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**PRELIMINARY STUDY OF RADIOACTIVE
WASTE DISPOSAL
IN THE VADOSE ZONE**

PURCHASE ORDER NO. 2649709

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ABSTRACT

To investigate the characteristics of the vadose zone with respect to radioactive waste disposal, the mechanics of unsaturated flow in arid regions and the geohydrology of four areas with a deep water table were studied.

The studies indicated that (1) arid sites with a water table deeper than 200 m can be found in at least three distinct geologic settings in the western United States, (2) the physics of unsaturated flow in soils and rock with interstitial porosity at low water contents, particularly under thermal gradients, is not yet completely understood, and (3) under certain conditions unsaturated flow can be so slow that analytic modeling of an unflawed repository is unnecessary to prove effective containment.

KEYWORDS: Radioactive waste disposal, mechanics of unsaturated flow, unsaturated soil, unsaturated rock, vadose zone, deep water table, hydraulic conductivity, volumetric water content, thermal gradient, modeling.

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LIST OF SYMBOLS

D	Diffusivity (l^2t^{-1})
D_0	Liquid Moisture Diffusivity (l^2t^{-1})
D_T	Thermal Moisture Diffusivity (l^2t^{-1})
$f(\theta)$	Function of θ
i	Cumulative Infiltration ($l^3l^{-2}t^{-1}t$) or (l)
K	Hydraulic Conductivity (lt^{-1})
K_0	Saturated Hydraulic Conductivity (lt^{-1})
l	Length (l)
m	Mass (m)
q	Moisture Flux ($l^3l^{-2}t^{-1}$) or (lt^{-1})
R	Rain Intensity (lt^{-1})
T	Temperature
t	Time (t)
u	Rate of Advance of Wet Front (lt^{-1})
v_0	Rate of Infiltration ($l^3l^{-2}t^{-1}$) or (lt^{-1})
x	Horizontal Distance (l)
z	Vertical Distance (l)
1D	One-dimensional Analysis
2D	Two-dimensional Analysis
∞	Infinity
α	Experimental Constant
β	Experimental Constant
∇	$\text{Del} = \tilde{i} \frac{\partial}{\partial x} + \tilde{j} \frac{\partial}{\partial y} + \tilde{k} \frac{\partial}{\partial z}$

θ	Volumetric Water Content ($l^3 l^{-3}$)
θ_n	Volumetric Water Content At $t = 0$ ($l^3 l^{-3}$)
θ_o	Volumetric Water Content At Saturation ($l^3 l^{-3}$)
ϕ	Total Potential Per Unit Volume ($\psi + \psi_g$) ($ml^{-1} t^{-2}$)
ψ_g	Gravitational Potential Per Unit Volume ($ml^{-1} t^{-2}$)
ψ	Matric Potential Per Unit Volume ($ml^{-1} t^{-2}$)
ψ_π	Osmotic Potential Per Unit Volume ($ml^{-1} t^{-2}$)

1. INTRODUCTION

The potential for disposal of radioactive waste materials in the unsaturated, or vadose, zone in arid regions has received relatively little attention.

Some characteristics of the vadose zone, such as extremely slow transport rates and availability in large, unpopulated tracts of federal land, are favorable with respect to the disposal of radioactive waste material; others, such as shallow depths, uncertainties regarding future climates and, for soils, low strengths and low thermal conductivities, are relatively unfavorable.

This study, authorized by the Lawrence Livermore Laboratory under Purchase Order 2649709, is concerned with the engineering feasibility of using the vadose zone as a repository, with respect to both the movement of moisture through soil and rock in arid regions and the hydrogeology of four areas with a deep water table. Excluded from this study are questions of heat flow, mined openings in soil, and future climates.

The report comprises four sections in addition to the Introduction. Section 2 is an Executive Summary, which includes the significant conclusions of the study. Section 3 is an outline of the status of knowledge of flow at depth in soils with very low water content. In Section 4, the hydrogeology is described for four selected arid areas in the western United States which have deep water tables. Section 5 is an outline of the status of information available for modeling radionuclide transport from a vadose zone repository.

2. EXECUTIVE SUMMARY

2.1 Introduction

Most concerns about the feasibility of geologic disposal of radioactive waste center on two deleterious effects of groundwater: canister dissolution and the transport of radio-nuclides. Placing the waste in the unsaturated, or vadose, zone above the water table may mitigate or eliminate these effects.

Two topics investigated in this study, the mechanics of unsaturated flow and the hydrogeology of four selected arid regions, show favorable conditions for radioactive waste disposal in the vadose zone. However, a lack of field data on the movement of soil moisture in arid regions precludes definitive conclusions at this time. Several topics not considered in this study will require investigation before the feasibility of the overall concept can be shown. A summary of results of the study follows.

2.2 Mechanics of Unsaturated Flow

- Although the fundamental physics of unsaturated flow in porous media at very low water contents and under non-isothermal conditions is still under research, the available data suggest that flow in arid site vadose zones, under the proper conditions, can be at velocities slower than 10^{-4} m/yr.
- Unsaturated flow in fractured rock is fundamentally different from unsaturated flow in soil or rock with interstitial porosity. No theory has been formulated for moisture transport in fractured rock and essentially no applicable field studies have been conducted. Observations suggest that rainfall percolates readily through fractured rock to the water table at velocities of about 10^{-2} to 10^{-3} cm/sec.
- Unsaturated flow in porous media at volumetric water contents greater than about 10% is understood well enough to be modeled for various engineering and agricultural applications. Migration of liquid wastes is studied with numerical models, as is the movement of irrigation water in agriculture. These models have been verified only in the upper few meters of soil and at water contents well above those observed at depth in arid vadose zones.

- At volumetric water contents less than about 10%, unsaturated flow is so poorly understood that for some field applications neither the rate nor the direction of flow can be predicted. Because this topic is of minor importance for most agricultural and engineering applications, where water contents are higher, it has received little attention until recently. Upper limits of flow rates can be estimated, however, by using data at water contents of about 10%.

- Experiments at the Hanford Reservation in Richland, WA during the past five years have shown that previous theories, applicable to wetter soils at shallower depths, which hold that pulses of infiltrating moisture can propagate relatively unmodified to the water table, are not applicable to relatively dry soils at depths greater than several meters. Instead, the experiments show that an initially distinct pulse of moisture becomes broad and indistinct as it moves downward. Summer evaporation begins to draw water back to the surface and eliminates most or all of the moisture pulse. Below approximately 12 meters, water content does not vary with time.

- If climatic variations should cause the ground surface to become inundated for a sufficient period of time, water could infiltrate downward under a gradient of one with a seepage velocity that approaches the saturated hydraulic conductivity, K_o . Eventually the entire soil profile between the ground surface and the original groundwater table becomes wetted. For example, with $K_o = 10^{-5}$ cm/sec, a groundwater depth of 1,000 m and an effective porosity of 0.30, inundation would cause percolation to the water table in about 100 years.

2.3 Hydrogeology of Four Selected Arid Regions

Four geologically distinct areas with relatively deep water tables were investigated to assess the potential for finding a repository location with a deep and stable water table. A brief summary of the four areas, presented in Table 2.1, shows the status of knowledge of each area with respect to both unsaturated and saturated hydrogeology. Specific conclusions include the following:

2.3.1 General

- Deep water tables occur wherever a highly transmissive aquifer and/or a small amount of groundwater recharge combine with an aquifer discharge point topographically much lower than the ground surface at the area of interest.

- Relatively little information exists regarding the flow of groundwater at depths below several hundred meters in arid regions because few people inhabit such regions, and because water supply comes largely from reservoirs rather than from wells. Many of the available data come from national laboratories such as the Nevada Test Site (NTS), the Idaho National Engineering Laboratory (INEL) and the Hanford Reservation (Hanford), and hydrogeologic studies cited in this report were selected largely from these areas.

2.3.2 Great Basin, Nevada

- This area includes the Nevada Test Site (NTS).
- The vadose zone extends as deep as 1,000 m beneath the ground surface in the Great Basin.

- The basins and ranges constitute two different types of vadose zones in the Great Basin. The vadose zone material in the basins, or flats, consists primarily of thick deposits of porous alluvium underlain by volcanic tuffs and minor carbonate rocks.⁽¹⁾ The vadose zone material in the ranges, which separate the basins, consists largely of volcanic tuffs underlain by carbonate rocks.

- Moisture flows interstitially through the soils in the vadose zone in the basins and through fractures in the rock in the vadose zone in the ranges.

- We have found no analytic studies of unsaturated flow at depth in either type of vadose zone in the Great Basin. In the basins, researchers have assumed that all rainfall infiltrating the alluvium evaporates, in which case downward percolation to the water table would be non-existent. However, a possible exception to this assumption is represented by the fact that surface waters flow into large earth cracks that exist in southern Yucca Flat. No confirmatory studies have been made. Observations at Rainier Mesa at NTS suggest that rainfall percolates quickly to the water table in the ranges, where no thick soil cover stores rainfall until it can be evaporated.

- Groundwater in the saturated aquifer beneath the vadose zone flows mainly through fractures and solution cavities. The complex geologic structure strongly controls both the direction and rate of groundwater flow.

- Groundwater flows at velocities of about 10 m/yr and has a maximum aquifer residence time, from Yucca Flat to

⁽¹⁾ A glossary of select geological terms is included at the end of the Executive Summary.

the discharge point, of about 10^4 years. Some data, however, suggest that flow in certain sections of the aquifer can be several orders of magnitude higher.

- No models have been developed to analyze groundwater flow in the Ash Meadows (NTS area) or White River groundwater systems of the Great Basin. Borg et al. (1976) concluded that due to complex geologic structure and a lack of parameter data, one-dimensional models would give results as reliable as those from two-dimensional models.

- The Great Basin region is tectonically active and moderate earthquakes and active faults are observed. The potential effect of these phenomena on radioactive waste containment in the unsaturated zone requires investigation.

2.3.3 SNAKE RIVER PLAIN, IDAHO

- This area includes the Idaho National Engineering Laboratory (INEL).

