surface heat flux is to be minimized, a number of surface treatments should be considered. Any barrier to water vapor flow should have some effect on heat loss. Possible techniques to consider are tillage mulching, compaction of soil layers, buried plastic or asphalt sheets, saturation of some near surface layer, or chemical treatment to block pore continuity. These techniques could be used either to prepare a site for an artificial aquifer or to decrease heat loss from existing shallow aquifers. The potential for such techniques is sufficient to warrant investigation but these measures probably do not require immediate attention.

Coupled heat and water flow calculations. Additional study of the processes of heat and water loss are required in order to better understand the dynamics of unconfined ATES. Although the parametric analysis presented here gives trends and ranges of behavior, there is a variety of specific site conditions which may affect heat storage. Modeling of the surface treatments described in the previous section would be an efficient way to analyze each treatment effect. Another area where additional calculations may be of value is the effect of atmospheric conditions on surface fluxes. Although the effect of the surface boundary is not as important as other parameters of the ATES system, additional research in this area would be important for analysis of the fine details of storage. Spatial variability of both surface conditions and soil properties will affect ATES. The effects of this variation should be assessed.

Finally, it should be noted that the effect of site properties on ATES must be determined in advance of an installation at a site. Additional system modeling at the scientific level is probably required before site inspection and assessment can be performed on an engineering basis. This modeling should be carried out in conjunction with an experimental program designed to make pertinent measurements in the unsaturated flow system. Simultaneous, coordinated work in these two areas will result in a better understanding of the relevant flow phenomena, the magnitudes of the component flows, and their importance to the overall problem of heat and water transport in unsaturated porous media.

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APPENDIX
ANNOTATED BIBLIOGRAPHY: HEAT AND MASS TRANSFER
IN UNSATURATED POROUS MEDIA

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ANNOTATED BIBLIOGRAPHY: HEAT AND MASS TRANSFER
IN UNSATURATED POROUS MEDIA

I. THEORY

A. Development of flow equations using nonequilibrium thermodynamics.

1. Groenevelt, P. H. and G. H. Bolt. 1969. Non-equilibrium thermodynamics of the soil-water system. J. Hydrol. 7:358-388.

Following standard texts on the subject matter, a combination of the Gibbs equation of thermostatics, (1), with appropriate conservation equations, (5,6,8), is shown to yield an expression, (10), for the local production of entropy (i.e., energy dissipation) accompanying the flow of heat and mass. This equation indicates that for viscous flow this local entropy production is not related unambiguously to the local rate of loss of pressure and potential energy as part of the energy may be transferred to the environment in the form of viscous stress energy or stored as kinetic energy.

The application of this equation to flow in porous media requires its integration over a finite volume as only surface averaged fluxes and potential differences over finite distances are accessible for experimental determination. It is shown in section 3 that in the process of integration the above energy transfer terms tend to disappear, yielding a straightforward expression for the entropy production, (16), in terms of the products of these averaged fluxes and their conjugated forces viz. the "macroscopic" gradients of the temperature and thermodynamic potentials.

As the particular fluxes and forces appearing in this basic equation are rather inconvenient for soil water systems, a regrouping of terms is shown to yield an alternative expression in which appear the fluxes of heat and soil solution, the diffusion fluxes of salts and the electric current. For the single salt system the conjugated forces are then the gradients of the temperature, hydraulic head, osmotic pressure and electric potential (19).

In section 5 generalized relationships between fluxes and forces are introduced (29) and the significance of the different transport coefficients in soils, especially the coupling coefficients, is discussed. In section 6 the magnitude of these coefficients in isothermal systems is calculated for an assumed model based on the presence of an electric double layer as the main cause of the selective action of the porous system on transport processes.

2. Taylor, S. A., and J. W. Cary. 1964. Linear equations for the simultaneous flow of matter and energy in a continuous soil system. SSSAP 28:167-172.

Some transfer processes in soil are considered with respect to recent developments in the theory of thermodynamics of irreversible processes. The theory provides a general and systematic guide for setting up interrelated equations describing the transport of one or more components through soil. Darcy's law and the "Diffusion equation" are shown to be components of this general development. Particular attention is given to the non-isothermal flow of moisture and energy through soil. A general relation between the fluxes induced by the combined gradients of moisture content, temperature, and solutes is also suggested.

3. Winterkorn, H. F. 1958. Mass transport phenomena in moist porous systems as viewed from the thermodynamics of irreversible processes. In H. F. Winterkorn (ed.), Water and its conduction in the soil. Highway Res. Board Spec. Rep. 40.

This article is a general discussion about the application of thermodynamics to soil water and heat transport. Just general statements are made about each phase. Such a topic would be "Some remarks on water transport in vapor phase" or "Some remarks on water transport in liquid phase."

B. Continuum mechanics derivation of flow equations.

1. Kay, B. D., and P. H. Groenevelt. 1974. On the interaction of water and heat transport in frozen and unfrozen soils. I. Basic theory: the vapor phase. SSSAP 38:395-404.

A theory is presented for water transport due to a difference in temperature and heat transport due to a difference in water pressure as they occur in frozen and unfrozen soils. It is shown by model considerations that the transport coefficients of these two phenomena, as far as the vapor phase is concerned, are equal. This coupling is related to the heat of vaporization, the heat of fusion and the heat of wetting of the soil water.

2. Nakano, M., and T. Miyazaki. 1979. The diffusion and nonequilibrium thermodynamic equations of water vapor in soils under temperature gradient. Soil Sci. 128:184-188.

Based on the observation that temperature is a function of distance, Fick's law of diffusion is extended as widely as possible to apply to the movement of water vapor in soils under temperature gradients. The general equations

developed as a result are expressed using (1) water vapor pressure, (2) water vapor density, (3) the Clapeyron equation, and (4) the form of phenomenological equations in nonequilibrium thermodynamics. From comparisons with the equations derived in several previous reports, we conclude that the earlier results may be valid only under the condition that the total water potential is larger than -10^6 centimeters.

3. Raats, P. A. C. 1975. Transformations of fluxes and forces describing the simultaneous transport of water and heat in unsaturated porous media. *Water Resour. Res.* 11(6):938-942.

Balances of mass for the water in n distinct phases and a balance of heat for the medium as a whole are formulated. Following Philip and deVries, it is assumed that the flux of water in each phase is proportional to the gradient of the pressure in that phase and that the diffusive component of the flux of heat is proportional to the gradient of the temperature. Clapeyron equations are used to express the gradient of the pressure in any phase in terms of the gradient of the pressure in a reference state and of the temperature. The reference state may be the water in one of the phases or the water in some measuring device such as a tensiometer or a psychrometer. Expressions for the total flux of water and for the diffusive flux of heat plus the convective flux of heat associated with the conversion from any phase to the reference state are shown to satisfy the Onsager reciprocal relations.

4. Rollins, R. L., M. G. Spangler, and D. Kirkham. 1954. Movement of soil moisture under thermal gradient. *National Res. Council Proc., Highway Res. Board Mtg.* 33:492-508.

An analysis of the flow of soil moisture under a thermal gradient is made for a closed system. This theoretical study indicates that an equilibrium state requires a cyclic transfer of moisture in the soil-water system. Water moving from the warm to the cold side in the vapor phase is returned to the warm side in the liquid phase. This analysis also indicates that the accumulation of moisture on the cold side may be the results of movement of water both as a liquid and as a vapor. However the analysis does not show anything about the relative amounts of moisture in the two.

Using principles from the above analysis, an apparatus for determining quantitatively the amount of water moving in the vapor phase in a steady state process is described. Water vapor moving through the soil system from the warm to the cold side is returned to the warm side through an external capillary tube, in which the rate of movement is measured. A differential equation is presented for

estimating the amount of water vapor moving in the system, assuming a molecular diffusion process.

A comparison of results of moisture distributions for circulating and noncirculating systems indicates that moisture movement in the liquid state is not the predominant method of moisture migration for the soil used.

Calculated rates of water movement, assuming a molecular diffusion process, were approximately a sixth of the values of measured rates. Data are presented showing the relationship between the thermal gradient and the quantity of flow. The effect of the dry density and the percent of air-filled void is evident from this data.

5. Sophocleous, Marios. 1979. Analysis of water and heat flow in unsaturated-saturated porous media. *Water Resour. Res.* 15(5):1195-1206.

The study described investigates the theoretical and practical aspects of modeling water and heat flow in an unsaturated-saturated flow domain. Field experiments were designed and complementary laboratory analyses were employed to measure the relevant physical quantities that would enable the significance of non-isothermal effects to be determined. The Philip and deVries non-isothermal flow model was adapted to the more general pressure (or matric) head-based approach and a new expression for the thermal liquid water diffusivity.

6. Sposito, G. 1978. The statistical mechanical theory of water transport through unsaturated soils. I. The conservation laws. *Water Resour. Res.* 14:474-478.

The macroscopic differential equations of mass and momentum balance for water in a rigid unsaturated soil are derived from first principles by using the methods of statistical mechanics. The derivation begins with the development, at the molecular level, of expressions for the mass and linear momentum densities of water in a soil. The derivatives of these expressions with respect to the time then provide for local balance equations that are ensemble averaged and local volume averaged in order to produce the macroscopic balance equations. The resulting macroscopic equations agree with those derived for mass and linear momentum transport in a soil on the basis of the continuum theory of mixtures. The unified insight into mass transport processes afforded by a statistical mechanical theory as compared to one evolved from a solely macroscopic view is discussed.

C. Description of flow mechanisms.

1. Cary, J. W. 1966. Soil moisture transport due to thermal gradients: practical aspects. *SSSAP* 30:428.

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.

2. Cary, J. W. 1967. The drying of soil: thermal regimes and ambient pressures. *Agr. Meteor.* 4:353-365.

Evaporation from a loam soil was studied in the laboratory during the falling-rate stage of drying. The primary variables were temperature 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.

3. deVries, D. A. 1958. Simultaneous transfer of heat and moisture in porous media. *Trans. AGU* 39(5):909-916.

Simultaneous differential equations for the transfer of heat and moisture in porous media under the combined influence of gravity and gradients of temperature and moisture content are developed. These equations are a

generalization of those derived by Philip and deVries (1957). In particular a consistent distinction is made between changes of moisture content in the liquid and the vapor phase.

The interaction between heat and moisture transfer in a steady state heat conduction is discussed in detail using numerical values for two soils, a clay loam and a medium sand. The behavior is found to be dependent on the boundary conditions for moisture transfer, on the direction of the temperature gradient and on the ratio of two moisture diffusivities entering in the analysis.

4. deVries, D. A. 1976. Heat transfer in soils. In D. A. deVries and H. H. Afgan (eds.), Heat and mass transfer in the biosphere. Scripta, Washington, D. C.

Heat is transferred in soils mainly by conduction. Transport of latent heat by water vapor diffusion in the air-filled pores must also be taken into account.

The calculation of the overall thermal conductivity and the volumetric heat capacity of a soil from its composition is treated.

Some general features of the storage and release of heat during the annual and diurnal temperature cycles and their influence on the energy balance of the earth's surface are discussed. The penetration of frost into a soil is touched upon.

The main unsolved problems in the field of soil heat transfer are connected with the combined transfer of heat and moisture in soils. Present theories and their limitations are briefly discussed and suggestions for further work are given.

5. Farrel, D. A., E. L. Greacen, and C. G. Gurr. 1966. Vapor transfer in soil due to air turbulence. Soil Sci. 102:305-313.

Their analytical mathematical works show that the depth of air movement due to pressure variations caused by gusty wind is on the order of several cm for soil mulches.

Also, in this near surface turbulent air flow region, Da is as much as 100x the laminar flow value.

6. Jury, W. A., and J. Letey. 1979. Water vapor movement in soil: reconciliation of theory and experiment. SSSAJ 43:823-827.

A large number of separate experimental studies of water vapor movement in response to thermal gradients are shown to be independent of water content over a wide range of water contents. Two models, proposed by J. R. Philip and D. A. deVries (1957) and J. W. Cary (1963), are rewritten in a compatible form and compared to the data. The phenomenological model of Cary describes the data if the

phenomenological coefficient β has a value between 1.0 and 3.5, whereas the mechanistic theory of Philip and deVries underpredicts the vapor transport observed in the experiments in all cases.

A modification of the theory of Philip and deVries is proposed which increases the influence of liquid water on vapor transport. In the new formulation the experimental information is adequately described and both models are rendered compatible. Both experimental and theoretical results suggest that water vapor movement may be described by a simple expression which is a known function of temperature, and is independent of water content over a wide range of saturation. Even without calibration it is suggested that this expression can describe water vapor flux to within a factor of 2.

7. Kimball, B. A., and E. R. Lemon. 1971. Air turbulence effects upon soil gas exchange. SSSAP 35:16-20.

The effects of air turbulence upon the exchange of gases between a soil and the atmosphere were investigated using a specially designed vapor exchange meter. The device measured the rate of evaporation of liquid heptane from a porous stainless steel plate buried in a porous medium. Natural air turbulence as indicated by both mean windspeed and root mean square air pressure fluctuations significantly affected the heptane evaporation rates in straw and coarse gravel, but had a decreased effect in media of smaller pore size. In silt loam, the regression coefficient for evaporation rate on windspeed was 20% of its value for straw. Increasing depth from 1 to 8 cm in coarse sand caused a 91% decrease. Although the heptane flux values are quite scattered, the data indicate that air turbulence can significantly increase the transport of water vapor through coarse mulches or through very shallow depths of soil, but that soil aeration is mostly a diffusive process.

8. Kuznak, J. M., and P. J. Sereda. 1957. The mechanism by which water moves through a porous material subjected to a temperature gradient. Part I. The introduction of a vapor gap into a saturated system. Soil Sci. 84:291-300.

This experiment was designed to detect liquid phase flow in porous media under a temperature gradient. They found no liquid water movement in their experiment.

9. Perrier, E. R., and O. Prakash. 1977. Heat and vapor movement during infiltration and redistribution. Soil Sci. 124:73-76.

Temperature and relative humidity were monitored in a horizontal column apparatus at selected sites during

infiltration into dry soil. As the wetting front approached a measurement site, the temperature was observed to rise; later the humidity was observed to increase, which was followed very closely by the wetting front. As the liquid front passed the measurement site, the temperature was observed to drop rapidly.

These data show that the advance of a wetting front involves several processes: (a) as dry soil is wetted, a large amount of heat is evolved; (b) evaporation during the wetting process supplies a vapor phase; (c) as liquid moves into dry soil, the vapor phase moves as a front immediately ahead of the wetting front; (d) a large portion of the heat evolved moves as a front well in advance of both the vapor and liquid fronts; and (e) the heat evolved subsequently heats the liquid front; however, due to evaporation and the thermal conductivity of water, the wetted soil behind the liquid front is cooled.

10. Philip, J. R., and D. A. deVries. 1957. Moisture movement in porous materials under temperature gradients. Trans. AGU 38:222-232.

A theory of moisture movement in porous materials under temperature gradients is developed which explains apparently discordant experimental information, including (a) the large value of the apparent vapor transfer, (b) effect of moisture content on net moisture transfer, and (c) the transfer of latent heat by distillation.

The previous simple theory of water vapor diffusion in porous media under temperature gradients neglected the interaction of vapor, liquid and solid phases, and the difference between average temperature gradient in the air-filled pores and in the soil as a whole. With these factors taken into account, an (admittedly approximate) analysis is developed which predicts orders of magnitude and general behavior in satisfactory agreement with the experimental facts.

An important implication of the present approach is that experimental methods used to distinguish between liquid and vapor transfer have not done so, since what has been supposed to be vapor transfer has actually been series-parallel flow through liquid "islands" located in a vapor continuum.

Equations describing moisture and heat transfer in porous materials under combined moisture and temperature gradients are developed. Four moisture-dependent diffusivities arising in this connection are discussed briefly.

11. Rose, D. A. 1963. Water movement in porous material. Part 1. Isothermal vapour transfer. Brit. J. Appl. Phys. 14:356-262.

The isothermal movement of water vapor through six different porous materials (both consolidated and granular) has been studied using cylindrical samples (5 cm diameter and 1.2 cm deep) in a simple diffusion cell. Two flow regimes occur. Initially, as moisture content θ/θ_T increases the apparent conductivity (k' cm sec⁻¹) increases rapidly, confirming predictions of Philip and deVries (1957) that a discontinuous liquid phase can enhance water movement by transmitting a vapour flux and thus acting as a short circuit path for vapour. When liquid continuity is established, however, further increase in θ/θ_T causes k' to rise slowly to a limiting value dependent on the thermal conductivity of the material, suggesting that liquid flow is inhibited by a back hydromotive force set up by unavoidable temperature gradients produced by condensation and evaporation. Apparent diffusivities (D' cm² sec⁻¹) increase rapidly as θ/θ_T increases, as predicted by Philip (1955) for vapor movement. The cell gives quick and reproducible estimates of water potential in the range pF2.5 to pF5.6.

12. Rose, D. A. 1968. Water movement in dry soils. I. Physical factors affecting sorption of water by dry soil. J. of Soil Sci. 19:81-93.

Semi-infinite columns of dry soil closed at one end had the other exposed to a turbulent atmosphere at constant relative humidity (0.98) and a range of constant temperatures. After varied times the water content of the columns was measured gravimetrically, in 1 cm layers, from which the total quantity taken up Q and the distribution of water content θ with time t and distance x were found. Assuming that the Boltzmann transform, $\lambda = xt^{-1/2}$ can be applied to the standard diffusion equation, two soil parameters are derived. A test of the assumption is that θ should be uniquely dependent on λ , and then the diffusivity is calculable. The second parameter, the sorptivity, is not independent of θ_i or the value maintained at $x = 0$.

Results show that in a clay soil, between pF5.8 and pF4.2, water moves predominantly as vapor: in a non-swelling silicate mineral (seipolite) there is significant liquid movement between pF5.1 and pF4.2. Long-continued use of organic manure has little effect on either D or S ; aggregate size has some effect, in opposite senses for a sand and a clay; a mulch or a still atmosphere affects S but not D ; evaporation suppressants decrease S ; ignition decreases S and greatly alters D ; and degradation of structure causes small changes in D .

13. Scotter, D. R., G. W. Thurtell, and P. A. C. Raats. 1967. Dispersion resulting from sinusoidal gas flow in porous media. Soil Sci. 104:306-308.

Random air pressure fluctuations resulting from wind effects or surfaces may be quite important to gas flow in near surface zone. These data show that, in the surface 20 cm, the dispersion coefficient may double under sinusoidal pressure fluctuations.

14. Taylor, S. A., and L. Cavazza. 1954. The movement of soil moisture in response to temperature gradient. SSSAP 18:351-358.

Using an air gap technique, it has been shown that the flow of moisture in the soil from warm to cool regions might occur largely in the vapor phase. This is accompanied by a return flow of liquid water in response to an induced flow potential gradient. The use of wire screens to make air gaps interferes with this process.

The measured diffusion coefficient for vapor flow was higher than had been expected from diffusion data, but was substantially in agreement with that of other workers. The reasons for this high diffusion coefficient might be that the diffusion of moisture and heat is coupled in soils; the phenomena of surface migration and molecular hopping of adsorbed water might also cause this increase.

The factors having greatest effect on the moisture distribution in continuous soil columns in a closed system are: (1) the diffusion and the convection of vapor or some related processes; (2) the evaporation and condensation occurring in the two end regions of the cylinders; (3) the hydraulic conductivity of the soil.

The thickness of the air gap in broken columns influences the amount of water that will be transferred in the vapor phase and must be considered whenever an air gap is used to prevent fluid flow.

The existence of two distinct drying and wetting moisture characteristic curves has been verified, and they play an important part in determining the liquid flow of moisture in the soil. The influence of this factor as well as the reason for the large vapor transfer need further study in relation to the transfer of soil moisture.

15. Wierenga, P. G., R. M. Hagan, and D. R. Nielsen. 1970. Soil temperature profiles during infiltration and redistribution of cool and warm irrigation water. Water Resour. Res. 6(1):230-238.

Soil temperatures were measured in field profiles irrigated with cool and warm water and also measured in nonirrigated soil. Difference in soil temperature between

plots after irrigation with cool (.4 1C) and warm (21. 6C) water, persisted longer at the lower soil depths. The time required after an irrigation for these differences to decline to less than 1C was 14 hr at the 1 cm depth, 36 hr at the 10 cm depth, and 42 hr at the 20 cm depth. A numerical procedure developed for predicting changes in soil temperature. Method gave good agreement between observed and calculated values in the subsoil but not in the surface soil. During infiltration heat transfer occurred mainly by mass movement, whereas after 10 hr soil heat transfer below 10 cm occurred mainly by conduction. 14 refs.

16. Woodside, L. V., and J. M. Kuznak. 1958. Effect of temperature distribution on moisture flow in porous materials. Trans. AGU 39:676-680.

Several experimenters have observed moisture flow rates through porous materials under a temperature gradient, which are many times the calculated rates based on vapor diffusion. To investigate this discrepancy, a temperature gradient was applied across an enlarged model of a particular, high-quartz wet sand, and the temperature distribution within the pore spaces was measured. For the 'air-dry condition' of the model, the average temperature gradient in the pore spaces was six times the overall applied gradient. Approximately this same ratio exists, therefore, between the respective vapor pressure gradients.

It is proposed that the flow mechanism consists of multiple vapor diffusion and liquid flow steps in series, the resistance to flow in the liquid phase being negligible, so that the flow rate is determined by the vapor pressure gradients in the pore spaces. The fact that these gradients are of the order of six times the overall gradient previously used to calculate the flow rate at least partially explains the discrepancy.

II. COUPLING OF HEAT AND MASS TRANSFER TO THE ATMOSPHERE

A. Effect of surface meteorological conditions on evaporation and sensible and latent heat losses.

1. Barton, I. J. 1979. A parameterization of the evaporation from nonsaturated surfaces. J. of Appl. Meteor. 18(1):43-47.

The parameterization of evaporation proposed by Priestley and Taylor (1972) is modified to take account of nonsaturated surfaces. It is suggested that evaporation can be expressed as a function of net radiation, air temperature, a constant related to the nature of the surface, and a variable that is dependent on moisture availability. The constant is equal to 1.26 for water and open grassland surfaces (as in Priestley and Taylor) but is nearer unity for some other surfaces. The variable is found to be directly related to near-surface soil moisture content for both bare and grassland surfaces.

2. Cook, D., and J. M. Norman. 1977. Measurement and prediction of increased evaporation from combined soil warming and irrigation. Preprint of 13th Agr. and For. Meteorol. Conf. AMS. pp. 117-118.

The combined influence of soil warming and irrigation on the soil surface energy budget has been investigated using an artificially heated plot and an unheated control. Net radiation, soil heat flux, evapotranspiration and drainage have been monitored over a 12-month period of irrigation on both plots. These data, along with routine meteorological and soil measurements, have been used to obtain all the terms of the surface energy budget equation for the heated and unheated plots on a daily basis. The results provide a quantitative assessment of the enhanced evapotranspiration water loss from an artificially heated soil and provide some of the essential information required for modeling the use of soil warming as a means of dissipating excess heat from large electric power generating stations.

3. Gardner, W. R., and D. Hillel. 1962. The relation of external evaporative conditions to the drying of soils. J. Geophys. Res. 67:4319-4325.

Evaporation from laboratory soil columns was studied as a function of potential evaporative conditions. The length of time a given evaporation rate could be maintained by the soil was in good agreement with an approximate solution of the isothermal equation for unsaturated flow. During the falling-rate period of drying, the evaporation rate was found to approach very nearly a function of the water content of the soil and to be nearly independent of

the potential evaporation rate. The cumulative evaporation at any time can be calculated from the flow equation. Temporary interruption of the evaporative process had little effect upon the cumulative evaporation over a long period of time.

4. Hanks, R. J., and N. P. Woodruff. 1958. Influence of wind and water vapor transfer through soil, gravel and straw mulches. Soil Sci. 86:160-164.

This paper attempts to expand the knowledge about the influence of wind on water vapor transfer through dry soil, gravel, and straw mulches.

5. Hanks, R. J., H. R. Gardner, and M. L. Fairbourn. 1967. Evaporation of water from soils as influenced by drying with wind or radiation. SSSAP 31:593-595.

The evaporation rate from three soils initially wet was made the same by adjusting the wind and radiation intensity. The wind treatment caused a temperature depression at the soil surface initially which nearly disappeared after about 5 days. The radiation treatment caused a temperature increase at the soil surface which increased with time. Soil water content profiles measured as a function of time showed the water content near the soil surface to be higher for the wind treatment than for the radiation treatment. The components of flow due to temperature and suction gradients for both liquid and vapor flow were estimated using the analysis of Philip and deVries. This analysis indicated the cumulative downward vapor flow due to temperature gradients amount to about 10% of the net upward flow in 40 days at the 5 cm depth for one soil for the radiation treatment. Computations of evaporation assuming isothermal conditions for soils initially wet to near saturation would probably estimate total evaporation within 10%, and probably be sufficiently accurate for many purposes.

6. Lykosov, V. N., and E. G. Palagin. 1978. Dynamics of cross-coupled heat and moisture transport in the atmosphere-soil system. Sov. Meteorol. Hydrol. 8:35-42.

Energy and mass exchange in the soil are studied by mathematical modeling. Attention is concentrated on procedures for calculating the temperature and moisture profiles, formulating underlying-surface balance conditions, and determining the heat and moisture fluxes at the atmosphere-soil boundary. An algorithm is proposed for solution of the corresponding problems and an example calculation is given.

7. Nagpal, N. K., and L. Boersma. Evaporation from two porous media during diurnal radiation cycles. Can. J. Soil Sci. 38:465-474.

The process of evaporation from porous media subjected to diurnal radiation cycles was studied. Information about pore size effects on evaporation was also obtained. Two porous media consisting of glass beads with diameters ranging from 53 to 74 μm and from 149 to 210 μm were used. Rates of water loss from the initially saturated media varied during the diurnal cycle. The initial rise in rate of surface irradiation increased the rate of water loss from both media. For the coarser material the rate of evaporation continued to increase for several hours after the maximum rate of irradiation was reached while for the finer material the rate of evaporation reached a maximum value even before the maximum rate of irradiation was attained. As a result, the ratio of the evaporation rates, $E_{\text{coarse}}/E_{\text{fine}}$, varied during the day from approximately 1.0 during periods of low evaporative demand to 2.60 during periods of high evaporative demand with an average value of 1.47 for the full day. The total amount of water lost per day from each medium remained constant during a 7-day experimental period, even though the surface layers dried out during periods of high irradiation. The pore size dependency of the rate of water loss was shown to be determined by the hydraulic conductivities of the porous media. The rates of water loss predicted by unsaturated water flow equations agreed well with the experimental data.

8. Philip, J. R. 1957. Evaporation and moisture and heat fields in the soil. J. Meteorol. 14:354-366.

Recent developments in the study of moisture and heat fields in the soil are applied to the problem of evaporation from bare soil surfaces. Most of the analysis is confined to steady conditions, but it definitely suggests that there are three phases to the desiccation of a given soil profile:

1. So long as the soil is sufficiently moist, the evaporation rate, E , is indistinguishable from that from a saturated surface, E_0 .
2. At intermediate moisture contents, E is independent of E_0 and depends only on the soil-moisture distribution.
3. When the surface layers of the soil are sufficiently dry, E is sensitive to the heat flux in the soil, and a negative correlation between E and E_0 may follow.

The first and second of these correspond to the well-known constant and falling rate phases of the "isothermal" drying of initially saturated soils and other porous media, which thus receive quantitative physical explanation.

There is experimental evidence of the existence of the third phase also.

Other results of the article include:

1. Development of a (steady state) model of the energy balance at imperfectly evaporating surfaces ($E < E_o$). This leads to a simplified but suggestive approach to the microclimate of bare soils.
2. Analysis of the spatial distribution of evaporation sites within the soil.
3. Analysis of the modification of the energy balance which results from the location of evaporation sites within the soil rather than at the surface.

9. Rose, C. W. 1968. Evaporation from bare soil under high radiation conditions. Int. Congr. Soil Sci. Trans. 9th. pp. 57-65.

This paper presents a theory that enables evaporation from bare soil to be calculated from water content and temperature profiles. The theory is of general application and is applied to data from a field experiment carried out under fairly high radiation conditions.

10. Staple, W. J. 1971. Boundary conditions and conductivities used in the isothermal model of evaporation from soil. SSSAP 35:853-855.

The surface boundary conditions in the isothermal model for computing evaporation from soil were expressed in terms of the wind speed, temperature, and humidity of the ambient air. Use of these parameters along with reduced nodal spacings near the surface produced evaporative fluxes that closely followed the differences in vapor pressure between soil and atmosphere. A simple evaporation experiment was used to show that evaporation rates and soil moisture distributions obtained with these revised boundary conditions were closer to experimental values than were those obtained with the usual conditions defined in terms of potential evaporation and moisture content.