- The vadose zone extends as deep as 300+ m beneath the ground surface on the Snake River Plain.

- The material in the vadose zone consists primarily of horizontally-layered flood basalts separated by relatively thin layers of clastic sediments.⁽¹⁾

- Because downward-percolating water flows through fractures and large-scale voids (such as lava tubes) in the basalt, unsaturated flow theory for porous media cannot be used to predict the rate of moisture movement through the basalt.

- We have found no studies of predictions of future climates or of the potential effect of future climates on the elevation of the water table in the Snake River Plain.

- A two-dimensional groundwater model that correctly accounts for the hydrologic water balance is based on the assumption that, on the average, 10% of the annual precipitation percolates to the water table. Other studies show that most of this recharge occurs below a small portion of the area and that areas of several square kilometers have a net infiltration of only about 1.5 cm/yr. Further studies are neces-

(1) A glossary of select geological terms is included at the end of the Executive Summary.

sary to determine if infiltration of this magnitude precludes the use of such locations for a vadose zone repository.

- Water flows in the water table aquifer beneath the vadose zone mainly through fractures and macroscopic voids with a maximum velocity of about 300 m/yr and an aquifer residence time of about 60 to 300 yrs.

- The numerous test wells and a geologically simple aquifer geometry have permitted the formulation of a reasonably-accurate, two-dimensional, finite-difference groundwater model that simulates the general transport of ^3H and ^{90}Sr through the aquifer.

2.3.4 Columbia River Plateau, Washington

- This area includes the Hanford Reservation.

- The vadose zone extends as deep as about 100 m beneath the ground surface at the Hanford Reservation.

- The material in the vadose zone consists of coarse-to-fine-grained sediments and a minor amount of horizontally-layered basalt at depth.

- Precipitation infiltrates the upper few meters of the vadose zone by interstitial, unsaturated flow.

- Comparison of field measurements with mathematical models show that the flow of moisture in the upper 12 meters of soil can be simulated, but flow of moisture at depth has not been modeled or measured.

- Water flows interstitially through sediments in the water table aquifer directly below the vadose zone.

- Water flows in the water table aquifer with a velocity of about 100 m/yr and has an aquifer residence time of several days to several hundred years, depending upon the distance from the Columbia River discharge point.

- Models for describing groundwater and contaminant movement have been developed and tested against analytical solutions and limited field data.

- Additional data and work are needed to reduce the uncertainty in the model predictions.

2.3.5 Copper Mine Area, Colorado Plateau, Arizona

- The vadose zone extends as deep as about 250 m beneath the ground surface at the Copper Mine area of the Colorado Plateau.
- The material in the vadose zone consists of medium- to fine grained, partially cemented, porous sandstone. Two to five m of relatively impermeable siltstone overlies the sandstone and apparently retards or eliminates infiltration of precipitation.
- The porous nature of the sandstone in the vadose zone suggests that any downward-infiltrating water would flow interstitially. Field experiments would be required to test this hypothesis.
- Because no field experiments have been conducted on the unsaturated zone since its discovery at the Copper Mine area, the rate and character of moisture flow remain unknown.
- A water infiltrating the vadose zone flows interstitially through the lower portion of the Navajo sandstone to the Colorado River about 15 km west of the Copper Mine area.
- Groundwater, where found to the west of the Copper Mine area, flows westward with an estimated velocity of about 1,000 m/yr and has an estimated aquifer residence time of about 15 yrs.
- Mathematical models have not been applied to predict groundwater flow in the vicinity of the Copper Mine area.

GLOSSARY

- Alluvium - A general term for clay, silt, sand, gravel, or similar unconsolidated detrital material deposited during comparatively recent geologic time by water.
- Volcanic tuff - A consolidated deposit of volcanic ash and dust with or without minor amounts of other sediments such as sand or clay.
- Carbonate rock - A sedimentary rock, e.g., limestone or dolomite, which consists of more than 50% by weight of carbonate minerals such as CaCO_3 or MgCO_3 .
- Flood basalt - Basalt is a dark-colored, fine-grained igneous rock, and flood basalts result from an extensive lava flow or successive lava flows of high-temperature, fluid basalt from fissure eruptions which accumulate to form a plateau.
- Clastic sediment - A sediment formed by the accumulation of fragments derived from pre-existing rocks or minerals and transported as separate particles to their place of deposition by purely mechanical agents (such as water, wind, ice and gravity), e.g., gravel, sand, silt, clay.
- Lava tube - A hollow space beneath the surface of a solidified lava flow, formed by the withdrawal of molten lava after the formation of the surficial crust.
- Vadose Zone - The unsaturated soil or rock above the water table and associated capillary fringe (also known as the "zone of aeration" or the "unsaturated zone").

TABLE 2.1 - SUMMARY OF HYDROGEOLOGIC INFORMATION
AT FOUR REGIONS WITH DEEP WATER TABLES

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Zone		Great Basin, Nevada	Snake River Plain, Idaho	Columbia River Plateau, Washington	Copper Mine Area, Arizona
Unsaturated (Vadose)	Maximum thickness of vadose zone	1,000 m	300+ m	100 m	250 m
	Geologic material in vadose zone	Alluvium; volcanic tuff (some zeolit- ized); limestones; sandstones, shales	Basalt with fine grained sedimen- tary interbeds; coarse to fine grained surficial sediments	Coarse to fine grained sedi- ments and minor basalt	Siltstones and sandstones
	Type of unsatu- rated flow	Interstitial in the alluvium, zeo- lite tuff, shales, and some limestones. Fracture in the tuff and some limestones	Flow in macro- scopic voids in the basalt; in- terstitial flow in sediments	Flow in macro- scopic voids in the basalt; in- terstitial flow in sediments	Interstitial (?)
	Status of know- ledge of unsatu- rated flow at depth	Unknown	Estimated to be about 2-3 cm/yr downward	Estimated to be extremely slow	Unknown
	Existing model	No	No	No	No

TABLE 2.1 - SUMMARY OF HYDROGEOLOGIC INFORMATION
AT FOUR REGIONS WITH DEEP WATER TABLES

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Zone		Great Basin, Nevada	Snake River Plain, Idaho	Columbia River Plateau, Washington	Copper Mine Area, Arizona
Saturated (Phreatic)	Type of flow in main aquifer	Fracture.	Flow in macro- scopic voids	Interstitial	Interstitial (?)
	Approximate flow velocity	10 m/yr	300 m/yr	100 m/yr	300 m/yr
	Approximate maximum ground- water residence time	10^4 yr	300 yr	200 yr	50 yr
	Status of know- ledge of hydro- logic parameters	Minimal	Good	Good	Minimal
	Existing model	No	Yes	Yes	No

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3. THE FLOW OF WATER IN UNSATURATED SOILS

3.1 Introduction

The flow of water in unsaturated soil exhibits several unique characteristics which favor the storage of nuclear waste in arid-site vadose zones. In general, water flows very slowly in soils which are not saturated, because liquid continuity, common to saturated flow, is severed, and water is attracted to soil particles by adsorptive and capillary forces.

Surficial soils in arid areas typically have very low moisture content, and little, if any, water infiltrates the soil profile deeper than about 10 to 20 m. Without flow of water through a depository, there can be no significant migration of water-borne radionuclides. Further, temperature gradients, such as may be created by the burial of high-level waste or spent fuel, cause whatever small amount of moisture that may exist to flow away from the heat source. Therefore, it appears that arid-site vadose zones may be feasible for nuclear waste disposal.

Prediction of the feasibility of vadose-zone repositories, (VZR) requires knowledge of both the mechanism by which water flows through unsaturated soil and the mathematical equations which have been proposed to model such flow. Consequently, this part of the report consists of an overview of the literature on unsaturated flow, and contains a general discussion of soil-water relationships, unsaturated flow characteristics, and current methods of simulating flow mathematically. The section concludes with a brief discussion of the applicability of mathematical simulation techniques to the modeling of vadose-zone repositories.

Appendix A contains a series of tables which annotate the most important literature that was reviewed. Appendix B contains plots of data that illustrate the basic soil-water relationships discussed in Section 3.2. Appendix C contains data from infiltration experiments that serve further to illustrate basic unsaturated flow phenomena. Appendix D contains a summary of laboratory and field methods for measuring the water retention and flow characteristics of unsaturated soils. Appendix E contains data on the flow of moisture in unsaturated soil when subjected to thermal gradients.

3.2 Soil-Water Relationships

3.2.1 Soil-Water Potential

Unsaturated soils consist of soil, water, and air - a three phase system. Typically, the potential energy (explained below) of such a system is less than that of free water at the same temperature, elevation, and atmospheric pressure (Gardner, 1960). The measure of this energy has received various names in the literature, the most common being soil-moisture potential (negative) or suction (positive) and is usually designated by the symbol Ψ . Hillel (1971) described the state of water in a soil as follows:

"Soil water is subject to a number of force fields, which cause its potential to differ from that of pure free water. Such force fields result from the attraction of the solid matrix for water, as well as from the presence of solutes and the action of external gas pressure and gravitation."

Therefore, the total soil water potential can be expressed as:

$$\phi = \Psi_m + \Psi_g + \Psi_\pi + \dots \quad (\text{Yong and Warkentin, 1975})$$

where ϕ , the total potential, has been defined as:

"the amount of work that must be done per unit quantity of pure water in order to transport reversibly and isothermally an infinitesimal quantity of water from a pool of pure water at a specified elevation and atmospheric pressure to the soil water (at the point under consideration)" (Hillel, 1971).

Ψ_m , the matric potential, is related to the surface tension of air-water interfaces and the adsorption forces of soil-water interfaces. This concept is illustrated in Figure 3.1. The adsorption forces, caused by hydration envelopes around soil particles, become very important in clays which possess electrical double layers and exchangeable cations (Richards et al., 1964).

Ψ_g , the gravitational potential, is related to the elevation of soil water above the reference pool. Figure 3.2 illustrates the flow of water in a soil column due to the action of a gravitational potential.