11. Westcot, D. W., and P. J. Wierenga. 1974. Transfer of heat by conduction and vapor movement in a closed soil system. SSSAP 38(1):9-14.

A study was made to determine the importance of vapor flow to the heat transfer process in soil. Soil temperature and water content were measured in sealed columns, 7.6 cm in diameter and 32 cm in length. The columns were inserted into a field plot and exposed to daily temperature fluctuations for periods varying from 1 to 6 days. A computer simulation model was developed to predict vapor movement and the transfer of heat by conduction and by vapor diffusion.

B. Transport resistances at the soil surface.

1. Aragaki, M., and K. Shirai. 1975. Evaporation from soils under temperature gradients. Bull. of Faculty of Agric., Mie Univ. 49:101-106.

In this paper, evaporation from soils under temperature gradients in a current of air has been studied theoretically and experimentally. Experimental research deals with the case of negative temperature gradients only.

2. Bailey, R. T., J. W. Mitchell, and W. A. Beckman. 1973. Convective heat transfer from a desert surface. ASME Pap. No. 75-HT-AA. 6 p.

The convective heat transfer relations for atmospheric flow over sparsely vegetated areas are reviewed and compared to existing relations for flow in rough ducts. Experimental convection coefficients obtained at four desert sites are compared to the analytical relations. The experimental equipment for measuring the heat transfer convection coefficient between air and ground in a desert environment consists of two electrically heated plates positioned flush with the ground. Measurements of power dissipation and surface temperatures allow direct calculation of the convection coefficient. The experimental heat transfer results are correlated with micrometeorological models from which a soil roughness height is calculated. This roughness parameter is shown to characterize air flow near the soil surface, and may be significantly different from the roughness parameter ordinarily determined from velocity profiles. A simplified heat transfer correlation is presented for desert surfaces.

3. Das, G., B. Singh, and R. V. Singh. 1977. Wind tunnel studies on the effect of chemical mulches on evaporation from water and soil surfaces. Indian J. Agric. Engr. 14(1):28-32.

The effect on evaporation due to application of chemical mulches namely hexadecanol and octadecanol on free water surfaces and soil surfaces under shallow water table conditions was studied in a wind tunnel. No evaporation was observed at wind velocities below 3.1 km/h. Evaporation from free water surface was reduced to 50% approximately by the use of hexadecanol at wind velocities less than 16.65 km/h.

4. Gardner, H. R., and R. J. Hanks. 1966. Evaluation of the evaporation zone in soil by measurement of heat flux. SSSAP 30:425-428.

Heat-flux plates were used for measuring the movement of heat during evaporation in bare columns of soil with and without surface treatment with hexadecanol. The evaporation zone was about 1 cm thick, and equal rates of evaporation occurred in control soil and after hexadecanol treatment when the zones were at the same depth after the treated zone had been passed. Diurnal fluctuation of applied heat energy slowed the evaporation process, without fully eliminating movement of the evaporation zone.

5. Gradwell, M. W. 1968. The effect of grass cover on overnight heat losses from soil and some factors influencing the thermal conductivities of soils. New Zeal. J. Sci. 11:284-300.

Masses of sensible heat from the 0-40 cm layers of silt loams during radiation (frost) nights were calculated; the effect of short grass cover in reducing these losses was greater than that of soil loosening. Thermal conductivity (calculated from thermal diffusivity arrived at from the differences between maximum and minimum temperatures at various depths) was doubled by an absolute increase₃ of 25% moisture content or by an increase of 0.3 g/cm³ dry-weight density.

6. Greb, B. W. 1966. Effect of surface-applied wheat straw on soil water losses by solar distillation. SSSAP 30:786-788.

Soil surface application of 1,120, 2,240, and 3,360 kg/ha of wheat straw (Triticum aestivum L.), equivalent to 30, 60, and 90% soil surface coverage, reduced water losses from a wet soil surface by solar distillation 16, 33, and 49%, respectively, for a 20-day period compared with no straw. A surface application of 6,720 kg/ha, or 180% soil coverage, reduced soil water loss only slightly more than did the 3,360 kg/ha application. Water losses and effectiveness of straw tended to diminish with time as soil water was depleted. The presence of clear plastic canopy increased daily soil temperature maximums at 2.5 cm depth by approximately 11.5°F. Maximum daily soil temperatures, both with and without plastic, were reduced nearly 3°F/1,120 kg per ha of added straw. The results suggest part of the mechanism by which more soil water is conserved under stubble mulch summer fallow than with clean fallow.

7. Guild, W. R. 1950. Note on heat transfer at the soil surface. J. Meteor. 7:140-144.

Heat balance and eddy diffusion are determined for nocturnal cooling in desert regions of Arizona. Convection accounts for as much as half of the transfer.

8. Hammel, J. E. 1979. Modeling tillage effects on evaporation and seed zone water content during fallow in Eastern Washington. Ph.D. dissertation. Washington State Univ., Pullman. 65 pp.

A mathematical model was developed to investigate the dynamic aspects of soil water flow in upper soil layers with emphasis on simultaneous heat and water vapor transport. The model was verified using measured meteorological variables for a 15 day period, soil hydraulic and thermal properties, and soil moisture and temperature distributions to the 60 cm depth.

9. Lyles, L., and B. E. Allison. 1979. Wind profile parameters and turbulence intensity over several roughness element geometrics (wind erosion of soil particles, water evaporation, and transport of gases). ASAE Trans. 22(2):334-338.

Estimating equations were developed for the mean velocity profile parameters in the logarithmic law and for longitudinal turbulence intensity. The estimates were based on wind tunnel measurements over several roughness element shapes, sizes, heights, and geometrical patterns. All the prediction equations had correlation coefficients >0.90 . These equations have application in wind erosion of soil particles, water evaporation, and transport of gases. They only apply to a particular set of conditions, which indicates a need for more universally applicable equations for the profile parameters and turbulence intensity.

10. Papendick, R. I., M. J. Lindstrom, and V. L. Cochran. 1973. Soil mulch effects on seedbed temperature and water during fallow in Eastern Washington. SSSAP 37:307-314.

Temperature and water content data are presented for depth and time under various tillage operations. The data show fairly clearly that mulching is very important for both heat and water transport.

The authors elucidate the mechanisms--qualitatively, at least: enhancement of seed zone water is dependent on water mulch and pore continuity from subsoil. The surface water mulch is inextricably tied to heat transfer. Both heat and moisture transfer are correlated with soil porosity. Both transport resistances increase with increasing porosity until sufficient vapor transport (turbulent flow) occurs so that the effective resistance decreases.

11. Schuepp, P. H. 1977. Turbulent transfer at the ground: on verification of a simple predictive model. Boundary Layer Meteor. 12:171-186.

Observations on heat transfer from ground-based plates and evaporation from free water surfaces in the laboratory and in the field are compared with predictions from a simple model. The model relates the convective transfer coefficient (or boundary-layer resistance) at any point on a surface to the momentum transfer (friction velocity) in the boundary layer immediately above it and should be applicable to practically any soil surface, open or vegetated.

Heat-transfer data showed a standard deviation of 25% between predictions and observations. Evaporation data provided only order-of-magnitude confirmation of the model because of uncertainty in effective water vapor density above free-water surfaces.

C. Advection

1. Itier, B., and A. Perrier. 1977. An analytical model of advection. Preprint of 13th Agr. and For. Meteorol. Conf. AMS. pp. 31-32.

Their model splits climatic regimes into upstream and downstream portions. Given values for height, constant roughness, windspeed profile, and variation in diffusivity, the model predicts the flux decrease with downstream distance.

2. Philip, J. R. 1959. The theory of local advection. J. Meteor. 16:535-547.

This paper is the first of a series in which the theory of local advection (the exchange due to horizontal heterogeneity) of energy and moisture will be developed and applied to a number of problems of practical and theoretical interest. The paper provides an introduction to the practical implications and physical basis of local advection, but it is mainly devoted to developing methods of analysis to be applied in later papers. The treatment aims to provide simple and rapid numerical procedures for the solution of advection problems.

Methods are given for solving the two-dimensional atmospheric-diffusion equation subject to "concentration," "flux," and "radiation" types of boundary conditions. Appendices give discussions of the properties of the functions entering the solutions and provide simple means for their computation. Extensive tables of the relevant functions are given for the case $m = 1/7$, $n = 6/7$ (m , n being the exponents in the power-law approximations to the vertical profiles of mean wind speed and eddy diffusivity, respectively).

The rudiments of a quantitative theory of advective inversion are developed, expressions being obtained for the equation of the inversion surface, and for the maximum height and downwind extent of the inversion.

3. Rosenberg, N. H. 1969. Advective contribution of energy utilized in evapo-transpiration by alfalfa in the central Great Plains. Agron. J. 61:879-886.

This paper presents information about the variation of evapotranspiration (ET) with diurnal and seasonal changes. The Penman and van Bavel methods for estimating ET are shown to yield incorrect estimates because the methods only use local atmospheric conditions.

4. Taylor, P. A. 1970. A model of air flow. Boundary Layer Meteor. 1:18-39.

A numerical model of airflow in the lowest 50-100 m of the atmosphere above changes in surface roughness and temperature or heat flux has been developed based on boundary layer approximations, the Businger-Dyer hypotheses for the non-dimensional wind shear and heat flux and a mixing length hypothesis.

Results have been obtained for several situations, in particular, airflow with neutral upstream conditions encountering a step change in surface temperature or heat flux with no roughness change. In these cases large increases in shear stress at the outer edge of the internal boundary layer are predicted. The case of unstable upstream flow encountering a step change to zero heat flux is also considered.

Two situations that may be encountered near the shores of the Great Lakes are considered.

III. THERMAL AND FLUID PERMEABILITY PROPERTIES OF UNSATURATED POROUS MEDIA

A. Effect of temperature on thermal and fluid permeability properties.

1. Chahal, R. S. 1964. Effect of temperature and trapped air on the energy status of water in porous media. Soil Sci. 98:107-112.

Larger values of temperature coefficient of pressure potential than would be expected from considerations of the surface tension-temperature relationships that have been reported can be explained in terms of changes in the volume of trapped air with temperature. Larger values of temperature coefficients would be expected to occur in fine-textured in comparison to coarse-textured soils, simply because there would be a larger volume of air trapped in them before or during desorption. Also, it is to be noted that the volume of trapped air in a system is not limited to whatever there is at saturation, depending upon cracks, cavities, and other deformities in the particles, but it is also related to the desorption curve. Any theoretical consideration of the effect of trapped air must take into account this relationship.

2. Chahal, R. S. 1965. Effect of temperature and trapped air on matric suction. Soil Sci. 100:262-266.

The results indicate that trapped air in the soil-water system may cause changes in matric suction many times larger than would be expected from the theory of surface-tension to temperature relationship, and therefore, point out the importance of temperature effects during laboratory and field investigation of the soil-moisture to energy relationships. Results of the present investigation lend further support to the remarks made earlier about soil-water to energy relationships as measured with the pressurized equipment.

3. Flocker, W. J., M. Yamaguchi, and D. R. Nielsen. 1968. Capillary conductivity and soil water diffusivity values from vertical soil columns. Agron. J. 60:605-610.

Water evaporated from vertical columns of Yolo loam, 100 cm long, for 200 days at ambient temperatures of 10, 20, or 30 $\pm$ 1 C. distribution of soil water was determined by gamma-ray attenuation at 5 cm intervals, concomitantly with determinations of soil water pressure. Values of capillary conductivity based upon measured fluxes and hydraulic gradients are presented, along with values of diffusivity based upon fluxes and water content gradients.

4. Gardner, R. 1955. Relation of temperature to moisture tension of soil. Soil Sci. 79:257-265.

A study was made to determine the effect of temperature on soil moisture tension of three soils at constant moisture percentages. Data are presented which show a consistent tendency for tension to decrease at approximately 0.008 atmosphere per degree rise in temperature within the range of 0 to 50 C. The possibility of temperature-induced moisture percentage gradients in the soil at the time the tensiometers were read raises a question regarding the part of the tension variation that can be attributed to the direct effect of temperature on soil moisture tension and the part that should be attributed to the indirect effect through temperature-induced moisture percentage gradients. Regardless of this question, however, the data show that variations due to temperature may be so great that they cannot well be ignored in studies of soil-plant relationship or soil moisture movement.

5. Jackson, R. D. 1963. Temperature and soil-water diffusivity relations. SSSAP 27:363-366.

Water held in porous media is under the influence of a number of forces. The energy of the water in these force fields may be described with the aid of the potential theory of classical physics or with the aid of classical thermodynamics. Desorption curves were determined for two sand fractions and a silt fraction at 48 and 44 C. Temperature coefficients of the potential calculated on the basis that surface tension-radius of curvature was the principle water-holding mechanism were compared with the experimental observations. The results indicated that in the fine sand (53-74 μ) and silt (13.5-18.0 μ) other mechanisms besides surface tension-radius of curvature may be involved in holding the water in the porous medium.

6. Jensen, R. D., M. Haridasan, and G. S. Rahi. 1970. The effect of temperature on water flow in soils. Water Resour. Res. Inst., Mississippi State Univ., State College. 43 pp.

This paper investigates the dependence of hydraulic conductivity, pressure head and water diffusivity in soil on temperature.

7. Peck, A. J. 1960. Change of moisture tension with temperature and air pressure: theoretical. Soil Sci. 89:303-310.

Classical treatments of the problems of water held in unsaturated soil have given scant attention to the effects of air bubbles trapped within the liquid-phase water. The present paper provides a detailed theoretical analysis of

the effects of trapped air, with special reference to the influences of temperature and external air pressure on moisture tension. The following effects are taken into account: (a) changes of bubble volume with temperature and air pressure, (b) changes of the volume of a fixed mass of water with temperature, and (c) changes of the surface tension of water with temperature.

The analysis enables the prediction of changes of moisture tension with external air pressure. Under conditions that are not uncommon, the rate of change of tension with temperature and air pressure becomes very large.

A very important consequence of these results is that, in general, moisture content-tension curves determined by suction and pressure methods will be different. These discrepancies suggest the need for a critical examination of pressure plate and pressure membrane methods.

The analysis also suggests a method of determining the volume of air trapped in an unsaturated soil.

In a paper to follow, the author's experimental results will be presented and discussed with the aid of the theory developed here.

8. Sepaskhah, A. R., and L. Boersma. 1979. Thermal conductivity of soils as a function of temperature and water content. SSSAJ 43(3):439-444.

Apparent thermal conductivities of loamy sand, loam, and silty clay loam soils were measured with a cylindrical heat probe at several water contents and temperatures. Values of λ were also calculated with the deVries model. Results show that the model may be used satisfactorily to calculate λ . However, improvements may be needed to account for the enhancement of vapor transfer at high temperatures (45°C) in medium to fine textured soils. The heat probe method of measuring the thermal conductivity requires a correction factor to allow for errors due to the entrapment of air at the surface of the heat probe, when used at water contents ranging from 0 to about 30% of saturation.

The contribution of vapor distillation to thermal conductivity was analyzed by comparing λ at 25 and 45°C. The ratios $\lambda_{45}/\lambda_{25}$ were nearly equal to unity when <5% of total pore space was filled with water. The ratios increased, due to vapor distillation, as water filled the pores and reached maximum values of 2.17, 2.71, and 2.96 when 22, 27, and 35% of total pore space was filled with water, corresponding to soil water potentials of -0.8, -9.0, and -8.0 bars for loamy sand, loam, and silty clay loam, respectively. As the water content increased further, the ratios decreased and approached unity when >50% of total space was filled with water.

9. Wilkinson, G. E., and A. Klute. 1962. The temperature effect on the equilibrium energy status of water held by porous media. SSSAP 26:326-329.

Water at -2 mb pressure was absorbed by horizontal columns of three soil materials of different texture at temperatures of 5.0, 13.5, 24.0, 33.0, and 42.5 C. From the water content distributions, soil-water diffusivities were calculated, assuming applicability of diffusion theory. Twofold to threefold differences were observed between diffusivities at 5.0 and 42.5 C. The water content distribution curves were not measurably influenced by temperature, thus, in the equation used to calculate diffusivity, the time required for water to penetrate the given distance was the only term differing with temperature.

The temperature dependence of the ratio of surface tension to viscosity appears to be the dominant factor in describing the temperature dependence of soil-water diffusivity. The square root of the surface tension viscosity ratio adequately accounts for the temperature dependence of penetrability and sorptivity.

B. Measurements of component fluxes.

1. Hutcheon, W. L. 1958. Moisture flow induced by thermal gradients within unsaturated soils. In H. F. Winterkorn (ed.), Water and its conduction in soil. Highway Res. Spec. Rep. 40.

Qualitative and quantitative data have been presented to illustrate the character and magnitude of thermally induced moisture flow within unsaturated soils, with particular reference to soil moisture conditions characteristic of moisture tensions in excess of 0.1 atmospheres.

2. Jackson, R. D., B. A. Kimball, R. J. Reginato, S. B. Idso, and F. S. Nakayama. 1975. Heat and water transfer in a natural soil environment. In D. A. deVries and H. H. Afgan (eds.), Heat and mass transfer in the biosphere. Vol. I. Transfer processes in the plant environment. Scripta, Washington, D. C. pp. 67-76.

Experiments conducted in a natural field soil yielded heat and water fluxes as functions of depth and time which were compared with theoretical values. Calculated and measured water fluxes agreed well at night but diverged during the day. Calculated and measured heat fluxes agreed well at night but diverged particularly at the shallow depths at other times.

3. Jackson, R. D., R. J. Reginato, B. A. Kimball, and F. S. Nakayama. 1974. Diurnal soil-water evaporation: comparison of measured and calculated soil-water fluxes. SSSAP 38(6):861-866.

The J. R. Philip and D. A. deVries theory and the "isothermal" theory were used to predict diurnal soil water fluxes near the soil surface. The predicted values were compared with those obtained by measurements of soil-water content, soil temperature, and evaporation. Previously measured soil-water diffusivities were used in the theoretical calculations. The thermal vapor diffusivities were calculated using both the "simple" and the "complete" theory of Philip and deVries.

4. Jury, W. A., and B. B. Bellantuoni. 1976. Heat and water movement under surface rocks in a field soil. 1. Thermal effects. SSSAJ 40(4):505-509.

Results of a field experiment designed to evaluate the effect of surface rocks on soil heat flow are presented. Temperature observations were made by thermocouples at 12 locations under and adjacent to rocks placed over bare soil. Continuous readings were taken for 24 and 48 hour intervals using seven different kinds of rock cover, ranging from large granite slabs to gravel piles. A simulation model was constructed to describe two-dimensional heat flow in a uniform soil with a rectangular rock on the surface. Using the measured surface temperature as boundary values and a soil thermal conductivity corresponding to the mean daily soil surface temperature, the simulation model adequately reproduced the observed temperatures under and adjacent to the rock.

5. Jury, W. A., and B. B. Bellantuoni. 1976. Heat and water movement under surface rocks in a field soil. 2. Moisture effects. SSSAJ 40(4):509-513.

Field and laboratory experiments were conducted to determine the effect of large surface rocks on water content changes in bare soil. Results demonstrated that water was retained longer under the rock than under adjacent bare soil, and suggested that horizontal differences in temperature can cause lateral migration of water vapor toward the rock.

6. Kimball, B. A., R. D. Jackson, F. S. Nakayama, S. B. Idso, and R. J. Reginato. 1976. Soil-heat flux determination: temperature gradient method with computed thermal conductivities. SSSAJ 40(1):25-28.

Surface soil-heat fluxes were determined for 6 days in a field of Avondale loam using the null-alignment method, the combination method, and four variations of the temperature gradient method with thermal conductivities computed from the deVries theory for particular reference depths. For all methods, calorimetry was used to obtain the surface flux from the flux determined for the reference depth.

7. Kimball, B. A., R. D. Jackson, R. J. Reginato, F. S. Nakayama, and S. B. Idso. 1976. Comparison of field-measured and calculated soil-heat fluxes. SSSAJ 41(1):18-25.

Soil-heat fluxes calculated using deVries' theory were compared with those experimentally determined in a field of Avondale loam at Phoenix, Arizona, on 6 days representing different seasons of the year. A fair agreement between measured and computed fluxes was obtained only after modifying the air shape factor curve and ignoring heat transfer due to water vapor movement. The omission of the latter implied that heat transfer by pure conduction was most important and that thermal and isothermal vapor fluxes exactly canceled during the day and were insignificant at night.

8. Mathes, R. K., and H. D. Bowen. 1963. Water vapor transfer in soil by thermal gradients and its control. ASAE Trans. 6:244-248.

This paper concluded that water vapor movement was dependent on the thermal curvature, the product of the diffusion coefficient gradient and the thermal gradient, and the thermal gradient squared. It also suggested that evaporation or condensation can be induced in a region if a temperature gradient and moisture exist in the region.

9. Rose, D. A. 1963. Water movement in porous materials. 2. The separation of the components of water movement. Brit. J. Appl. Phys. 14:491-496.

Vapor conductivities h_v estimated from hydrogen diffusion in porous solids of decreasing moisture content rise slowly from zero at saturation to a maximum near dryness, and then fall rapidly again to zero at complete dryness. Liquid conductivities computed from pore size distributions, and estimated from previous water transport experiments, decrease continuously from saturation to complete dryness. For each of six different porous materials, the proportion of water movement as vapor falls rapidly as θ increases, becoming negligible in the moisture range important for plant growth.

Six stages in the wetting of a porous system can be identified. Once adsorption is complete, unimpeded vapour transfer occurs, together with liquid movement. The latter takes the form, as θ increases, of distillation (a discontinuous liquid phase acting as a short-circuit for vapor), followed by surface creep (flow in continuous liquid films with k_l increasing as $(\theta/\theta_T)^n$ with $n \approx 3$) and unsaturated hydraulic flow. Finally, at saturation, $k_v = 0$, and saturated hydraulic flow obtains.

10. Rose, D. A. 1966. Water transport in soils by evaporation and infiltration. Symposium on water in the unsaturated zone (UNESCO, Wageningen).

The equation of vertical flow is adapted to describe the evaporation of water from, and the uptake of water by a deep uniform soil free from vegetation. The theory is examined by laboratory measurements on long columns of 0.5 - 1.0 mm aggregates of three soils from Rothamsted and Woburn.

Evaporation from soils initially at a uniform moisture content--field capacity--decreases as the square root^{1/2} of time, and causes moisture profiles invariant with zt (where z = depth below soil surface and t = time), which differ for each soil. Drying diffusivities calculated from these profiles have an unusual shape, decreasing to a minimum and then increasing as the moisture content (θ) falls from field-capacity to air-dryness. The uptake of water treated in the same way yields wetting diffusivities which also display a prominent minimum as θ increases from air-dryness to field capacity.

The shape of these diffusivity-moisture content relations is discussed and the concept of sorptivity adapted to yield a parameter specifying the water supplying power of a soil when evaporation is limited by soil moisture.

11. Rose, C. W. 1968. Water transport in soil with a daily temperature wave. I. Theory and experiment. Aust. J. Soil Res. 6:31-44.

The theory of water transport in soil is summarized for situations where flux can arise from gradients in temperature as well as in soil water suction and gravitational potential. The general differential equation describing such transport was doubly integrated to provide a form of equation suitable for the analysis of experimental data.

An experiment is described in which profiles of water content and temperature were measured in the top 15 cm of bare soil for a period of 6 days and nights following saturation. High daytime radiation conditions resulted in maximum temperature gradients associated with the¹ daily temperature wave in the soil approaching $10\text{ }^{\circ}\text{C dm}^{-1}$, thus favoring vapor phase in relation to liquid phase water transport. The basic soil data necessary for the quantitative analysis of moisture fluxes into liquid and vapor components were determined.

When water content at the eight sampling depths was plotted against time from saturation there was superimposed on a regular decline in water content an approximately sinusoidal fluctuation of daily period with a maximum around 0400-0600 hr local time and a minimum about 1600 hr. These results may be interpreted as indicating the existence

of a vapor flux in the soil of magnitude comparable with the liquid flux. This hypothesis is examined quantitatively in Part II.

12. Sutor, J., and V. Novak. 1968. The effect of temperature gradient on movement of soil water. Int. Congr. Soil. Sci. Trans. 9th. (Adelaide, Australia) I:85-93.

The influence of a temperature gradient within a soil column upon the loss of water from the exposed surface of one end was investigated experimentally using a packed column 15 cm long. The surface was maintained cooler than the soil beneath it, and a set of water content profiles in the soil was obtained for the conditions; initial water content of the soil $0.21 \text{ cm}^3/\text{cm}^3$, temperature gradient 1 C/cm , mean temperature 29 C , bulk density of the soil 1.20 g/cm^3 and seven relative vapor pressures at the evaporating surface between 0.01 and 1.0.

Similar experiments were performed with isothermal water movement in the soil. The evaporation was about three times as large, when the temperature gradient assisted the flow of water to the surface. By computation, using the experimentally determined values of temperature gradient and water flux, the coefficient for water flow under a temperature gradient was determined to be $3 \times 10^{-5} \text{ cm}^2/\text{sec/deg C}$ with a maximum at a water content of $0.21 \text{ cm}^3/\text{cm}^3$.

C. Measurements of component diffusivities and conductivities.

1. Cary, J. W. 1965. Water flux in moist soil: thermal versus suction gradients. Soil Sci. 100:168-175.

Simultaneous flows of heat and moisture were measured across a loam soil sample under various thermal and pressure gradients. The data were collected in a manner such that the liquid and vapor flow components could be separated and such that the liquid component could further be divided into that flowing due to pressure differences and that due to thermal differences. It was found that a temperature gradient of 0.5 C/cm at a soil moisture suction of 5 cm of Hg would move as much water through the soil as a pressure gradient of 2 cm of water per cm. At a suction of 34 cm of Hg this temperature gradient was equivalent to 250 cm of water per cm. Vapor transport through the soil was several times greater than could be predicted with Fick's law and the diffusion of water vapor in air, even though at soil moisture suctions of less than 18 cm. Hg this vapor transfer accounted for less than half of the net thermally driven water flux. A linear thermodynamic moisture flux equation was developed in terms of commonly used flow coefficients and variables. This equation predicted fluxes in agreement with experiment observations. A mechanism for unsaturated flow in the liquid phase through soil is proposed.

2. Cassel, D. K., D. R. Nielsen, and J. W. Biggar. 1969. Soil-water movement in response to imposed temperature gradients. SSSAP 33(4):493-500.

Redistribution of soil water within insulated, uniformly packed, horizontal samples of unsaturated Columbia fine sandy loam at several soil-water contents was studied in response to imposed temperature gradients ranging from 0.5 to 1.0 C/cm; observed net water flux showed acceptable agreement with that predicted by theory of J. R. Philip and D. A. deVries; Fick's law and modified Taylor-Cary irreversible thermodynamic equation both underpredicted observed fluxes.

3. Fritton, D. D., D. Kirkham, and R. H. Shaw. 1970. Soil water evaporation, isothermal diffusion, and heat and water transfer. SSSAP 34(2):183-189.

Experimental and theoretical curves of cumulative soil-water evaporation vs. time, as well as the resulting depth-water distributions, were compared for wind- and radiation-induced evaporation in soil columns. It was found that the isothermal diffusion theory would predict cumulative evaporation for wind and radiation treatments but would not predict water distributions once vapor movement within the soil was significant. A heat and water transfer equation fit an experimental water distribution curve when vapor movement was dominant.

4. Gardner, H. R., and R. J. Hanks. 1966. Evaluation of the evaporation zone in soil by measurement of heat flux. SSSAP 30:425-428.

Heat flux plates were employed to measure the movement of heat within a soil column during the process of water evaporation. The movement of the zone of evaporation was followed for a bare soil column with and without a surface treatment of hexadecanol. Under these conditions the evaporation zone was approximately 1 cm thick, and equal rates of evaporation were found for the check and hexadecanol treatments when the zones were at the same depth after the treated zone had been passed. Diurnal fluctuation of applied heat energy slowed the evaporation process but the evaporation zone movement was still evident.

5. Gee, G. W. 1966. Water movement in soils as influenced by temperature gradients. Ph.D. Thesis. Washington State Univ., Pullman. 77 pp.

This thesis makes a qualitative examination of the mechanisms of flow in a soil column.

Results from temperature gradient experiments indicate that linear equations with constant coefficients cannot be

used to describe the flow induced by thermal gradients. Taylor-Cary and Philip-deVries were not good estimators of flow with time.

6. 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.

The effect of a temperature gradient on the movement and distribution of water in soil has been examined in closed columns of soil for a wide range of initial water contents. Small amounts of soluble salts acting as a tracer served to distinguish between liquid and vapor movement.

In all except the wettest and driest columns of soil, there was a transfer of water toward the colder end of the column and of salts toward the hotter end.