ψ_{π} , the osmotic potential, is related to the effect of solute concentration on the vapor pressure of soil water. The osmotic effect becomes important when vapor diffusion is an important contributor to moisture flow (Richards et al., 1964).

The dots at the end of the equation signify that other potential terms are theoretically possible.

3.2.2 Water Retention

Reduction of the matric potential of a free-draining soil induces a flow of water out of the pore spaces and consequently lowers the soil-water content. Increasingly larger imposed suctions (more negative potentials) cause increasingly smaller pores to empty and the water content continues to decrease (Hillel, 1971; Horton and Hawkins, 1965; Remson et al., 1959; Yong and Warkentin, 1975). Figure 3.3 illustrates this phenomenon for three types of soil. Notice that clay has a tendency to retain more moisture than a sand, even when subjected to suctions of several thousand atmospheres.

The moisture content of sand is controlled chiefly by capillarity (soil structure) while the moisture content of clay is controlled by adsorption (soil texture and specific surface area). This makes it very difficult, if not impossible, to predict a priori, the soil-water retention characteristics of a soil which contains a mixture of sand and clay.

Referring again to Figure 3.3, notice that each soil is represented by two water-retention curves, one for wetting (increasing water content) and one for drying (decreasing water content). This phenomenon, known as hysteresis, has added to the difficulty of developing equations to model the flow of moisture in unsaturated soils (Klute, 1952; Biswas et al., 1966; Bear et al., 1968; Hillel, 1971; Yong and Warkentin, 1975).

3.2.3 Field Capacity

If a soil is first saturated and then allowed to drain under the influence of gravity, the water content will decrease until the drainage becomes "imperceptibly small" (Remson et al., 1959; Horton and Hawkins, 1965; Hillel, 1971; Yong and Warkentin, 1975). The water content of the soil at this time has been termed the "field capacity" of the soil. Although quantification of field capacity is difficult and subjective, the concept of field capacity is useful in describing the behavior of moisture in unsaturated soils, especially during infiltration. Infiltration, the processes by which

moisture is added to the soil profile from the ground surface, is described in Section 3.3.4.

3.3 Mechanisms of Unsaturated Flow

This section on the mechanics of flow in unsaturated media is presented in three parts: flow under isothermal conditions, flow due to a thermal gradient, and infiltration of surface water downward through the vadose zone. For the first two cases, the discussion consists of a description of (1) the phenomena thought to be operative, (2) the mathematical equations that are used to predict flow, and (3) the techniques available for measurement of the parameters that influence unsaturated flow and for which values are required to solve the equations.

The portion on infiltration contains a description of the factors which control the variation of water content with depth in the vadose zone. Also included is a discussion of the theoretical predictions of infiltration rates under certain conditions and field observations of the vertical extent of infiltration in arid areas.

3.3.1 Isothermal Flow

Water flows from areas of high potential (low suction) to areas of low potential (high suction). The rate at which water flows along a potential energy gradient is a function of the magnitude of the applied gradient and the hydraulic conductivity of the soil. Hydraulic conductivity is a measure of the ease with which water can be transmitted through the soil. For a saturated soil, the hydraulic conductivity, K , is essentially constant, for many soils, and can easily be measured. In unsaturated soils, however, the hydraulic conductivity is not a constant or intrinsic property of a particular soil. As a soil drains due to an applied potential gradient, the larger pore spaces empty first and there is a corresponding reduction in the conductivity of the soil. Thus, air-filled pore spaces essentially become barriers to flow by decreasing the cross-sectional area and increasing the tortuosity of the liquid flow path (Hillel, 1970; Yong and Warkentin, 1975; Childs, 1969).

The hydraulic conductivity of unsaturated soils is a function of both moisture content and soil-water potential. Furthermore, both of these factors are interdependent: as the potential decreases (increasing suction) the moisture content also decreases and, thus, the hydraulic conductivity decreases.

Figure 3.4 illustrates the relationship between soil-water potential and hydraulic conductivity for two soils of different texture.

It is interesting to note in Figure 3.4 that the unsaturated hydraulic conductivity of clay may actually exceed that of sand at large negative values of soil-water potential. This is due to the greater liquid continuity in clay. This is contrary to saturated hydraulic conductivity which is typically independent of head (positive soil-water potential) and is consistently higher for coarse-grained material than for fine-grained material. Therefore, in an unsaturated fine-grained soil profile, a sand layer, rather than enhancing flow, may actually impede moisture flow, at least temporarily.

3.3.1.1 Equations

L. A. Richards (1931) extended Darcy's law to unsaturated flow with the assumption that hydraulic conductivity, K , was a unique function of matric potential, Ψ , (the soil-water potential with gravitational and osmotic components excluded), i.e., $K = K(\Psi)$. He proposed that

$$q = -K(\Psi) \nabla H, \quad (1)$$

where q = flow rate and ∇H = potential gradient ($H = \Psi + \Psi_g$ in the terminology of this paper).

This approach cannot account for hysteresis of the $K(\Psi)$ function so the use of equation (1) must be limited to monotonic potential (or moisture content) change.

A more general approach, which would include transient situations, would be to combine equation (1) with the continuity equation:

$$\frac{\partial \theta}{\partial t} = -\nabla \cdot q \quad (2)$$

therefore,

$$\frac{\partial \theta}{\partial t} = \nabla \{K(\Psi) \nabla H\} \quad (3)$$

where θ = volumetric moisture content (the ratio of the volume of water held in the soil to the total volume of the soil).

Expressed in terms of one-dimensional vertical flow with ψ_q expressed as elevation z , equation (3) becomes

$$\frac{\partial Q}{\partial t} = \frac{\partial}{\partial z} \left\{ K(\psi) \left(\frac{\partial \psi}{\partial z} \right) \right\} + \frac{\partial K(\psi)}{\partial z} \quad (3a)$$

Hillel (1971) describes a modification of the above mathematics by introducing a "diffusivity" term, D , from the theory of heat flow and assuming no hysteresis (and hence $K(\psi) = K(\theta)$:

$$D(\theta) = K(\theta) \frac{\partial \psi}{\partial \theta} \quad (4)$$

Therefore, for one-dimensional vertical flow (Philip, 1957a):

$$\frac{\partial Q}{\partial t} = \frac{\partial}{\partial z} \left(D \frac{\partial \theta}{\partial z} \right) + \frac{\partial K}{\partial z} \quad (5)$$

Use of the word "diffusivity" does not necessarily mean that water movement occurs by diffusion; it is a term, borrowed from the field of thermodynamics, used in heat flow equations. The above equation is not applicable when hysteresis is large, where the soil is layered, or where temperature gradients exist (Hillel, 1971).

Various numerical solutions to equations (3) and (5) have been proposed and verified for specific conditions (Klute, 1952; Philip, 1957a, b; Geisel et al., 1963; Green et al., 1964; Jackson, 1964; Rubin et al., 1964; Wind and Van Doorne, 1975). For example, soil scientists apply the flow equation to the prediction of water balances when designing agricultural irrigation programs. They must estimate how much of the annual rainfall infiltrates the soil, and how much subsequently evaporates, to assure that sufficient water remains in the soil to feed the crops. Table 1 of Appendix A contains a summary of solutions of the equation with their assumptions and applications.

3.3.1.2 Parameters

Common to all solutions of the flow equations is the need to define the value of hydraulic conductivity or diffusivity over the range of soil-water potentials and water contents that exist in the field. Many techniques have been used to measure these parameters either in the laboratory or in the field and several of the most common are described in detail in Appendix D.

Measurement of soil characteristics in the laboratory is necessarily restricted to small samples of soil which are either "undisturbed" or "reconstituted." The tests are necessarily only approximations to in-situ behavior, since even the most careful soil sampling does not provide totally "undisturbed" samples, nor does reconstitution of disturbed samples accurately reproduce the in-situ soil structure. Nevertheless, in the absence of better techniques, the soil scientist often relies on the results of laboratory tests to predict the flow of moisture in unsaturated soils. Croney and Coleman (1960) discuss some of the earlier techniques of testing unsaturated soils and some of the more contemporary techniques are discussed below and in Appendix D.

Hydraulic conductivity, K , and diffusivity, D , can be measured directly in either steady state or transient experiments. Steady state tests involve applying a potential gradient to a soil column and measuring the amount of water flowing through the soil. Transient experiments, typically referred to as outflow or infiltration experiments, involve measuring water content as a function of time after applying a change in potential.

The results of three types of laboratory test can be combined to yield a plot of θ vs Ψ for a large range of Ψ . All three techniques involve applying a known potential to a soil sample and measuring the resulting equilibrium water content. For potentials greater than -0.8 bar (-0.8 atm) a tension plate assembly can be used to apply a potential to the base of a sample (Youngs, 1964). Potentials down to about -25 bars can be attained with a pressure plate assembly by applying a positive pressure to the top of the sample (Youngs, 1964). Potentials below -25 bars require indirect methods such as moisture blocks. Appendix D includes a discussion of these and other measurement techniques.

Remson et al. (1959) used tensiometers to measure soil-water potential in-situ while studying ground-water recharge due to infiltration. A tensiometer is a water-filled, permeable, ceramic cup or cell connected to a manometer. When the cup is placed in the soil, the soil suction tends to pull water out of the cup which results in a negative pressure reading on the manometer.

For lower potentials, thermocouple (or thermistor) psychrometers have been used successfully (Richards et al., 1964; Letey et al., 1969; Kay and Low, 1970; Isaacson et al., 1974). The psychrometer measures vapor pressure and temperature, thus permitting calculation of soil-water potential.

Field measurements can also be taken to determine the in-situ unsaturated hydraulic conductivity of a soil. The most common technique is referred to as the sprinkling or infiltration method and has been used by several researchers to test theoretical predictions (Green et al., 1964; Rose et al., 1965; Rubin et al., 1964). This method involves wetting the soil surface and determining the profiles of water content and potential as moisture redistribution takes place.

Appendix D contains a review of existing techniques for both laboratory and field measurements of water content, potential, conductivity and diffusivity.

3.3.2 Non-Isothermal Flow

Since Bouyoucus (1915) first demonstrated that moisture flows from areas of high temperature toward areas of low temperature, many scientists have investigated the mechanisms of moisture flow in unsaturated soil and have tried to predict flow rates mathematically. Early investigators proposed theories of flow which were based on the results of their own tests. However, there were discrepancies between test results and several phenomena remain unexplained.

It is generally accepted that vapor flow accounts for most of the moisture transfer due to applied thermal gradients (Rollins et al., 1954; Gurr et al., 1952; Taylor and Cavazza, 1954; Philip and DeVries, 1957; Cary and Taylor, 1962a, b; Cary, 1965, 1966). The primary mechanisms of thermally induced flow are thought to be as follows (Cary, 1966):

1. Surface tension gradients: Increasing temperature decreases surface tension and causes movement down the thermal gradient due to the induced surface tension gradients. Since moisture flows from warm zones to cool zones in saturated soil also (with no air-water interfaces), surface tension cannot be the sole driving force.
2. Soil-water potential: As temperature decreases, the soil-water potential also decreases (increasing suction) and a flow is induced from zones of high temperature to zones of low temperature. This idea formed the basis of the equations presented by Philip and DeVries (1957) discussed later in this section.

3. Random kinetic energy: Energy changes due to variation in hydrogen bond distribution caused by thermal gradients may also contribute to flow.
4. Osmotic gradients: Most dissolved salts will diffuse through a solution from warm to cool areas (Soret effect). Water tends to follow the salt in order to maintain chemical equilibrium.

The influence of temperature on flow increases as the water content decreases because vapor flow accounts for more moisture transport than liquid flow at low moisture contents.

3.3.2.1 Equations

Mathematical description of the flow of moisture due to thermal gradients in unsaturated soil is presently in a nascent state. The principal reason is that the flow of moisture and heat is a coupled phenomenon: as thermal gradients cause moisture to flow, the flow of moisture (by evaporation and condensation) causes the transfer of heat. Therefore, equations of flow must combine heat transfer equations with mass transfer equations, both of which are non-linear.

Philip and DeVries (1957) made one of the first attempts to model moisture flow caused by thermal gradients. They combined diffusion theory with the theory of liquid flow due to capillarity and proposed an approximate equation for moisture flow:

$$-q = D_0 \frac{\partial \theta}{\partial z} + D_T \frac{\partial T}{\partial z} + K \frac{\partial \psi}{\partial z} \quad (6)$$

where $D_0 \frac{\partial \theta}{\partial z}$ equals the flow of moisture due to a moisture content gradient,
 $D_T \frac{\partial T}{\partial z}$ equals the flow of moisture due to a thermal gradient, and
 $K \frac{\partial \psi}{\partial z}$ equals the flow of moisture due to gravity.

Cary and Taylor (1962a, b) proposed a different set of equations based on the thermodynamics of irreversible processes. They subsequently revised their equations after further laboratory experiments (Taylor and Cary, 1964; Cary, 1965).

In subsequent work by Cassel et al. (1969), Jury (1973), and Jury and Miller (1974) there were conflicting results between experiments and predictions using the two

aforementioned approaches. Some experiments confirmed the equations of Philip and DeVries, while others confirmed the equations of Taylor and Cary. In most experiments, however, moisture flow was several times that predicted by the equations of Taylor and Cary. Jury (1973) summarized previous work and illustrated the similarity between the Philip-DeVries equations and the Cary-Taylor equations. Conflicts between experimental results and proposed theories probably resulted from differences in measurements of the coefficients D_0 , D_T and K . To date, no exact solutions of Equation (6) exists. Several approximations have been proposed but have not been verified by experimentation.

3.3.2.2 Parameters

The coefficients, D and K , must be measured experimentally. K can be measured as a function of θ as previously described, and the variation of K with T "can be assumed to be proportional to the temperature variation of the reciprocal of the coefficient of viscosity of water" (Jury, 1973).

Jury describes several experiments which facilitate the calculation of D and K for θ greater than 7%. For θ less than 7%, there was not good agreement between theory and experiment. However, there was good agreement between theory and experiment for soils at higher water contents. Jury (1973) concluded that "more work remains to be done...before a workable model for transport of moisture in the presence of temperature gradients is ready for application to field situations."

In general, it is accepted that moisture flows along thermal gradients in the direction of decreasing temperature, as originally stated by Bouyoucos (1915). Experiments demonstrating this phenomenon were conducted by Gurr et al. (1952), Taylor and Cavazza (1954), and Rollins et al. (1954). However, although there was a net flux of moisture from the warm side to the cool side, they noticed that salt, which had been used to trace the flow of moisture in the liquid phase, accumulated at the heat source. The implication of such tests is that a mechanism for transporting soluble salts (in the liquid phase) away from a heat source may not exist in an unsaturated medium and that the majority of moisture flow occurs in the vapor phase. However, the salt tracer experiments were on closed systems so the results may not be applicable to free-draining field soils. Further, the results seem to contradict the aforementioned Soret effect, that most dissolved salts will diffuse through a solution from warm to cool areas.

Appendix E contains data on thermally-induced flow in unsaturated soils and some additional discussion of the characteristics of flow and factors that control flow rates.

3.3.3 Infiltration

The entry of water from the soil surface into the soil profile is called infiltration. As water enters the soil profile (from a constant supply at the surface) it will distribute itself as shown in Fig. 3.5. This typical profile was first described by Bodman and Colman (1943) and Colman and Bodman (1944) and has been verified mathematically by Philip (1957c).

Figure 3.5 shows that the moisture profile during infiltration is divided into four characteristic zones: (1) a thin zone (on the order of a few cm) at the soil surface reaches a water content approximating pore space saturation; (2) below this saturated zone the water content decreases with depth to a value approximately halfway between field capacity and saturation (the transmission zone) and remains essentially constant during infiltration; (3) below the transmission zone the water content decreases with depth in the wetting zone and at a given depth increases rapidly with time as infiltration continues; (4) the wetting zone stops abruptly at the wet front which marks the lower extent of infiltration and is characterized by a very steep gradient of moisture content.

There has been some debate (Bear et al., 1968; Hillel, 1971) as to whether the saturation zone actually exists or is a result of experimental error. Regardless, the obvious conclusion from Figure 3.5 is that most of the flow is unsaturated (Bear et al., 1968; Hillel, 1971).

Figure 3.6 illustrates the variation of soil-water potential gradient with time. With time, the potential gradient approaches a constant value equal to or slightly greater than -1 . Thus, the only driving force remaining after long periods of infiltration is gravity.

Remson et al. (1959) studied the flow of rainfall from the ground surface to a shallow water table in a field monitoring program involving a series of tensiometers installed throughout the soil profile. They observed that:

"Applied water does not distribute itself throughout the length of a soil column unless it is in sufficient quantity to bring the entire soil column to field capacity level. Instead, it percolates downward as a nearly saturated belt, leaving little islands of moisture in its wake."

They further observed that:

"Enough water must be applied to wet the entire zone of aeration {unsaturated zone above the water table} to the field capacity level before substantial ground water recharge can occur" (emphasis added).

So, if a soil profile is relatively dry, perhaps due to a long period of low rainfall, substantial rainfall may be required just to wet the profile to field capacity. If sufficient water is not added, groundwater recharge can only occur by very slow drainage or by vapor transfer. Indeed, Isaacson et al. (1974) reported a net moisture loss from the very dry (6% volume moisture content) unsaturated zone on the Hanford Reservation in Richland, WA. This loss was explained as being due to an excess of evapotranspiration over rainfall.

Isaacson and his co-workers (1978) indicated that current data from the lysimeter experiments (Isaacson et al., 1974; Brownell et al., 1975; Last et al., 1976) are not conclusive. Experiments during the past five years have shown that the "wet front at the field capacity" concept of downward percolation may not apply to relatively dry soils at depths greater than several meters. Instead, an initially distinct pulse of moisture becomes broad and indistinct as it moves downward. Summer evaporation begins to draw water back to the surface and eliminates most or all of the moisture pulse. Below approximately 12 meters, water content does not vary with time. This contradicts previous theories, applicable to wetter soils at shallower depths, which hold that pulses of infiltrating moisture can propagate relatively unmodified to the water table.

Horton and Hawkins (1965) performed laboratory experiments to trace the flow of water during and after infiltration and found:

1. "the percolation of groundwater through the soil to the water table is accomplished throughout most of the flow path by downward displacement of water previously retained by the soil at field capacity."

2. "even with abnormally large amounts of rain, and with the soil at field capacity, most of the rainwater that infiltrates large pores flows to smaller pores by the time it has penetrated the large pores a few feet" (below the ground surface).

The first conclusion confirms the findings of Remson et al. (1959), but only at depths of a few meters. The second confirms a well-documented characteristic of unsaturated flow: in unsaturated soil, since water is always under a potential less than atmospheric, water will not move into larger pores where the potential is greater than or equal to atmospheric.

Aylor and Parlange (1973) and Frind et al. (1976) confirmed and extended this second conclusion. Their data indicate that the wetting front in a fine-grained soil stops at a sand or air boundary. The moisture content then increases at the boundary until the suction in the fine-grained soil becomes less than that across the boundary. If there is no restriction to lateral flow, water moves laterally in the fine-grained soil because of the reduced potential there. Thus, accumulation at the boundary is delayed. Frind et al. (1976) refer to this phenomenon as the "wick effect."

3.4 Mathematical Simulation of Unsaturated Flow

3.4.1 Modeling a Vadose Zone Repository

Section 5 of this report presents a discussion of the feasibility of a vadose zone repository (VZR) and elucidates the need for developing a mathematical model to predict the flow of moisture in the unsaturated zone. Such a model requires knowledge of four basic aspects of the hydrologic system:

1. Flow equations.
2. Hydraulic parameters of the porous media.
3. Geometry of the system.
4. Boundary and initial conditions.