This result is considered to be due to a net transfer of water from hot to cold, in which water evaporating from the hotter soil moves as a vapor into colder soil, where it condenses and returns as a liquid when a favorable gradient of pressure potential has been established. It is shown from theoretical considerations that, when the soil is sufficiently wet to permit liquid flow, a state of equilibrium cannot be reached, and continuous circulation of the soil water takes place.

The maximum net transfer of water from the hot to the cold end occurred in columns the initial water contents of which were approximately one third of the moisture equivalent.

Movement of water in the liquid phase is shown to have occurred at a low water content, which corresponded to a pressure potential of -0.8×10^7 ergs/gm (pF 3.9).

7. Hadas, A. 1968. Simultaneous flow of water and heat under periodic heat fluctuations. SSSAP 32:297-301.

Theoretical predictions of thermally induced moisture transfer have been compared to measured quantities. Soil samples at different initial moisture contents were subjected to sinusoidal heat waves having an amplitude of 6 C, and various periods of time. Temperatures were measured at various distances from the heat source, using copper-constantan thermocouples and recording the emf output.

The moisture distribution after the 1st, 2nd, 6th, and 16th cycle was determined gravimetrically.

The measured net transfer quantities were found to be larger than the theoretically predicted values by factors ranging from 0.8 to 8.0. The discrepancy between predicted and measured values of net transfer may be attributed to the incompleteness of the theoretical models.

8. Hadas, A. 1969. A comparison between predicted and measured values of the thermal conductivity of a moist soil under steady and fluctuating thermal regimes. Israel J. Agric. Res. 19:151-160.

Predicted theoretical thermal-conductivity values agreed with experimentally measured values under steady state conditions, but under thermally fluctuating conditions heat conductivity was greater than predicted; differences increased with moisture content and pulse duration. The discrepancies are explained by assuming vapor transfer by thermally induced mass flow.

9. Hadas, A. 1975. Drying of layered soil columns under nonisothermal conditions. Soil Sci. 119:143-148.

An experiment was carried out to clarify the effects of steady and fluctuating nonisothermal conditions on evaporation from soil profiles having an aggregated top layer of different thicknesses and aggregate sizes.

It was found that there are optimal ranges of aggregate size and depth of the top layer which result in reduced evaporation losses from layered soils. These results differ from theoretical predictions and the discrepancies are discussed and explained. Under fluctuating nonisothermal conditions the water losses were reduced to a larger extent than under steady nonisothermal conditions. These results showed that employing an isothermal diffusion model to nonisothermal evaporation will introduce large errors into the predictions. Furthermore, any evaporation experiment carried out to check or evaluate the process under field conditions should be carried out under simulated field conditions.

It is suggested that to conserve water in the soil profile by tillage operations, the soil should be tilled prior to the introduction of the water into the soil.

10. Jackson, R. D. 1964. Water vapor diffusion in relatively dry soil. I. Theoretical considerations and sorption experiments. SSSAP 28:172-176.

Water transfer was measured in three soil materials at water content ranges corresponding to a relative pressure range of 0.97 to about 0.05. At these water contents vapor flow is probably the predominant transfer mechanism. The data were obtained from a transient sorption experiment and diffusion coefficients were calculated from analyses of water content distribution curves. The evaporation-condensation reaction between the liquid and vapor phases was treated by using simultaneous diffusion-reaction theory. Within the water content ranges used the diffusion coefficients increase and then decrease with increasing water content.

11. Jackson, R. D. 1964. Water vapor diffusion in relatively dry soil. II. Desorption experiments. SSSAP 28:464-466.

Water vapor diffusion during desorption was measured in three soil materials at water contents corresponding to a relative pressure range of 0.97 to about 0.05. The data were obtained from a transient desorption experiment conducted at 27°C. Diffusivities, calculated from analysis of water content distribution curves, increase and then decrease with increasing water content. The desorption diffusivity-water content curves exhibit a less distinct maximum, with the maximum occurring at a higher water content than do the sorption diffusivity-water content curves.

12. Jackson, R. D. 1964. Water vapor diffusion in relatively dry soil. Part III. Steady state experiments. SSSAP 28:467-470.

Steady-state water vapor diffusion coefficients were measured for sorption and desorption on Adelanto loam and Pachappa loam at 27°C. The sorption diffusivities for both soil materials increase, and then decrease, with increasing water content. The desorption diffusivities for Adelanto loam are constant throughout the water content range studied, whereas the desorption diffusivities for the Pachappa loam exhibit a maximum similar to the sorption diffusivities. For both soil materials the transient and the steady-state diffusivity-water content curves have nearly the same shape and magnitude. Diffusivities calculated from adsorption isotherm data are similar to the measured diffusivities. Comparison of data from the three methods indicates that liquid diffusion becomes noticeable at water contents corresponding to relative pressures of 0.5 to 0.7 for Adelanto and 0.8 to 0.9 for Pachappa.

13. Jury, W. A. 1973. Simultaneous transport of heat and moisture through a medium sand. Ph.D. Thesis. Univ. of Wisconsin, Madison.

Steady state and transient equations for simultaneous flow of heat and moisture are developed in this thesis.

The author concludes that temperature gradients must be included in water transport models, particularly if the thermal diffusivity is large. Secondly, the surface layer of soil must be coupled to both the soil below and the atmosphere above.

More works remain to be done before a workable model for moisture transport in presence of a temperature gradient is available for field use.

14. Jury, W. A., and E. E. Miller. 1974. Measurement of the transport coefficients for coupled flow of heat and moisture in a medium sand. SSSAP 38(4):551-557.

Primary and cross-coupling transport coefficients were measured for simultaneous flow of heat and moisture in a medium sand in the liquid-dominated regime of moisture flow. Results were compared with the theoretical models of Philip and deVries and showed a good agreement with deVries' model for thermal conductivity, and a measured value for the thermal moisture diffusivity over five times in excess of the calculated values.

15. Letey, J. 1968. Movement of water through soil as influenced by osmotic pressure and temperature gradients. Hilgardia 39:405-418.

The equation used to describe steady-state water flow through a soil in response to hydraulic pressure (or suction in unsaturated soil), osmotic pressure, and temperature gradients is $J_v = L_{vv} \text{ grad } P + L_{vD} \text{ grad } \pi - L_{vq} \text{ grad } T$. Published research results indicate that L_{vq} is relatively independent of suction when suctions are greater than about 0.06 bar, becomes quite small when suctions are lower than 0.06 bar, and increases in a predictable manner as the average temperature increases. A value of $3.0 \times 10^{-3} \text{ cm deg}^{-1} \text{ hr}^{-1}$ is suggested for L_{vq} for estimates of water flow in response to temperature gradients. L_{vD} can be considered to be zero for suctions less than about 0.25 bar except under conditions of very high clay content. The value of L_{vD} is not likely to be greater than 10% of L_{vv} (3% is suggested for approximate calculations) at suctions between 0.25 and 1 bar. L_{vD} would more nearly equal L_{vv} at greater suctions, but no data are available on this relationship at higher suctions. The suggested values of L_{vq} and L_{vD} are based on relatively few data and are to be considered as approximate. Additional data are required for more specific values for these coefficients.

16. Moore, N. R., Jr. 1973. Simultaneous heat and mass transfer in moist porous media. PhD. Thesis. Univ. of Arkansas.

The convective heat and mass transport phenomena occurring in granular, porous media that were partially saturated with a wetting liquid were experimentally investigated. Experiments were performed with small glass beads and with granular aluminum oxide as the porous media which contained liquid water. The results show the effects of temperature, heat flux, and initial liquid content on the thermal conductance of the media and indicate the character of the convective transport phenomena.

It was found that the thermal conductance of a porous medium whose particle size is small enough for the surface tension effects of a wetting liquid to dominate depends on the amount of liquid contained in the medium and may be significantly increased by convective transport. A nonuniform temperature field in the medium induces vapor diffusion which may result in simultaneous diametrically opposed liquid flows of relatively large magnitude in the same region of the medium, or in a nonuniform liquid distribution that under some conditions proceeds to local drying.

17. Scotter, D. R. 1976. Liquid and vapor phase transport in soil. Aust. J. Soil Res. 14:33-41.

Measurements of the desorption soil water diffusivities for vapor and liquid phase flow in a loamy sand and a clay at water contents covering the range between air dry and permanent wilting point are reported. The liquid and vapor components were separated using soil columns maintained at various air pressures. Vapor diffusivities were also calculated from desorption isotherms. At some water contents the measured vapor diffusivities were up to an order of magnitude greater than the calculated values. Liquid phase flow was detected in soils nearly air dry, suggesting some solute transport can occur under such conditions. The liquid and vapor diffusivities were found to be of comparable magnitude over a range of water contents, rather than each cutting off sharply as has been suggested in the literature. The measured component diffusivities at 298K were used to calculate the total diffusivities at other temperatures, using the temperature coefficients of the basic physical parameters governing liquid and vapor flow, and good agreement was found between measured and calculated values.

18. Weeks, L. V., S. J. Richards, and J. Letey. 1968. Water and salt transfer in soil resulting from thermal gradients. SSSAP 32:193-197.

Rate equations developed from the theory of thermodynamics of irreversible processes were used to analyze water transfer induced by thermal gradients along a sealed cylindrical horizontal column of Pachappa soil. Functional relations to calculate values of coefficients in the flow equations were determined from experimental data which were obtained during steady-state flow conditions. The relations were used to compute values of coefficients in order to calculate flow rates past selected positions along the column at various times during transient flow periods. Suction head (h) values were maintained in a range that could be measured with tensiometers which were located at three positions along the column. Liquid water transfer toward the warm end was assumed to be due to an

induced suction head gradient. This transfer was confirmed by sectioning the column at the end of the experiment and measuring the concentrations of Na^+ and Cl^- and the EC of saturation extracts prepared from the section.

D. Data or techniques which may allow prediction of thermal properties.

1. deVries, D. A. 1963. Thermal properties of soils. In W. R. van Wijk (ed.), Physics of plant environment. pp. 210-236.

This is the most useful presentation of deVries' theoretical technique for calculating soil thermal properties. The paper includes sections on heat capacity and thermal conductivity of soils, influence of moisture movement, and examples of thermal conductivity calculations for sand, clay and peat.

2. Domby, C. W., and H. Kohnke. 1956. The influence of soil crust on gaseous diffusion. SSSAP 20:1-5.

A method is described for forming crusts on soils in special containers in the laboratory using artificial rainfall, and for measuring gaseous diffusion through these soils without further disturbance of their structure.

Unless the surface is completely impervious, the rate of diffusion through a soil does not depend solely on the properties of this layer. When various proportions of the soil surface were sealed with paraffin, the rate of diffusion was not reduced in proportion to the area covered.

The effects of soil crusts upon diffusion depended to a very great extent on the soil moisture tension. There were no significant differences in the rates of diffusion through air-dry silt loam having different degrees of crusting. Surface crusts on this soil restricted diffusion only at low moisture tensions; the wetter the soil, the greater the influence of the crust on diffusion. Except in the very wet range, even minor variations in soil moisture tension had more influence on diffusion than did the presence of a thin surface crust.

A saturated soil crust could provide a very effective seal against diffusion at the time when soil aeration is most critical, i.e., during and for a short time after a rain or irrigation. Except during a period of complete sealing, however, crusted soils actually may permit more rapid diffusion and be better aerated than mulched or other uncrusted soils, because of differences in moisture content.

3. Fosberg, M. A. 1977. Heat and water transport properties in conifer duff and humus. USDA-FS Res. Pap. RM-195.

This paper represents an attempt to develop similitude relationships between water and air permeability properties and soil horizon depth, bulk density, and porosity.

The data are presented in a table but there isn't much quantitative information in the text about how well the correlations fit. A qualitative inspection of the data shows that the field measured soil properties are not sufficient to characterize the permeability properties.

4. Gerayzade, A. P. 1974. Linear relations between the thermal diffusivity coefficient and the particle-size composition of soil. *Sov. Soil Sci.* 6:621-624.

The thermal diffusivity coefficients of soils of four groups are determined. The fact that the thermal diffusivity coefficient increases linearly with decreasing content of physical clay in soils is confirmed. An empirical formula of this relation is derived.

5. Ghildyal, B. P., and R. P. Tripathi. 1971. Effect of varying bulk densities on the thermal characteristics of lateritic sandy clay loam soil. *J. Indian Soc. Soil Sci.* 9(1):5-10.

The specific heat of the soil was $0.23 \text{ cal.g}^{-1}\text{.}^{\circ}\text{C}^{-1}$. The unsteady state method of studying heat conduction in one dimension was used to determine the relationship of thermal conductivity to, and the relationships between, bulk density, volumetric heat capacity, degree of saturation, void ratio, porosity, and thermal diffusivity of the soil.

6. Globus, A. M., and A. V. Aref'ev. 1971. Dependence of thermal properties of soils on moisture pressure and thickness of the water film. *Pochvovedenie* 11:100-104.

At a particular volume weight, the relation between the coefficient of thermal conductivity of a moderately loamy, a sandy-loamy and a sandy soil and the thickness of the film of water around the soil particles was described approximately by a curve. At a particular thickness of the film, the coefficient of thermal conductivity and the coefficient of temperature conductivity were related almost linearly to volume weight. A simplified relation between the coefficient of conductivity and thickness of the film is derived.

7. Grable, A. R., and E. G. Siemer. 1968. Effects of bulk density, aggregate size and soil water suction on oxygen diffusion, redox potential and elongation of corn roots. *SSSAP* 32:180-186.

Corn seeds (*Zea mays* L. 'WF9 x 38-11') were planted in plastic cylinders of soil at different bulk densities (0.93, 1.03, 1.13, and 1.23 g cm⁻³) or aggregates of different sizes (<0.5, 0.5-1, 1-2, 2-3, and 3-6 mm). Soil water suction during germination was maintained at constant levels (0, 3, 18, 48, and 68 cm of water).

Germination began and root elongation rates sharply increased from zero to a maximum in each soil sample as soil water suction increased above the air-entry value or bubbling pressure. Over these ranges of suction, diffusion of O₂ controlled rate of root elongation. Diffusion of O₂ in soil was determined primarily by air porosity. Bulk density and aggregate size per se had little effect on diffusion or concentration of O₂ at seed depth. Bulk density and aggregate size greatly influenced soil water desorption, the depth of soil to which O₂ could diffuse, and the thickness of water films at equal air porosities. Between approximately 0 and 20 cm of suction, thickness of water films limited root elongation. Measured O₂ concentrations in soil agreed with concentrations predicted by diffusion theory. Redox potentials were correlated with O₂ levels. Root elongation rates decreased in some samples at the highest suctions used, probably because of increased soil strength.