This section reviews these aspects as they relate to the mathematical simulation of moisture flow in unsaturated soil and presents some limitations to the application of current theory in predicting the feasibility of the VZR concept.

3.4.2 Equations

The mathematical simulation of water flow in unsaturated soils at high water contents under isothermal conditions is well documented. Many computer codes exist which solve the general non-linear differential equations of flow. Most research has been related to the prediction of infiltration into agricultural soils (i.e., Philip, 1957a, b, c, d, e, f, and g) with some work oriented toward shallow waste disposal (Frind et al., 1976).

The equation of one-dimensional flow in unsaturated soils under isothermal conditions at water contents above field capacity can be expressed in two forms (Section 3.3.1 and Appendix A).

$$1. \quad \Psi\text{-based equation:} \quad \frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{\partial \Psi}{\partial z} \right) + \frac{\partial K}{\partial z} \quad (3a)$$

$$2. \quad \theta\text{-based equation:} \quad \frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left(D \frac{\partial \theta}{\partial z} \right) + \frac{\partial K}{\partial z} \quad (5)$$

where the time rate of change of water content, $\frac{\partial \theta}{\partial t}$, is characterized as a wet front advancing through the soil profile. The symbols used in the equation are described in Section 3.3.1. Two- and three-dimensional forms of the above equations have also been solved.

The mathematical techniques for the simulation of water flow in relatively dry unsaturated soils subjected to thermal gradients (at water contents well below field capacity) have not yet been verified by experiment. The equations most commonly cited in the literature are of the diffusion type first proposed by Philip and De Vries (1957); expressed here in one-dimensional form for simplicity:

$$-q = D_{\theta} \frac{\partial \theta}{\partial z} + D_T \frac{\partial T}{\partial z} + K \frac{\partial \Psi}{\partial z} \quad (6)$$

where $D_{\theta} \frac{\partial \theta}{\partial z}$ equals the flow of moisture due to a moisture content gradient,
 $D_T \frac{\partial T}{\partial z}$ equals the flow of moisture due to a thermal gradient, and
 $K \frac{\partial \Psi}{\partial z}$ equals the flow of moisture due to gravity.

To date, no exact solution of Equation (6) has been developed; however several numerical approximations have been proposed. One such approximation by Baca and King (1978) involves a finite element computer simulation and shows consider-

able promise. However, non-isothermal flow at very low moisture contents is so slow that no experimental techniques are currently available to check predictions made by such computer programs. Further, solution of Equation (6) requires the measurement of the coefficients D_0 , D_T and K , which is extremely difficult and presents a serious limitation to the application of current theory.

3.4.3 Parameters

The equations of unsaturated flow presented above are all of the phenomenological type, hence their solution requires the experimental measurement of certain flow parameters (i.e., D , K , ψ and θ). For the prediction of isothermal flow at high water contents, there are many techniques available by which the required parameters can be measured. These techniques are mentioned in Section 3.3.1 and further explained in Appendix D. For the prediction of non-isothermal flow at low water contents, however, there are no widely accepted techniques for the measurement of the parameters D and K .

Both D and K vary considerably as the water content of the soil changes. Figures 3.7 and 3.8 indicate K can vary several orders of magnitude with very little change in θ and as θ becomes increasingly small, its effect on K increases. Although D is not affected so drastically by θ (Figure 3.9), its experimental measurement at low moisture contents has received little attention in the literature.

3.4.4 Geometry

Geometric considerations for radionuclide transport modeling in the vadose zone are relatively simple compared with modeling in the phreatic, or saturated zone. Gradients are essentially vertical, except in the immediate vicinity of the waste canisters, so one-dimensional models are adequate to represent the physics of flow in-situ. For detailed modeling of waste dissolution, two- or three-dimensional modeling would be required.

3.4.5 Initial and Boundary Conditions

Because moisture flow under non-isothermal conditions is coupled with heat flow, the boundary conditions for a repository model would include time-varying temperatures at the ground surface and at the waste canisters in addition to the fixed moisture content at the water table and the time-varying moisture contents near the ground surface.

3.4.6 Limitations To the Applications of Current Theory

Although many experimenters have verified the theory of isothermal flow in unsaturated soil, several limitations restrict its application. For example, (1) the equations apply only to soils at water contents near or above field capacity, (2) the equations probably do not apply to unsaturated flow through fractured media such as rock, and (3) few documented cases exist in which the experimenters applied isothermal flow equations to the prediction of flow in soils other than agricultural soils in the upper few meters.

The application of moisture flow equations under non-isothermal conditions has many serious limitations: first, the general flow equation has yet to be verified by laboratory or field experiments, and second, the parameters required in flow equation cannot be measured at low water contents. Further, as the water content of soil decreases, flow becomes affected by other potentials, such as osmotic gradients, which are not accounted for in Equation (6).

Before the present theory of unsaturated flow can be applied reliably to the prediction of radionuclide transport from vadose zone repositories, further experimental research must be performed. Experiments must be conducted to verify the flow equations themselves, and improvements must be made in the reliability of laboratory tests to determine flow coefficients.

3.5 Summary

The flow of water in unsaturated soil exhibits several unique characteristics which favor the storage of nuclear waste in arid site vadose zones:

1. Water flows very slowly in soils which are not saturated because liquid continuity is severed and water is attracted to soil particles by adsorptive and capillary forces;
2. The flow rate in unsaturated soil decreases significantly as water content decreases;
3. Water in unsaturated soil is at a negative potential (suction); hence, water will not enter large voids which are at atmospheric pressure (zero potential);
4. The rate at which water infiltrates unsaturated soil is influenced by the intensity and duration of rainfall in addition to the total amount of rainfall.

5. In unsaturated soils which are subjected to thermal gradients, water flows toward lower temperatures.

The above principles suggest that vadose zones in arid regions may offer feasible locations for repositories since they consist of large areas with low potential for radionuclide migration. Arid areas typically contain soils with very low moisture content; little, if any, net infiltration of water into the soil profile; and extremely low rates of moisture flow at depth.

The flow of water in soils at water contents near or above field capacity can be modeled mathematically, and such models have been tested in both field and laboratory experiments. It must be emphasized, however, that these models are based on theories that have been tested at high water contents, near the surface, and for short time periods. At very low water contents, water movement is a coupled phenomenon including the simultaneous flow of moisture and heat. Several equations have been proposed to model such flow but, as yet, these equations have not been tested experimentally. Further, the coefficients required to solve the proposed flow equations are difficult to obtain, especially at the very low water contents typical of arid areas. Extension of the theories to the low water contents, depths, thermal regimes and time spans that characterize the hydrology of arid site vadose zone repositories requires additional research.

Fig. 3.1—Water in an unsaturated soil is subject to capillarity and adsorption, which combine to produce a matrix suction. (From Hillel, 1971)

Fig. 3.2—A model illustrating unsaturated flow (under a negative potential gradient) in a horizontal column. (From Hillel, 1971)

Fig. 3.3— Representative water retention curves showing hysteresis. Note that for sand, the water content drops significantly as the negative soil-water potential (suction) increases (from Yong and Warkentin, 1975).

Fig. 3.4- Plot of hydraulic conductivity as a function of negative soil water potential (suction). Note that as suction increases, hydraulic conductivity for sandy soils becomes extremely low, limiting soil moisture movement significantly (from Hillel, 1971).

Fig. 3.5 Water content vs. depth for two times during infiltration.
(Revised, from Yong and Warkentin, 1975)

Fig. 3.6 Soil suction during infiltration at different times for a sandy loam.
(From Yong and Warkentin, 1975)

Fig 3.7 Comparison of measured and calculated conductivity for Ida soil. (Kunze et al., 1968)

Fig 3.8 Relation between hydraulic conductivity and volumetric water content for Tubac gravelly sandy loam with and without stones. (Mahuy et al., 1975)

Fig 3.9 Variation of diffusion coefficients with moisture content for Grenville silt loam. (Gupta and Staple, 1964)

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4. GEOLOGY AND HYDROLOGY OF FOUR AREAS WITH A DEEP WATER TABLE

Several geologic environments include large areas with deep water tables. In general, two factors favor deep water tables: (1) an aquifer that is highly transmissive relative to the amount of recharge, and (2) a large vertical relief between the point(s) of aquifer discharge and the area of interest. This section is a summary of an investigation of groundwater systems within four geologic provinces in the western United States and shows why deep water tables occur in them, with specific reference to analyzing the stability of the water tables in future periods of increased precipitation.

4.1 Great Basin, Nevada

4.1.1 Introduction

Some of the deepest water tables in the United States are found in the southern Great Basin, where climatic and geologic factors are conducive to deep drainage of aquifers. The unsaturated zone is locally almost 1,000 m thick and consists of several different materials, each with various implications for siting a nuclear waste repository.

The Great Basin is a subsection of the Basin and Range physiographic province (Penneman, 1931). The topography of the Basin and Range province is typified by more or less northerly-trending, parallel-to-subparallel mountain ranges and intermontane valleys, or basins (Figure 4.1). The Great Basin subsection is usually defined as that portion of the province with interior drainage (i.e., no surface drainage to the sea).

Mifflin (1968) has utilized all available geologic and hydrologic information for most of the Great Basin (the state of Nevada) to delineate groundwater flow systems (Plate 4.1). He and Bateman and Hess (1978) have catalogued the information on each flow system or portion of a flow system. Over much of the state of Nevada (the northwestern two-thirds) groundwater flow systems are localized within topographic basins and the water table beneath valley floors is shallow. In the southern and eastern portions of the state, on the other hand, a given flow system may underlie several topographic basins, and the water table may be hundreds of meters deep in places. This section will examine the best known regional flow system, the Ash Meadows system, and the largest system, the White River system (Plate 4.1).