8. Gupalo, A. I. 1959. Thermal properties of soil in relation to its moisture content and compaction. *Pochvovedenie* 4:40-45.

Temperature conductivity increases with increasing soil moisture until the latter retards plant growth. The rate of equalization of temperature is thus greater in moister soils and warming up occurs to a greater depth. The coefficients of temperature conductivity and heat conductivity increase with increasing compaction, the nature of the increase depending on soil moisture. The mechanism of heat transfer in moist soil depends on the mechanism of moisture transfer at different moisture contents.

9. Hadas, A. 1977. Heat transfer in dry aggregated soil. 1. Heat conduction. *SSSAJ* 41(6):1055-1059.

The thermal conductivities of aggregated soils were determined experimentally in the laboratory. The thermal conductivity decreases as the size of the aggregates and the total and interaggregate porosities increase. Good agreement between the measured and predicted thermal conductivity values, by two methods, was found only for small aggregates, but failed increasingly as the aggregate size increased. The discrepancies between measured and predicted values are discussed.

10. Jones, H. E., and H. Kohnke. 1952. The influence of soil moisture tension on vapor movement of soil water. SSSAP 16:245-248.

The pF at which vapor movement was initiated and at which maximum movement occurred increased with a decrease in particle size.

The volume of unsaturated soil pores, not their size, seems to govern the soil moisture content at which vapor movement begins.

Maximum vapor movement occurred at moisture tensions slightly below the wilting point.

Temperature induced vapor pressure gradients seem to be mainly responsible for water vapor transfer in soils.

13. Nakshabandi, G. A., and H. Kohnke. 1965. Thermal conductivity and diffusivity of soils as related to moisture tension and other physical properties. Agri. Meteor. 2:271-279.

The same increments of moisture have an entirely different effect upon the increase in thermal conductivity of different mineral soils, if the comparison is based on the quantity of water added. However, the relationships between thermal conductivity and soil moisture tension are very similar for these same soils.

Thermal conductivity of dry soils increases with the bulk density. It also increases with the grain size. These increases are small, however, compared to those resulting from moisture changes.

The relationship between soil moisture tension and thermal diffusivity is similar for the three soils investigated. It appears that in the moist range for the same tension the diffusivity is the greater, the coarser is the soil texture.

14. Skaggs, R. W., and E. M. Smith. 1968. Apparent thermal conductivity of soil as related to soil porosity. ASAE Trans. 11:504-507.

This paper presents an analysis of the factors affecting the thermal conductivity of soil. The effect of soil porosity is studied in detail and a theoretical method for the evaluation of thermal conductivity in terms of soil porosity is discussed. Experimental evidence is presented verifying the method for Maury silt loam soil.

15. Tripath, R. P., and B. P. Gildyal. 1974. Thermal properties of fine texture mollisols under varying moisture conditions. J. Indian Soc. Soil Sci. 22(2):103-108.

The specific heat of three soils, Phoolbagh clay loam, Beni silty clay loam, and Haldi loam was calculated using

the radiation calorimeter and found to be 0.231, 0.224 and 0.207 Cal g⁻¹ C⁻¹, respectively. Thermal conductivity in all the soils increased rapidly with per cent saturation up to about 50% of pure saturation. The maximum value of the thermal diffusivity occurred around 50% pore saturation.

IV. MEASUREMENTS OF HEAT AND WATER MOVEMENT

A. Hot water storage in porous media.

1. Injection experiments.

- a. Cormary, Y., P. Iris, J. P. Marie, G. deMarisly, H. Michel, and M. F. Zaquine. 1978. Heat storage in a phreatic aquifer: Campuget experiment (Gard, France). Proc. TES in Aquifers Workshop, Berkeley, California. pp. 88-93.

A small quantity of hot water (1400 m^3 at 40 C) was injected during 20 days, and recuperated after 4 months in a confined aquifer, 3 m thick. The aim of this experiment was: 1) to measure in situ thermal parameters of the aquifer, and 2) to experiment with heat storage. Only 30% of the heat was recovered, at an average temperature of 15 C , for a withdrawal of 3000 m^3 , and a normal temperature of 14 C in the aquifer. This was not surprising, considering the size of the experiment.

- b. Mathey, B. 1977. Development and resorption of a thermal disturbance in a phreatic aquifer with natural convection. J. Hydrol. 34:315-333.

Development and resorption of a thermal disturbance with natural convection in a phreatic aquifer during a 5 month injection-pumping cycle was studied. Evaluation of thermal balance revealed insignificant losses but an important "dilution" of the disturbance by thermal conduction.

- c. Molz, F. J., A. D. Parr, P. F. Andersen, and V. C. Lucido. 1979. Thermal energy storage in a confined aquifer: experimental results. Water Resour. Res. 15:1509-1514.

To aid in testing the idea of storing thermal energy in aquifers, an experiment was performed by Auburn University in which $54,784 \text{ m}^3$ of water was pumped from a shallow supply aquifer, heated to an average temperature of 55 C , and injected into a deeper confined aquifer where the ambient temperature was 20 C . After a storage period of 51 days, $55,345 \text{ m}^3$ of water were produced from the confined aquifer. Throughout the experiment, which lasted approximately 6 months, groundwater temperatures were recorded at six depths in each of 10 observation wells, and hydraulic heads were recorded in five observation wells. In order to prevent errors due to thermal convection, most of the observation wells recording temperature

had to be backfilled with sand. During the 41 day production period, the temperature of the produced water varied from 55 to 33 C, and 65% of the injected thermal energy was recovered. At no time was an appreciable amount of free thermal convection observed in the storage formation. The dominant heat dissipation mechanisms appeared to be hydrodynamic thermal dispersion and possible mixing of cold and hot water induced by clogging and unclogging of the injection-production well. On the basis of laboratory and field studies, it was concluded that clogging of the injection well, which constituted the major technical problem during the experiment, was caused by the freshwater-sensitive nature of the storage aquifer. Due to the relatively low concentration of cations in the supply water, clay particles would swell, disperse, and migrate until they became trapped in the relatively small pores connecting the larger pores. Surging the pump and back washing the injection well would dislodge the clogging particles and temporarily improve the storage formation permeability. The phenomenon seems largely independent of temperature because it was reproduced in the laboratory with unheated water. It may, however, depend on pore velocity. Future research should be directed toward procedures for selecting storage aquifers that will have minimal susceptibility to clogging and other geochemical problems. Procedures for overcoming such difficulties are needed also because clogging and related phenomena will be more the rule than the exception. Designing an aquifer thermal storage system for maximum energy recovery would involve selecting an appropriate aquifer, analyzing the effects of hydrodynamic thermal dispersion and thermal convection if it is predicted to occur, anticipating geochemical problems, designing the optimum supply-injection-production well configuration and injecting a sufficiently large volume of heated water to realize economies of scale related to increasing volume surface area ratio.

- d. Molz, F. J., J. C. Warman, and T. E. Jones. 1978. Aquifer storage of heated water. 1. A field experiment. *Ground Water* 16(4):234-241.

Electrical power plant and solar heating systems have been proposed wherein confined ground-water aquifers are used as temporary storage reservoirs for thermal energy in the form of moderate to high temperature water (60 C-204 C). The Water Resources Research Institute of Auburn University has performed an aquifer storage experiment involving warm water (36 C). The objectives of the experimental program were to begin actual testing of the concept of heat storage in

aquifers and to provide data for calibration of mathematical models describing the simultaneous transport of water and heat. Phase I consisted of drilling an exploratory well at the selected field site near Mobile, Alabama. Phase II involved construction of the central injection well, three observation wells, and performance of preliminary pumping tests. Phase III was devoted to construction of the remainder of the observation well field, performance of final pumping tests, and measurement of aquifer thermal properties; while Phase IV was devoted to a cycle of warm-water injection, storage, and recovery. It was concluded that heat storage aquifers must have low natural pore velocities, and care must be taken not to clog the injection well with solids or precipitated chemicals. Swelling of clays in the storage formation must be minimized, and hydraulic pressures capable of breaching the confining layers must be avoided. Mechanical and chemical clogging problems may be minimized by using formation water as influent to the heating system. For a 36 day storage of 2 million gallons, the calculated energy recovery factor of 0.69 was considered promising. Future research should be directed toward experimental studies involving larger volumes of water and high-injection temperatures. Study should be directed also to the geochemistry problem and the effect of high temperatures on the mechanical and hydraulic properties of clay confining layers.

- e. Werner, D., and W. Kley. 1977. Problems of heat storage in aquifers. J. Hydrol. 34:35-43.

A hydrothermic field experiment near Krefeld (Federal Republic of Germany) has been used to study the propagation of heat in groundwater. For two months water at a temperature of 15 C was fed through a well pipe to the water-bearing strata, the distribution of temperature in the surrounding space was measured and its variation with time was observed.

A theoretical model has been developed to simulate the field experiment. This consists of a time-space differential method with the help of which good conformity between theory and experiment is obtained.

The aim of this study is to ascertain the physical bases for the storage of heat in groundwater. This techno-economic and environmental problem is posed because of the enormous quantities of cooling water necessary in the operation of atomic power stations. A functional theoretical model can help to decide how far the idea of storing heat in an aquifer is technically feasible.

2. Other related studies.

- a. Akbarzadeh, A., and G. Ahmadi. 1979. Under ground thermal storage in the operation of solar ponds. *Energy (Oxford)* 4(6):1119-1125.

The thermal interaction between a large solar pond and the surrounding ground is considered. For a given sinusoidal variation of the temperature at the bottom of the pond, the time-dependent temperature profiles in the ground are calculated and the corresponding heat fluxes to or from the ground as functions of time are obtained. The temperature variations in the ground for several years are plotted and the heat transfer between the solar pond and the ground thermal storage is discussed. The efficiency of the heat recovery is studied and its significance is pointed out.

- b. Greenland, D. E. 1969. Soil heat flow investigations at Cass: South Island high county. *New Zeal. J. Agr. Res.* 12:352-366.

Soil heat flow was measured in a high country valley at different seasons when rain followed a dry spell. The thermal conductivity and heat capacity of the soil were computed. Periodic heating at the surface was demonstrated, but a model describing heating through all depths was not satisfactory. The heat flow for the periods investigated was related to the other components of the heat balance at the surface.

- c. Kemper, W. D., W. R. Walker, and J. Sabey. Trans-seasonal storage of energy in moist soils.

These soil scientists stress minimization of surface to volume and self mulching of a storage system where the water is trapped in a plastic barrier.

Their data for heat and water flux from bottom heated systems suggests that initial water content and soil texture only influence start up time for system and not ultimate efficiency.

- d. Lilleleht, L. U., L. M. Fafarman, and J. T. Beard. 1978. Earth-conducted heat losses from thermal storage systems. *Thermal Storage and Heat Transfer in Solar Energy Systems Symp.*, ASME, New York. pp. 41-44.

In an effort to reduce the storage costs of solar thermal energy, it has been proposed to use the ground to provide some or all of the insulation of the storage medium and some of the thermal capacity as well. Analyses of heat losses from such systems are often

quite difficult, especially for the more complicated geometries and due to their cyclical nature of operation. This paper presents a new method of estimating the steady-state heat conduction losses from storage systems in the shape of segments of spheres. The use of orthogonal toroidal coordinates is illustrated for predicting these losses from constant temperature objects imbedded in earth with another constant surface temperature.

- e. Shelton, J. 1975. Underground storage of heat in solar heating systems. *Solar Energy* 17:137-143.

The thermal interaction between underground heat storage and the surrounding ground is studied. The storage medium may be water in a tank, rocks, the ground itself, etc. For the case of hemispherical geometry, analytic steady-state solutions and numerical solutions to select time-dependent situations are presented and discussed.

The quasi-steady-state average daily net heat loss into the surrounding ground from uninsulated underground heat-storage facilities is estimated to be only a few per cent of the heat storage capacity for typical home-size units in low-conductivity ground in the absence of ground-water flow. The economics of added insulation under these circumstances appears to be critically site and system specific.

The time required for the average heat-loss rate of a new system to approach the steady-state value is on the order of a year.

The ground surrounding a typical heat-storage facility contributes very little to the system's storage capacity either on a daily or a seasonal time scale.

- f. Yuan, S. W., and L. S. Galower. 1975. A central solar energy utilization system. *Int. Sem. on Future Energy Production: Heat and Mass Transfer Problems*. Dubrovnik, Yugoslavia.

This research is concerned with the long-duration earth storage of solar heat for space and hot water heating. Preliminary analysis indicates that the properties of the ground are excellent for this purpose and will achieve a heat capacity adequate to store summer heat for winter use without excessive thermal loss due to ground conductivity. Conservative calculations were also made that sufficient heat quantity can be transferred from the earth storage of an acre of land to heating requirements of at least twelve average-size houses in most areas of the United States.

B. Irrigation with warm water in moist soils.

1. deWalle, D. R., D. D. Fritton, L. T. Kardos, and D. P. Knievel. 1978. Soil warming for waste heat disposal crop growth enhancement and waste water renovation. Inst. for Research on Land and Water Resources. Pennsylvania State Univ., University Park.

To examine the feasibility of utilization and dissipation of reject heat from power production, the effects of year-round soil-warming on subsurface and surface heat exchange, crop growth and development and renovation of municipal waste water were determined in central Pennsylvania. The soil in a 0.09 ha test plot was artificially heated by circulating water heated to 38 C through a buried, parallel pipe network. Both the heated plot and an adjacent unheated control area were irrigated weekly with about 1.5 cm of municipal sewage effluent. Measurements of soil heat flow, pipe heat loss, soil temperatures, pipe temperatures, and soil thermal conductivity were used to test theoretical steady-state models describing heat loss from a buried pipe network. The fraction of pipe heat loss dissipated by evapotranspiration was determined through the use of weighing microlysimeters. A horizontal atmospheric advection model was used to scale the small plot heat loss data to a full-sized area. Crop phases of the research focused on the effect of the warm soil on winter survival, maturation, development, yield, quality and pest relationships for alfalfa, orchardgrass, tall fescue, winter wheat, winter barley, snapbeans and sudan-grass. Crop research also included a greenhouse study of the effects of soil temperature and compaction of root growth of tall fescue. The influence of the warm soil on the chemical composition of the applied sewage effluent was determined from samples periodically collected with suction lysimeters and analyzed for pH, $\text{NO}_3\text{-N}$, organic-N, Cl^- , total soluble salts and soluble orthophosphate.

2. Elwell, D. L., W. L. Roller, and A. Ahmed. 1978. Waste heat for root-zone heating: A physical study of heat and moisture transfer. Second Waste Heat Management and Utilization Conf. Proc., Univ. of Miami, Mech. Eng. Dep. Coral Gables, Fla. p. 28-38.

A study of the use of power plant waste heat for root-zone heating in greenhouses is being undertaken. As the first part of this study, the physical properties of a system that supplies heating water through buried pipes and irrigation water through a constant level water table have been characterized. Three different soils (sand, peat-vermiculite and silt-loam) have been investigated at heating water temperatures of 25, 30, 35 and 40 C. The resultant heat and moisture transfer figures are internally

consistent and in good agreement with comparable results of other investigators, where available. The water table irrigation system is capable of maintaining suitable soil moisture content for plant growth up to the highest temperatures investigated, and the soil heating system can supply up to 30% of the total, seasonal heating load of a greenhouse while producing root-zone temperatures up to 30 C. These results will now serve as a base for continuing study with lettuce growing on the system.

3. Rykbost, K. A., and L. Boersma. 1974. Soil and air temperature changes induced by subsurface line heat sources. Ag. Expt. Stn. Spec. Rep. 402. Oregon State Univ. Corvallis. 105 p.

In regions where soil temperatures limit plant growth, artificial soil warming may be an economically feasible practice. This hypothesis was evaluated in a soil warming experiment near Corvallis, Oregon. These experiments were prompted by the observation that multiple use of waste heat discharge in the condenser cooling water of thermal power plants may well become an important consideration in the development and siting of these plants. The thermal discharge might be used to achieve increased soil temperatures by circulating warm water through a subsurface pipe network. The effect of increased soil temperatures on crop growth was described in a separate report (Rykbost et al., 1974). This report describes energy balances and water regimes of soils heated above natural temperatures.

4. Sepaskhah, A. R., L. Boersma, L. R. Davis, and D. L. Slegel, 1973. Experimental analysis of a subsurface soil warming and irrigation system utilizing waste heat. ASME Pap. No. 73-AW/HT-11. 12 p.

Soil warming by condenser cooling water as one part of an integrated system for waste heat utilization is considered. Laboratory and field experiments are analyzed to determine effective thermal conductivity, moisture content and usage, and energy dissipation. The effective thermal conductivity of soils is presented as a function of mineralogical composition. Both laboratory and field experiments show drying of the soil near the heat source to occur, decreasing the thermal conductivity of the soil. However, energy loss and temperature distribution is greatly enhanced by supplying water near the heat source by a subsurface irrigation system. Calculations indicate the land surface area required to dissipate the waste heat from a 1000-MWe power plant using soil warming to be very large.

5. Skaggs, R. W., D. C. Sanders, and C. R. Willey. 1974. Evaluation of the use of waste heat for soil warming in the Southeast. Water Resour. Res. Inst., Univ. of North Carolina, Rep. 103. 100 p.

Soil temperatures and heat flux measurement were analyzed to determine the effect of soil warming in combination with surface and subsurface irrigation on the soil temperature regime and on the amount of heat that can be transferred to the soil system. The following crops were grown on the test plots: cabbage, cucumber, snap bean, strawberry, sweet corn and sweet potato. The influence of soil warming and other environmental factors on the growth, development and yield of these crops was evaluated. The mathematical model of Kendrick and Havens was used to project the heat flux measurements to prototype scale systems. It was determined that a soil warming system of the type studied would require approximately 16,000 acres to handle the waste heat from a 1000 megawatt plant during the summer in North Carolina. Approximately 3800 acres could be warmed during the coldest winter months.

C. Evaporation from shallow water tables.

1. Gardner, W. R. 1958. Some steady state solutions of the unsaturated moisture flow equation with application to evaporation from a water table. Soil Sci. 85:228-232.

A transformation is given which makes possible the exact solution of some steady-state unsaturated-flow problems and approximate solutions of some transient problems. Solutions of the steady-state problem in one dimension are given for several different relations between capillary conductivity and soil suction.

Steady-state evaporation from a soil in which there is a water table is examined. The maximum evaporation rate is shown to be related to the capillary conductivity and depth to water table in a simple fashion. It is concluded that movement of water in the vapor phase is relatively unimportant in this connection. The influence of a surface mulch is also considered.

2. Hadas, A., and D. Hillel. 1968. An experimental study of evaporation from uniform soil columns in the presence of a water table. Int. Congr. Soc. Soil Sci. Tans. 9th. (Adelaid, Australia) 1:67-74.

The paper presents the results of laboratory experiments on evaporation from uniform soil profiles in the presence of a water-table.

Theoretical solutions describing steady-state evaporation predict that when external evaporativity is increased the soil eventually attains a maximal evaporation rate which depends upon water-table depth and profile transmissivity and which remains constant regardless of how high external evaporativity may become. Our experiments showed that instead of leveling off at the maximal evaporation rate, as evaporativity was further increased the

actual evaporation rates began to decrease. This may result from gradual salt accumulation, structural effects or changes in boundary conditions (i.e., the formation of a two-layer condition in which a dry surface layer eventually acts as a diffusion barrier). A comparison of the measured evaporation rates and those predicted theoretically showed that both Gardner's and Anat's theories overestimated the maximal rate, especially for shallow water-table conditions. Modification of Gardner's theory to correct for excessively high conductivity values near saturation improved the estimate somewhat. Cyclic changes in evaporativity or in water-table depth apparently induced hysteretic effects which decreased the evaporation rate.

3. Hadas, A., and D. Hillel. 1972. Steady-state evaporation through non-homogenous soils from a shallow water table. *Soil Sci.* 113(2):65-73.

In the work presented here an attempt was made to study the effects of soil profiles layering on evaporation from a shallow water table, under isothermal conditions, and to test the assumptions and theory developed to date.

4. Hellwig, D. H. R. 1978. Evaporation of water from sand. 6. The influence of the depth of the water table on diurnal variations. *J. Hydrol.* 39(1/2):129-138.

Evaporation from sand tanks in an arid climate was estimated by measuring the amount of water needed to maintain the water table at a prescribed fixed level. In earlier experiments of this kind two distinct peaks were found, one in the afternoon and the other at sunrise.

To obtain more reliable information on the phenomenon of the peak at sunrise, hourly data from an evaporation experiment which lasted 847 days were evaluated. The results of 518 days were used for a diurnal evaluation. It was confirmed that two peaks occur during a 24 hr period. One peak is related to radiation and the deeper the water table is kept below the sand surface the later it occurs. This was the absolute diurnal peak for those tanks in which the sand surface was never dry. It was a relative diurnal peak for those tanks in which the sand surface was normally dry.

The other peak appears at sunrise and is related to the air temperature. This peak was independent of the depth of the water table up to a depth of 400 mm below the sand surface. Down to a water-table depth of 600 mm below the sand surface the time lag at which this peak occurred relative to the lowest air temperature increased for up to 2 h. This peak was a relative diurnal peak for those tanks in which the sand surface was never dry. It was the absolute diurnal peak for those tanks in which the sand surface was normally dry.

It seems as if the evaporation peak at sunrise depends on the presence of a condensation surface.

5. Saxena, G. S., G. S. Taylor, and R. E. Franklin. 1971. Effect of environmental factors on evaporation rates from soils in the presence of a water table. *J. Indian Soc. Soil Sci.* (19(1):23-29.

Laboratory studies were conducted to assess the effect of external environmental factors on evaporation rates from soils in the presence of a water table. Columns of fine quartz sand and Miami silt loam were used, and water table was maintained at desired depths. Radiant energy was supplied by an incandescent lamp suspended above the surface of soil column, and air turbulence was created by an electric fan. Humidity and temperature were not controlled. An integrated effect of all ambient factors was measured in terms of evaporation rate from a free water surface, which was referred to as potential evaporation rate. It was observed that evaporation rate was mainly governed by external conditions when the water table was within 60 to 90 cm of the surface. The nature of the soil played no significant role. For deeper water tables, the capillary conductivity of the soil became the main controlling factor. Lowering the water table appreciably reduced the evaporation rates, but even from very deep water tables, the evaporative losses continued at the rate of about 0.05 cm/day. This can lead to salinization of soils in a period of time, depending upon the concentration of salts in the ground water. Additional studies were made on measurement of capillary conductivity. Gamma-ray attenuation technique was employed for non-destructive and frequent sampling of moisture content in soil profile above the water table.

6. Tripathi, R. P., and B. P. Gildyal. 1978. Transport and evaporation of water from soil columns under conditions of high water-table. *J. Indian Soc. Soil Sci.* 26(4):313-319.

Soil-water evaporation under 10, 30, 50, 70, and 85 cm water-table depths was studied in Haldi loam soil of Pantnagar. Evaporation potential of 2.3 cm/day was applied by passing heated air of constant temperature and speed over the soil surface. Under steady state temperature conditions the maximum temperature zone occurred at 5 cm depth causing a temperature gradient towards the surface and also in a downwards direction. These temperature fields resulted in thermal moisture transfer both in vapor and liquid phases in the direction of a decreasing temperature gradient. The thermally induced vapor flow of moisture was 0.02 cm/day whereas liquid phase transport ranged from 0.12 to 0.57 cm/day downwards, and from 0.047 to 1.120 cm/day towards the surface, from the maximum temperature zone

under the different water-table depths. Thermal vapor flow increased when the water content in the soil was decreased.

D. Sensing of subsurface water from surface temperature.

1. Cartwright, K. 1974. Tracing shallow groundwater systems by soil temperatures. *Water Resour. Res.* 10(4):847-855.

Circulating water is known to affect the temperature of the rock through which it flows. Analysis of existing theoretical equations suggests that shallow groundwater flow will also affect the surface soil temperature. Soil temperatures therefore might be used to delineate small, shallow groundwater flow systems: both recharge and discharge zones might be distinguished. Temperature data taken in the field during summer and winter support the theoretical conclusions and agree with the groundwater flow patterns inferred from hydrologic data. In general, winter soil temperature profiles have less data scatter than profiles of summer temperatures. The greater the rate of groundwater flow, the greater the temperature difference between recharge and discharge zones. Temperature variations ranged from 0.75 C in clayey glacial till to 5 C in sand.

2. Huntley, D. 1978. On the detection of shallow aquifers using thermal infrared imagery. *Water Resour. Res.* 14(6):1075-1083.

Patterns appearing on a thermal infrared image are primarily a function of the temperature of the earth's surface. Diurnal surface temperatures are affected by the thermal inertia and solar reflectance of the surface by meteorologic variables, by evaporation rates, and by subsurface temperature effects such as the effect due to varying groundwater depth. Variations in surface material result in variations in both thermal inertia and solar reflectance, causing a change in both the magnitude of temperatures through the diurnal cycle and the amplitude of diurnal temperature variation. Variations in groundwater depth produces an effect on surface temperature that is discriminable from variations in solar reflectance and thermal inertia but is not discriminable from variations in evaporative cooling rates.

3. Meyer, C. R., J. A. Tunheim, and D. G. Moore. 1975. Heat flow-temperature model for remotely mapping near-surface water tables by thermography. *Proc. South Dakota Acad. Sci.* 54;23-32.

A theoretical study was made relating to the detection and mapping of near-surface ground water by the use of

remotely sensed thermal emittance. Differences in surface temperatures expected from the presence of a near-surface water table due to the large heat capacity of water were calculated. These thermal anomalies were evaluated using the model both during their development and during several diurnal cycles. The model considers heat transfer in two identical soil layers with the lower boundary temperatures differing by 1 C to 3 C as reported in the literature. Calculations using this model give the following results: a) The surface thermal anomaly is from 20% to 40% of the subsurface anomaly present at a 50 cm depth. b) Variations in the surface thermal anomaly over a 24 hr period are about 5% to 10% of the anomaly magnitude. c) A greater difference between the effective atmospheric temperature and the surface soil temperature results in a slightly larger thermal anomaly. d) If the surface thermal anomaly is diminished due to environmental conditions, it will redevelop in several hours after normal conditions are restored.

4. Thompson, G. E. K. 1976. Fluctuations in near-surface ground temperature, Taupo, New Zealand. *New Zealand J. of Geol. and Geophys.* 19(3):352-364.

Near-surface ground temperatures fluctuated rapidly, over 15 C in one day, at a site in Taupo Borough where heat transfer is by convection from the underlying hydro-thermal system. There was a good inverse correlation between near-surface ground temperature and atmospheric pressure, suggesting that a decrease in atmospheric pressure promotes vapor emission at the hot ground-water surface, and increases the rate of convective heat transfer, and that the near-surface ground temperature. Conversely, an increase in atmospheric pressure inhibits vapor emission and results in a lowering of near-surface ground temperature.

V. COMPUTER MODELS

- A. Aston, A. R., and A. M. Gill. 1976. Coupled soil moisture, heat and water vapour transfers under simulated fire conditions. *Aust. J. of Soil Res.* 14:55-66.

Good job of computer modeling the system with just a few exceptions: They never measure thermal conductivity or separate it into conduction and vapor flow. Also, they solve vapor flow, heat flow, and liquid flow separately and connect them with sources and sinks to account for evaporation and condensation. This is important for systems which operate over a large range of temperatures.

- B. Baca, R. G., I. P. King, and W. R. Norton. 1978. Finite element models for simultaneous heat and moisture transport in unsaturated soils. *Proc. 2nd Int. Conf. on Finite Elements in Water Resources.* Pentech Press, London. p. 19-21.

A finite element model for use in the analysis of transport processes in arid site vadose zones is described and demonstrated. The numerical model is designed to provide detailed simulations of the interactions between climatic conditions and natural moisture movement in a dry soil system. The theoretical framework of the models considers processes of coupled heat and moisture transport, soil-atmosphere interactions and spatially varying soil properties. The mathematical formulation of the non-isothermal model consists of the coupled form of the pressure based flow equation and the conduction equation. The flow equation models moisture flow as a function of the capillary, gravity and thermal driving forces. The heat equation considers conduction and surface fluxes arising from atmospheric heating and cooling.

- C. Eckert, E. R. G., and E. Pfender. Heat and mass transfer in porous media with phase change. Int. Heat Transfer Conf. Hemisphere, Washington, D. C., 6th. 6:1-12.

The paper aims at a clear presentation of the important features characterizing combined heat and mass transfer in a porous medium and specifically in soils. A description of the soil characteristics leads to an idealized model and the assumption of capillary dominated conditions is used to illustrate the way in which soils hold liquid water for equilibrium in the absence of gravity or with gravity. The basic equations describing the transport of liquid water under isothermal conditions and the transport of vapor and energy in the presence of temperature gradients are developed.

- D. Erh, K. T., D. R. Nielsen, and J. W. Biggar. 1971. Two dimensional heat transfer in porous media with steady-state water flow. SSSAP 35:209-214.