4.1.2 The Ash Meadows Groundwater Basin

4.1.2.1 Introduction

The Ash Meadows groundwater basin, named for the discharge point of the system, is the best known of the regional flow systems due to geologic and hydrologic investigations associated with the operation of the Nevada Test Site (NTS) located within it (Figure 4.2). The test site is a facility operated by the Department of Energy and is used to test nuclear explosives. Between 1951, when testing began, and June 1977, 472 announced tests were performed at the site, 388 of which were underground tests. Geologic mapping, boreholes, tunnels, and aquifer test wells have been used to determine the character of the geologic and hydrologic environment. Observation wells are used to monitor the migration of radio-nuclides from explosion sites.

Topography in the NTS area is fairly typical of the Basin and Range province (Figure 4.2). A series of more or less northerly-trending, parallel-to-sub-parallel mountain ranges and intermontane valleys provides most of the relief, except for a few mesas in the northwest corner of the test site. The elevation of both the ranges and the valley floors decrease to the south and west in the vicinity of NTS (Borg et al., 1976). Many of the valleys are undrained (topographically closed) and contain ephemeral lakes called playas.

The NTS is situated in the most arid part of Nevada, the driest state. Precipitation in the area increases with altitude. For instance, valley floors usually receive between 8 and 16 cm/yr, while the highest mountain ranges near NTS may receive in excess of 50 cm/yr, much of it as snow (Borg, et al., 1976). Evaporation, as measured from pans, averages 282 cm/yr, and, while evaporation from soils is certainly much less, it is reasonable to assume that, in the valleys, the potential annual evapotranspiration greatly exceeds annual precipitation (Borg et al., 1976). Precipitation may locally exceed evapotranspiration in the uplands, especially during winter months (Borg et al., 1976).

4.1.2.2 Geology

4.1.2.2.1 Introduction

The geology of the Basin and Range province is an important factor in the configuration of regional flow systems and in the occurrence of deep water tables within those systems. In this section, the stratigraphy and geologic structure of NTS will be discussed in the

context of their effects on the hydrologic system, and data will be presented on the hydrologic properties of the various rock units. Although there are some hydrologic data for each unit, data points are so sparse, and measurements so approximate, that reliable modeling is probably impossible.

4.1.2.2.2 Stratigraphy

The stratigraphy of the NTS area is described below and is summarized in Table 4.1. Table 4.2 gives some hydrologic parameters for rocks at NTS. The formation thicknesses presented are maximum composite dimensions. Normally, the units present at any one locality are much thinner.

Late Precambrian To Late Paleozoic Sedimentary Rocks - The oldest rocks occurring at NTS are late Precambrian to late Paleozoic sedimentary rocks. These sediments, consisting of limestones and dolomites (carbonate rocks) and shales and sandstones (clastic rocks), reach a maximum composite thickness of 12,000 m (Borg et al., 1976). This portion of the stratigraphy has been broadly divided into four units: an upper carbonate sequence (a maximum of 1,100 m thick), overlying an upper clastic sequence (a maximum of 2,400 m thick), a lower carbonate sequence (a maximum of 5,000 m thick), and a lower clastic sequence (a maximum of 2,800 m thick) (Borg et al., 1976; Winograd and Thordarson, 1975).

The upper and lower carbonate sequences perform hydrologically as aquifers. The carbonates, which have low interstitial permeability, have been greatly deformed, and flow through interconnected fractures is the dominant groundwater transport mechanism. The fractures are irregularly distributed and may channel flow along pathways approaching 20 km in length. Therefore, the carbonates cannot be regarded as hydrologically homogeneous (Winograd and Thordarson, 1975; Winograd and Pearson, 1976; Winograd, 1978). Fractures remain open to a depth of at least 1,300 m (Winograd and Thordarson, 1975; Winograd, 1978). Estimates of the transmissivity of the lower carbonate aquifer range from 16 m²/day to 485 m²/day and average 162 m²/day (Borg et al., 1976; Winograd and Thordarson, 1975). The upper carbonate aquifer is thought to have similar hydraulic properties (Borg et al., 1976).

Although the Paleozoic and Precambrian clastic strata were subjected to the same deformation

as the carbonate rocks, they lack a hydrologically effective fracture porosity because of the plastic deformation of the shales and siltstones. As a consequence, flow through the units is controlled by the low interstitial permeability, which ranges from 3×10^{-8} to 4×10^{-6} m/day in the lower clastic unit (Borg et al., 1976; Winograd and Thordarson, 1975). Transmissivity of the lower 600 m of this clastic unit is less than $7.5 \text{ m}^2/\text{day}$ (Borg et al., 1976; Winograd and Thordarson, 1975). An improvised test of the upper clastic unit suggests a transmissivity of less than $7.4 \text{ m}^2/\text{day}$ (Borg et al., 1976). These relatively impermeable rocks act as barriers to flow and are called aquitards.

Mesozoic Rocks - The only rocks of Mesozoic age are a few small granitic intrusions. These are of minimal hydrologic importance.

Tertiary Volcanic Rocks - Overlying the old sedimentary rocks are up to 4,000 m or more of Tertiary volcanic rocks. These rocks are mostly tuffs with subsidiary amounts of rhyolite and basalt, and were deposited from Oligocene to Pliocene time (Borg et al., 1976; Winograd and Thordarson, 1975). The tuffs range from completely-welded to non-welded; some have been extensively reworked by running water, and some have been zeolitized.

Approximately the upper quarter to half of the volcanic section, made up of lava flows, welded ash-flow tuff, and bedded tuff, performs as an aquifer (Borg et al., 1976; Winograd and Thordarson, 1975). Fracture flow is the dominant mode of groundwater transport in the lava flow and welded-tuff units, while interstitial flow is indicated in the bedded tuff (Borg et al., 1976). Fractures are cooling joints and openings along bedding planes. Transmissivity of the lava-flow aquifer has not been determined; that of the bedded tuff, as determined in one well, may be about $16 \text{ m}^2/\text{day}$ with a hydraulic conductivity of 3×10^{-3} to 1.6×10^{-1} m/day (as determined from 11 samples, Winograd and Thordarson, 1975). Three tests of the welded-tuff aquifer measured transmissivity ranging from 2.5 to more than $845 \text{ m}^2/\text{day}$ (Borg et al., 1976).

The lower one-half to three-fourths of the volcanic section is broadly classified as the tuff aquitard, although its hydrologic properties may locally range into those of an aquifer (Borg et al., 1976). The unit is characterized by unconnected fractures and low interstitial permeability, the latter due to zeolites and clays deposited in the pores. Groundwater flow is controlled by this low interstitial permeability, which ranges from 1.6×10^{-5} to 1.2×10^{-3} m/day in one lithology of the many that occur in the

aquitard (Borg et al., 1976; Winograd and Thordarson, 1975; see Table 4.2 for more details).

Pliocene To Recent Sediments -

The youngest geologic materials in the NTS area are alluvial and lake deposits, locally over 1,000 m thick, which fill the intermontane valleys and were deposited from Pliocene to Recent times (Borg et al., 1976). The deposits, washed into the valleys from surrounding mountains, are still slowly aggrading. These valley-fill sediments are made up of discontinuous lenses of material of widely varying properties. In general, the material becomes finer with increasing distance from the mountains. Where playas occur, in the topographically lowest parts of the valleys, a surficial deposit of silt, as much as 20 to 30 m thick, may exist (Borg et al., 1976). Caliche is irregularly distributed throughout the entire thickness of the valley-fill, and, where present, it poorly cements the alluvium (Borg et al., 1976; Winograd and Thordarson, 1975).

The valley-fill sediments act as an aquifer where they are saturated. Groundwater flow is interstitial, and the unit has a transmissivity of 30 to 164 m²/day (Borg et al., 1976; Winograd and Thordarson, 1975).

4.1.2.2.3 Structure

Two major periods of deformation are recognized in the Ash Meadows area. The older is a late Mesozoic episode of folding and low-angle faulting (thrusting) which affects only the old, pre-Tertiary rocks, and the younger is a Pliocene-Pleistocene episode of block-faulting (high-angle faulting) which is responsible for the present topography (Borg et al., 1976). Local deformation associated with the extrusion of the volcanic rocks occurred: several collapse calderas and at least one resurgent dome are recognized (Byers et al., 1976). The major episode of block faulting ended in the Pleistocene period, but minor movements still occur and residual stresses exist in the earth's crust, as evidenced by seismicity, recent fault-scarps, recently re-activated faults, and earth cracks in playa floors (Snyder, 1973; Byers, 1978; Claassen, 1978).

Figure 4.3 is a cross-section of a typical topographic basin. The Basin and Range topography is fault-controlled. Any of the rock units may crop out in the ranges, and the upland may or may not be covered by a thin soil or alluvium. Thick sections of alluvium fill the valleys. Importantly, the volcanic and the valley-fill aquifers are generally not continuous across the faults, while the carbonate aquifer is thick enough so that continuity is maintained.

4.1.2.3 Hydrology

4.1.2.3.1 Introduction

Regional groundwater flow in the NTS portion of the Ash Meadows groundwater system occurs in the carbonate aquifers, and is generally in a south-southwest direction toward Ash Meadows, where large springs discharge from the system (Figures 4.2 and 4.4). Figure 4.5 is a simplified, generalized cross-section of the system. The flow system outside of NTS is not as well documented as that portion within NTS. The subtleties of this groundwater system are important in understanding the reasons for, and controls on, the regional (deep) water table.

4.1.2.3.2 Groundwater Recharge

Groundwater recharge is controlled by topography in several ways. As discussed above, precipitation increases with increasing altitude. In addition, evapotranspiration decreases with altitude. Thus, the net amount of water available for recharge increases with increasing altitude. Also, the fractured tuffs and carbonates crop out in the highlands, or are covered by only a thin veneer of alluvium, enabling precipitation to enter the fractures directly and rapidly descend below the zone of evapotranspiration (Thordarson, 1965; Winograd and Pearson, 1976). Recharge is probably minimal under steep slopes covered with alluvium, as the water runs off before substantial infiltration can occur (Claassen, 1978). Recharge under valley floors, where thick accumulations of alluvium occur, is considered to be negligible because evapotranspiration greatly exceeds precipitation in the lowlands, and because infiltration in porous media, especially in the fine-grained sediments on the playa floors, is limited (Borg *et al.*, 1976). Confirmation of this assumption requires field investigation.