Water entering columns of a sandy loam, clay loam and a clay was kept at a temperature below or above the ambient temperature for several values of soil water velocity. Values of thermal conductivity generally depended on flow velocity at values $<0.3 \text{ cm min}^{-1}$. The axial thermal conductivity was 1.1 times greater than the radial thermal conductivity for all soils and flow velocities.

- E. Goodrich, L. E. 1978. Efficient numerical technique for one-dimensional thermal problems with phase change. Int. J. Heat Mass Transfer 21(5):615-621.

A new numerical scheme for one-dimensional heat flow problems with phase change is presented. The technique, which continuously monitors the progression of the phase interface, is unusual for the high accuracy achieved without sacrifice to computing efficiency. These data are needed by designers of roadways in cold climates for calculating the depth of frost or thaw penetration.

- F. Guymon, G. L. and R. L. Berg. 1977. Galerkin finite element analog of coupled moisture and heat transport in algiol soils. In W. E. Gray, G. F. Pinder, and C. A. Brebbia (eds.), Finite element in water resources, Proc. Int. Conf. on Finite Elements in Water Resources, 1st. Pentech Press, London. pp. 3.107-3.114.

This paper briefly summarizes recent research of the authors, with more detailed information and applications presented elsewhere. The specific application currently under study concerns modeling of the frost heave and thaw consolidation process. Numerous other applications are also possible; e.g., modeling redistribution of moisture in the freezing soil profile in agricultural regions and thermal stability of frozen embankments.

- G. Slegel, D., L. Davis, and L. Boersma. 1974. Simultaneous heat and mass transfer in soils with subsurface heated porous pipes. Seminar on Heat and Mass Transfer in the Environ. of Veg: Heat and Mass Transfer in the Biosphere. Pt. 1. Transfer Processes in the Plant Environ. (Dubrovnik, Yugoslavia), Pap. 6. Scripta, Washington, D. C. pp. 87-96.

The development of transient equations describing the heat and mass transfer in soils is presented. Both liquid and vapor phases are considered. The results of solving these equations numerically on a digital computer for temperature and moisture content are presented. Boundary conditions were chosen to represent a subsurface soil warming and irrigation system consisting of a series of submerged parallel porous, heated pipes. The spacing and depth of the pipes were varied to find their effect on temperature and moisture content distributions.

- H. Zienkiewicz, O. C., C. J. Parekh, and A. J. Wills. 1973. Application of finite elements to heat conduction problems involving latent heat. Rock Mech., Felsmech. Mec. Roches 5(2):65-76.

A finite element method is developed for the solution of heat conduction problems which involve latent heat. The method is superior to other well-known approaches to these problems in that it allows a wider range of material properties and boundary conditions to be dealt with, such as are encountered in complex engineering operations like ground freezing. The method is applied to simple triangular finite elements in this paper, although it could be extended to other types of element including three-dimensional ones. Several example problems are discussed and illustrated, and comparisons are made with other approaches where these can also be used.

VI. WATER AND HEAT TRANSPORT UNDER FREEZING TEMPERATURE CONDITIONS

- A. Fuchs, M., G. S. Campbell, and R. I. Papendick. 1978. Analysis of sensible and latent heat flow in a partially frozen unsaturated soil. SSSAJ 42(3):379-385.

A heat flow theory has been developed which can be used to predict freezing and thawing cycles of a temperate soil. The latent heat of fusion is incorporated into the classical heat flow equation with the heat capacity and the thermal conductivity of the soil modified to include terms which account for the phase transformation. Computations for a Palouse silt loam soil show that the ice-liquid phase transformation occurs over a defined temperature range where the apparent heat capacity and the apparent thermal conductivity may take on values which are several orders of magnitude larger than those either in the unfrozen or in the near completely frozen soil.

- B. Groenevelt, P. H., and B. D. Kay. 1974. On the interaction of water and heat transport in frozen and unfrozen soils: II. The liquid phase. SSSAP 38(3):400-404.

The phenomena of thermoosmosis and thermofiltration in frozen and unfrozen soils are considered from a phenomenological point of view. It is shown by model calculations that the transport coefficients of these two phenomena, as far as the liquid phase is concerned, are equal. This coupling is related to the heat of wetting and heat of fusion.

- C. Harlan, R. L. 1973. Analysis of coupled heat-fluid transport in partially frozen soil. Water Resour. Res. 9(5):1314-1323.

A Darcian approach was applied to the analysis of coupled heat-fluid transport in porous media with freezing and thawing. With the aid of a numerical model, freezing-affected water redistribution to frozen soil was examined and the effects of soil type and initial conditions on the response of a hypothetical soil column were studied. The rate of upward movement of water to a freezing zone decreased from coarse to fine textured soil and with increasing depth to the water table.

- D. Higashi, A. 1953. On the thermal conductivity of soil, with special reference to that of frozen soil. Trans. AGU 34:737-748.

Using the Angstrom principle, a new apparatus for measuring the thermal diffusivity of soil was designed and the thermal diffusivity of frozen and of wet soil was measured. Thermal conductivity was computed from the diffusivity and the volume specific heat of the soil. Moisture relationships of the thermal diffusivity and the conductivity of frozen soils are represented by an empirical formula of exponential form. Samples of various modes of segregation of ice were artificially prepared and their thermal diffusivity measured.

- E. Kay, B. D., and P. H. Groenevelt. 1971. On the interactions of water and heat transport in frozen and unfrozen soils. 1. Basic theory: the vapor phase. SSSAP 38(3):395-400.

A theory is presented for water transport due to a difference in temperature and heat transport due to a difference in water pressure as they occur in frozen and unfrozen soils. It is shown by model considerations that the transport coefficients of these two phenomena, as far as the vapor phase is concerned, are equal. This coupling is related to the heat of vaporization, the heat of fusion and the heat of wetting of the soil water.

- F. Kudyavtev, V. A., E. D. Yershov, V. G. Cheverev, Y. P. Lebedenko, L. V. Shevchenko. 1977. Moisture transport and formation of segregation ice in freezing and thawing particulate soils. Heat Transfer (Soviet Res.) 9(1):38-45.

The results of a laboratory study of moisture transport in freezing and thawing soils are presented. A new mechanism of formation and growth of cryogenic structures, with allowance for the thermophysical and physicomachanical heat and mass transfer conditions, is suggested.

- G. Penner, E. 1970. Thermal conductivity of frozen soils. Can. J. Earth Sci. 7:982-987.

The thermal conductivity of 2 frozen soils, a clay and a silty clay, at temperatures between 0 and -22 C was measured by a thermal probe and transient heat flow technique. The method compared well with estimates calculated by the deVries method, except between temperatures of 0 and -2 C. Values showed a trend of increased thermal conductivity as the temperature was lowered as ice has a higher thermal conductivity than water.

- H. Taylor, G. S., and J. N. Luthin. 1978. Model for coupled heat and moisture transfer during soil freezing. Can. Geotech. J. 15(4):548-555.

The paper presents simulation analyses of heat and water flow during soil freezing. The analyses utilize a modified form of a model presented by R. L. Harlan. The model is tested by using experimental data reported by C. Dirksen and R. D. Miller, and Y. W. Jame and D. I. Norum. The basic elements of the model are the heat and water flow equations. An implicit finite-difference scheme is used to solve these two equations for the boundary conditions used in the experiments. The agreement between simulated and experimental results is illustrated for temperature, water, and ice contents.

VII. LIQUID CONVECTION IN SATURATED POROUS MEDIA

- A. Fel'dman, G. M. 1975. Calculation of free thermal convection in coarse-grained soils. Heat Transfer (Sov. Res.) 7(6):93-108.

The free thermal convection in coarse-skeleton soils was calculated for the conditions of the internal and the external problem. In the first case, the unsteady-state problem was solved, whereas the steady-state problem was solved in the second case. The formulas derived allow determining the effective thermal conductivity accounting for water circulation.

- B. Katto, Y., and T. Masuoka. 1967. Criterion for the onset of convective flow in a fluid in a porous medium. Int. J. Heat Mass Transfer 10:297-309.

The purpose of the paper is to make a careful study of the breakdown of stability of a layer of fluid subject to a vertical temperature gradient in a porous medium, and to give a conclusive criterion for the onset of convection currents.

Through theoretical examinations, it is pointed out that the phenomenon is affected with a specially defined thermal diffusivity, and that there are possibilities for the ordinary theory based upon Darcy's law to be applicable even when the permeability of the porous medium becomes considerably high. Careful experiments also are carried out in which the difficulties to generate the convective flow in the porous medium under reasonable temperature gradient are overcome by the use of the porous medium of comparatively high permeability as well as by the use of a compressible gas as the fluid.

Satisfying agreement of the experimental results with the theory is obtained to provide conclusions that the criterion for the onset of convective flow is certainly given by the equation: $Ra k/l^2 = 4\pi^2$ where Ra is Rayleigh number, k the permeability, and l the vertical thickness of the porous medium, but that the thermal diffusivity included in Ra must be defined as the thermal conductivity of porous medium divided by the specific heat capacity of fluid.

- C. Minkowycz, W. J., and P. Cheng. 1976. Free convection of water in a porous medium. Int. J. Heat and Mass Transfer 19:805-813.

An analysis is made for free convective flow about a vertical cylinder embedded in a saturated porous medium, where surface temperature of the cylinder varies as x^λ , a power function of distance from the leading edge. Within the framework of boundary-layer approximations, exact solution is obtained for the special case where surface temperature varies linearly with x , i.e., $\lambda = 1$. For other values of λ , approximate solutions based on local similarity and local non-similarity models are obtained. It is found that the local similarity solutions are sufficiently accurate for all practical purposes.

Analytical expressions for local surface heat flux and overall surface heat flux are obtained.

- D. Nield, D. A. 1968. Onset of thermohaline convection in a porous medium. *Water Resour. Res.* 4:553-560.

The problem of the onset of convection, induced by buoyancy effects resulting from vertical thermal and solute concentration gradients, in a horizontal layer of a saturated porous medium, is treated by linear perturbation analysis. It is shown that oscillatory instability may be possible when a strongly stabilizing solute gradient is opposed by a destabilizing thermal gradient, but attention is concentrated on monotonic instability. The eigenvalue equation, which involves a thermal Rayleigh number R and an analogous solute Rayleigh number S , is obtained, by a Fourier series method, offer a general set of boundary conditions. Numerical solutions are found for some special limiting cases, extending existing results for the thermal problem. When the thermal and solute boundary conditions are formally identical, the net destabilizing effect is expressed by the sum of R and S .

- E. Wooding, R. A. 1963. Convection in a saturated porous medium at large Rayleigh number or Peclet number. *J. Fluid Mech.* 15:527-544.

When the dimensions of a convective system in a saturated porous medium are sufficiently great, diffusion effects can be neglected except in regions where the gradients of fluid properties are very large. A boundary-layer theory is developed for vertical plane flows in such regions. In special cases, the theory is equivalent to that for laminar incompressible flow in a two-dimensional half-jet, or in a plane jet or round jet, for which similarity solutions are well known.

A number of experiments have been performed using a Hele-Shaw cell immersed in water, with a source of potassium permanganate solution located between the plates. At very small values of the source strength, a flow analogous to that of a plane jet from a slit is obtained. The distance advanced by the jet front, or cap, is measured as a function of time, and the velocity is found to be nearly proportional to the velocity of the fluid on the axis of the steady jet behind the cap, as given by the similarity law of Schlichting and Bickley. At large values of the source strength, a two-dimensional 'broad jet' of homogeneous solution descending under gravity is produced; the shape of the flow region can be calculated with little error from potential theory, neglecting the effect of the mixing layers.

A possible example of a mixing layer observed in a geothermal region is examined. The theoretical form of the temperature distribution is calculated numerically, taking into account the large viscosity variation with temperature and also the possibility of a large permeability variation. These effects

are found to have less influence upon the solution than might have been expected. Quantitative values obtained for the physical parameters are consistent with other geophysical observations.

VIII. JOURNAL ABBREVIATIONS LIST

Ag. Expt. Stn. Spec. Rep.	Agricultural Experiment Station Special Report
Agr. and For. Meteorol. Conf. AMS	Agricultural and Forest Meteor- ology Conference, American Meteorological Society
Agr. Meteor.	Agricultural Meteorology
Agron. J.	Agronomy Journal
ASAE Trans.	Transactions of the American Society of Agricultural Engineers
ASME Pap.	American Society of Mechanical Engineering Paper
Aust. J. of Soil Res.	Australian Journal of Soil Research
Boundary Layer Meteor.	Boundary Layer Meteorology
Brit. J. Appl. Phys.	British Journal of Applied Physics
Can. Geotech. J.	Canadian Geotechnical Journal
Can. J. Earth Sci.	Canadian Journal of Earth Science
Can. J. Soil Sci.	Canadian Journal of Soil Science
Heat Transfer (Sov. Res.)	Heat Transfer (Soviet Research)
Highway Res. Bd. Spec. Rep.	Highway Research Board Special Report
Int. Congr. Soc. Soil Sci. Trans.	Transactions of the Congress of the International Society of Science
Int. Congr. Soil Sci. Trans.	Transactions of the Congress of the International Society of Soil Science
Int. J. Heat Mass Transfer	International Journal of Heat and Mass Transfer
Israel J. Agric. Sci.	Israel Journal of Agricultural Science
J. Fluid Mech.	Journal of Fluid Mechanics

J. Geophys. Res.	Journal of Geophysical Research
J. Hydrol.	Journal of Hydrology
J. Indian Soc. Soil Sci.	Journal of the Indian Society of Soil Science
J. Meteor.	Journal of Meteorology
J. of Appl. Meteor.	Journal of Applied Meteorology
J. of Soil Sci.	Journal of Soil Science
National Res. Council Prac. Highway Res. Board Mtg.	National Research Council Proceedings, Highway Research Board Meeting
New Zeal. J. Agr. Res.	New Zealand Journal of Agricultural Research
New Zealand J. of Geol. and Geophys.	New Zealand Journal of Geology and Geophysics
New Zeal. J. Sci.	New Zealand Journal of Science
Proc. South Dakota Acad. Sci.	Proceedings of the South Dakota Academy of Science
Soil Sci.	Soil Science
Sov. Meteorol. Hydrol.	Soviet Meteorology and Hydrology
Sov. Soil Sci.	Soviet Soil Science
SSSAJ	Soil Science Society of America Journal (formerly SSSAP)
SSSAP	Soil Science Society of America Proceedings
Trans. AGU	Transactions of the American Geophysical Union
Water Resour. Res.	Water Resources Research

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