Principal recharge areas are considered to be the Spring Mountains, the Sheep Range, and the Belted Range (Figure 4.2) (Borg *et al.*, 1976; Winograd and Thordarson, 1975; Winograd and Pearson, 1976). Subsurface flow from valleys to the north and east of the Nevada Test Site transfers water from these distant ranges to the saturated zone underlying NTS (Borg *et al.*, 1976; Winograd and Friedman, 1972). Lesser highlands, such as Rainier Mesa may also represent recharge areas (Thordarson, 1965). Occasionally, water may infiltrate downward to the water table underneath outcrops of the carbonate in relatively low, dry mountain ranges, for instance, the Specter Range, after anomalous rainfall (Winograd, 1978).

Nuclear test-related features (i.e., explosion-cavity collapse craters, recently reactivated faults, and associated surface cracks) may be anomalous sources of recharge in the valleys. The collapse craters can divert intermittent surface waters into the rubble chimney (Borg et al., 1976; Byers, 1978; Claassen, 1978), permitting rapid infiltration of surface water below the zone of evapotranspiration. There are also deep cracks on the playa surfaces, unrelated to nuclear testing, which occur elsewhere in the Basin and Range province, and which are observed to take large quantities of water (Doty, 1978; Byers, 1978; Claassen, 1978).

4.1.2.3.3 Groundwater Discharge

The discharge point of the Ash Meadows groundwater basin is the fault-controlled spring line at Ash Meadows, 35 km southwest of NTS (Figures 4.2, 4.4, 4.5). Total discharge along the spring line is approximately 2.1×10^7 m³/yr (Borg et al., 1976).

An undetermined amount of groundwater may flow under and past Ash Meadows toward Death Valley. This quantity, although believed to be smaller than the discharge at Ash Meadows, is one of the major uncertainties involved in groundwater velocity estimates (Winograd, 1978).

4.1.2.3.4 Groundwater Flow

Hydraulic communication between intermontane basins is not established through the valley-fill and volcanic aquifers because they are generally discontinuous between Basin and Range fault blocks (Figure 4.3). Thus, flow in these units is principally downward to the carbonate aquifer, although there may be some lateral component of flow within a topographic basin. Groundwater movement through the volcanic rocks is rapid in the aquifer portion of the unit, but slow in the aquitard portion. The slow rate of movement in the aquitard portion, estimated to be 1.5×10^{-4} to 6×10^{-2} m/yr, is the controlling factor in recharge through the volcanic rocks to the carbonate aquifer (Winograd and Thordarson, 1975).

The valley-fill and volcanic aquifers are saturated only in the structurally deepest parts of the valleys (Borg et al., 1976), except locally in the highlands where the volcanic aquitard perches water. A perched water-body has been well-documented under Rainier Mesa, hundreds of meters above the regional water table and approximately 500 m below the top of the Mesa (Figure 4.6) (Thordarson, 1965). Water percolates to the perched-water zone rapidly because the fractured volcanic aquifer crops out at the surface, or is

covered by only a thin veneer of soil, and precipitation can enter the fractures and rapidly descent below the zone of evapotranspiration. Estimates of the velocity of downward percolation in the fractured vadose-zone material range from 25 to 400 m/yr based upon tritium age-dating of the perched water (Thordarson, 1965). These estimates are confirmed by the observation that the water level changes in response to short-term (seasonal) changes in climate (Claassen, 1978). Hydrologic data indicate that water in the perched zone slowly leaks to the regional water table below (Thordarson, 1965).

Groundwater reaching the carbonate aquifer can flow laterally because the continuity of the aquifer is generally maintained over a wide area (Figure 4.3). This continuity is best maintained within and between the valleys at NTS (i.e., Yucca, Frenchman and Jackass Flats) and the discharge point at Ash Meadows. Between these valleys and valleys peripheral to NTS (e.g., Indian Springs Valley, Emigrant Valley and others, Figure 4.2), such continuity is not well maintained; relatively impermeable clastic rocks are faulted against, and interrupt, the carbonate aquifer (Figure 4.7), and barriers to flow are developed. Thus, between the central valleys of NTS and Ash Meadows, the hydraulic gradient is low, commonly 0.2 to 4.0 m/km (Winograd and Thordarson, 1968). On the other hand, between the NTS valleys and the peripheral valleys, hydrologic gradients are much steeper, on the order of 30 to 250 m/km (Figure 4.7; Winograd and Thordarson, 1968). Underflow from the peripheral valleys presumably is slow through the clastic-rock hydrologic barriers to the saturated zone underneath the central valleys.

Estimates of lateral flow velocity in the carbonate aquifer, based on permeability, aquifer cross-section, and total flow volume estimates, range from 1.8 m/yr to 180 m/yr beneath Yucca Flat and from 180 m/yr to 18,000 m/yr beneath the Specter Range (Borg *et al.*, 1976; Winograd and Thordarson, 1975). Estimates based on a two-well tracer test range from 18 to 410 m/yr (Borg *et al.*, 1976; Claassen and Cordes, 1975). Carbon-14 dates of water in the aquifer range from 10,200 to 28,000 years and imply a flow velocity of about 2 m/yr beneath Yucca Flat (Grove *et al.*, 1969).

The complex structure imposed upon the carbonate aquifer controls local flow rates and directions in a number of ways. The prevalence of fracture flow in the aquifer means that groundwater flow directions locally may not be perpendicular to potentiometric gradients inferred by borehole data, and may be inherently unpredictable. The irregular distribution of fractures may lead to channels of faster-than-average flow within the aquifer. Such channeling is inferred by hydrologic and hydrochemical investigations (Winograd and Thordarson, 1975; Winograd and Pearson, 1976).

4.1.2.3.5 Groundwater Levels

The regional water table under Yucca and Frenchman Flats lies from 150 to almost 600 m below the surface (Borg et al., 1976). The interconnection of these with other topographic basins in the regional groundwater system results in a groundwater discharge (Ash Meadows spring line) topographically hundreds of meters below the floors of the flats. This factor, plus the extremely gentle regional potentiometric gradient and lack of hydraulic barriers to flow between the flats and Ash Meadows, sets up the favorable geometry for deep water tables (Figure 4.5).

On the other hand, the depth to water under the peripheral valleys, such as parts of Emigrant Valley and Indian Springs Valley, is much less, sometimes as little as 0 to 70 m (Figure 4.7; Winograd and Thordarson, 1968). Figure 4.7 shows that the shallow water tables are due to clastic barriers to groundwater flow between the peripheral and the central valleys, behind which barriers the water table is high.

It is impossible to generalize the configuration of the water table under the mountain ranges within the Ash Meadows groundwater system. Under some highlands where recharge occurs, the regional water table may mound or water may be perched in the volcanic rocks. Under other highland areas, underlain by clastic rocks, potentiometric gradients steeper than the land-surface gradient can result in a decrease of the depth to the water table (Figure 4.7a). In other cases, the water level may remain relatively constant, and topographic changes result in an increase of the depth to the water table (Figures 4.7 b, c).

4.1.2.3.6 The Unsaturated Zone

The unsaturated zone is thickest over wide areas under the intermontane valleys. Under Yucca Flats, the water table is as deep as 580 m (Borg et al., 1976). Under some highlands, the regional water table may be locally somewhat deeper. In the valleys, the unsaturated zone comprises great thicknesses of alluvium and, depending on local geologic structure, some volcanic rocks and carbonate rocks (Borg et al., 1976). Under the highlands, the unsaturated zone includes mostly the volcanic and sedimentary rocks, with perhaps a thin soil cover.

The material in the unsaturated zone has important consequences for determining or modeling unsaturated flow. For instance, there is at present no theory describing unsaturated flow in fractured media. Also, there are only minimal field data on infiltration in fractured media at NTS (see above). On the other hand, unsaturated flow in porous media is better understood, and is described in detail in Section 3.0.

A number of factors seem to make fractured media less desirable for a repository site than porous media. One is that available evidence indicates that infiltration through fractured media is considerably faster than through porous media (see above). Evidence also indicates that recharge from precipitation on porous media may not infiltrate more than a few meters below the surface, while precipitation falling on outcropping or thinly-covered fractured rocks may reach the water table hundreds of meters below (see above). In addition, the theory of unsaturated flow shows that water in unsaturated (fine-grained) porous media will not flow into a repository opening. On the other hand, in fractured media, where the fractures may be too wide to support a meniscus, water could flow into a mined cavity in the unsaturated zone.

4.1.2.4 Climatic Changes and Water Level Fluctuations

The principal question with respect to relying on the unsaturated zone as a barrier to radionuclide migration is whether the water table may rise, inundating the repository and eliminating the barrier. Thus, the effect of a change in climate upon the water table must be evaluated.

World climate has clearly fluctuated between relatively cold-moist and hot-dry extremes over the last 600,000 to 1,000,000 years in cycles tens of thousands of years long (Antevs, 1952). The Basin and Range province was never glaciated itself during the cold-moist periods, but was subjected to greater precipitation and cooler temperatures. During these times, called pluvial periods, lakes formed (or enlarged) in some topographically closed basins which are dry (or contain small lakes) today (Morrison, 1965; Flint, 1971). The last pluvial period ended roughly 10,000 years ago, after which warming to a period of maximum aridity 6,000 years ago occurred (Antevs, 1952). Since then, climate has grown slightly less arid (Antevs, 1952; Wells and Berger, 1967). Minor climatic cycles lasting from several years to several thousands of years are superimposed upon the gross fluctuations (Antevs, 1952).

Geologic, hydrologic, anthropologic and paleontologic evidence has been used to estimate climatic parameters in the Basin and Range province during the last pluvial, and different conclusions have been reached (Antevs, 1952; Wells and Berger, 1962; Wells and Jorgensen, 1964; Flint, 1971; Snyder and Langbein, 1962). However, it is clear that at no time did precipitation exceed evapotranspiration in the region as a whole (Snyder and Langbein, 1962), and no estimates of an increase in precipitation greater than 50% over present rates have been made (Winograd, 1978). Thus, while return to a pluvial climate is likely (within the next 20,000 years or so), if future pluvials are similar to past ones, the water available for recharge will increase slightly, but will still probably be limited.

There are two possible scenarios that might pertain under pluvial conditions. One is that the water table might simply rise. The other is that a groundwater system similar to those now existing in other Basin and Range valleys (besides Yucca and Frenchman Flats) could develop.

In other Basin and Range valleys, where the rocks below the valley-fill are even less transmissive than the volcanic aquitard at NTS, or where recharge within a topographic basin is much greater than within basins at NTS, an intrabasin flow regime different from the predominantly downward flow at NTS occurs. Some relatively small amount of water may leak downward out of the valley fill, but most groundwater circulates within the alluvium. Downward flow (recharge) near the mountains, lateral and downward flow toward the valley floor, and upward discharge at the floor near the center of the basin, are the salient features of such a circulation (Figure 4.8). The Great Salt Lake Basin is an example of this kind of system, where the great amount of recharge from the Wasatch Range is probably the significant factor in its occurrence (Winograd, 1978; Arnow, 1978; Bolke and Waddell, 1972). Establishment of this kind of system in Frenchman and Yucca Flats is unlikely, even in the event of future pluvial periods with rainfall of magnitude similar to those in the past, because the recharge possible from the small ranges surrounding the flats is small as compared to the present recharge from the Wasatch Range (Winograd, 1978).

Hydrologic factors pertinent to NTS indicate that little, if any, rise in the water table would be expected in the event of wetter climate. The extremely low gradient and relatively high transmissivity of the regional aquifer suggests that increased recharge would not raise the groundwater table very much. If flow increased to the degree

that the discharge capacity of the Ash Meadows fault line were exceeded, groundwater might "back up." However, there are numerous fault lines analogous to the one at Ash Meadows, only several tens of meters above the presently discharging area which could act as "safety valves" (Winograd, 1978). Apparently, the only way that the water table could rise hundreds of meters under NTS is if all the topographically low discharge and potential discharge sites were blocked, or if they were to rise tectonically hundreds of meters relative to NTS valley floors (Winograd, 1978). Reversal of tectonic movements that have occurred over the past several million years is unlikely, at least within a time frame important to nuclear waste disposal, and such reversals in the past are not indicated in the geologic record in Nevada (Byers, 1978).

The U. S. Geological Survey is pursuing a two-part program to look for evidence of higher water tables under NTS in the past. One recently started project is to examine the detailed interstitial mineralogy of material now above the water table (Winograd, 1978). The other is to search for old deposits of precipitates from springs (potential "safety valves") which may have discharged in the past up the hydraulic gradient from the presently-discharging fault zone. To date, none have been found (Winograd, 1978).

The deposits associated with existing springs at Ash Meadows do imply greater hydraulic heads and/or flow rates in the past. (Winograd and Thordarson, 1975; Winograd, 1978). The distribution of desert pupfish in spring discharge ponds implies surface water connections between the springs in the past, and, indeed, lake deposits do occur at Ash Meadows (Winograd, 1978).

4.1.2.5 Groundwater Modeling

Borg et al., (1976) describe the possibility of modeling saturated flow and radionuclide transport in the Ash Meadows groundwater system:

"One- and two-dimensional mathematical models exist that could be used for the prediction of transport of radionuclides in groundwater flow from sites of underground nuclear explosions. One-dimensional models are probably sufficient. Their use can account for hydrodynamic dispersion, radioactive decay, convective transport, and instantaneous-equilibrium and linear-isotherm types of reversible sorption of a radioactive solute in groundwater flow.

Both models require as input parameters known or estimated values of transmissivity, storage coefficient, hydraulic head, groundwater flow velocity, dispersivity, distribution coefficients, medium effective porosity, medium bulk density, initial source radionuclide concentrations in groundwater, radionuclide decay constants, and dimensions of the radioactive contaminant pulse. With the exception of radionuclide decay constants and medium bulk densities, values of the parameters are either lacking altogether or are known imperfectly. Thus, currently only crude predictions of radionuclide transport at NTS can be made. Reliable and convincing predictions await further evaluation of the parameters that control transport in the NTS area."

A coarse sampling net for monitoring radionuclide migration at NTS has been set up, but has not as yet detected any above-background concentrations of radionuclides (Borg *et al.*, 1976). However, greater-than-background tritium concentrations were accidentally encountered in water from two holes 465 and 361 m from the nearest explosion sites (Crow, 1976). Based upon these findings, it is assumed that tritium is the only radionuclide to migrate significant distances in the NTS environment, and crude calculations of tritium migration have been made (ERDA, 1977). In these calculations, minimum travel-time between Yucca Flat and Ash Meadows of 300 years and complete mixing of contaminated water with water entering the groundwater system from all sources down gradient from Yucca Flat are assumed. Results show that the concentration in water discharging at Ash Meadows (of tritium of a given original concentration in Yucca Flat) will be reduced by a factor of at least 10^9 . Radioactive decay is responsible by itself for a reduction factor of 10^7 . The 300-year travel-time is probably unrealistically short, as Carbon-14 dates of water emerging at Ash Meadows imply an average groundwater residence time in excess of 10,000 years (Grove *et al.*, 1969; Winograd and Pearson, 1976).

4.1.3 Other Regional Flow Systems - The White River Groundwater System

4.1.3.1 Introduction

The geologic environment present at NTS extends off-site. The distribution of Basin and Range topography is shown in Figure 4.1, of the thick carbonate rocks in Figure 4.9, and of the tuffs in Figure 4.10. In light of the above discussion of the Ash Meadows flow system, similar

regional groundwater basins with deep water tables should be expected within the carbonate rock province of Figure 4.9. Mifflin (1968), in delineating flow systems within Nevada, recognizes approximately 18 such regional basins (Plate 4.1). Bateman and Hess (1978) report in a hydrologic inventory of Nevada playas that the deepest water table under a playa in Nevada occurs slightly more than 975 m below the surface of Desert Lake, 55 miles north of Las Vegas (Figure 4.2). (N.B.- The eastern boundary of the Ash Meadows groundwater system is in dispute, and Desert Lake could be considered part of the Ash Meadows groundwater system.)

The largest of the regional flow systems is the White River system (Mifflin, 1968). Its maximum dimensions are approximately 380 km long by 110 km wide, and the system may include up to 13 intermontane valleys aligned in a north-south strip along the White River Wash in eastern Nevada (Plate 4.1, Fig. 4.11, Eakin, 1966). The Wash was occupied by a Pleistocene (pluvial) river, and forms an axial topographic low roughly down the center of the groundwater basin (Eakin, 1966). Since the regional land-surface gradient is to the south, the Wash is generally the topographically lowest point of any at or north of a given latitude (Eakin, 1966).

Considerably less research has been done on the White River system as compared with the Ash Meadows system.

4.1.3.2 Geology

The White River area is very similar to the Ash Meadows area topographically, stratigraphically, and structurally. Briefly, a 10,000 m-thick section of Paleozoic sediments, more than half of which are highly fractured carbonate rocks, underlies many hundred to a few thousand meters of Tertiary volcanic rocks. Few measurements of hydrologic properties of the rocks have been made, but the properties are probably similar to those of rocks at NTS (Eakin, 1966). Tertiary to Recent block faulting deformed the bedrock into a series of parallel-to-subparallel ranges and valleys trending roughly north-south. The valleys are filled with sediment, locally nearly 600 m thick, that was eroded from adjacent ranges (Eakin, 1966).

4.1.3.3 Hydrology

Interbasin flow is recognized in the White River area (Figure 4.11). It is expected because of the thick, highly transmissive carbonate aquifer that underlies an area of high regional relief, and is implied by water budget studies, hydrochemical data, observations of springs, and some borehole data (Eakin, 1966). As in the Ash Meadows system, recharge is confined mostly to highland regions, and discharge is confined to a series of relatively large, uniformly-discharging springs. In the White River system, these springs are located at topographically low outcrops of carbonate rock in the White River Wash. Regional flow occurs within the carbonate rocks, and relatively long regional flow paths are implied (Eakin, 1966). Subsurface controls on hydraulic gradients, local flow rates, and local flow directions analogous to those in the Ash Meadows system occur in the White River system. The interaction of normally low gradients, topographically low discharge points, and high regional relief leads to areas where the water table is locally as deep as 300 m or more below the land surface (Figure 4.12, Bateman and Hess, 1978).

4.1.3.4 Climate Changes

The climatic variations to be expected in the Great Basin are discussed in Section 4.1.2.4.

4.1.3.5 Groundwater Modeling

Modeling the saturated flow regime in the White River system has not been attempted.

4.1.4 Differences Between the White River Area and the Ash Meadows Groundwater Basin

The White River groundwater system is larger than the Ash Meadows system (20,000 km² vs 12,000 km²), and involves greater volumes of flowing groundwater (12.4 x 10⁷ m³/yr vs 2.1 x 10⁷ m³/yr) (Eakin, 1966; Borg et al., 1976). It is an elongate feature while Ash Meadows basin is more equidimensional (Plate 4.1, Figure 4.11). The size and shape of the White River system might presumably lead to longer groundwater flow paths (and subsurface residence times). However, this presumption must be carefully examined.

Regional groundwater flow in the Ash Meadows system is at an angle to basin and Range structures, and, while several springs discharge the system, all the springs are localized along one fault. All the water entering the head of the system must pass through the whole basin before discharging. Flow paths from Yucca Flat to the discharge point are on the order of 100 km.

Regional flow in the White River system is almost due south, down the length of the basin, and there is some lateral component of flow toward the axially-located Wash (Figure 4.11). This flow is roughly parallel to Basin and Range structures, so faults do not control discharge as they do in the Ash Meadows area. Rather, the locus of discharge is the Wash, where the carbonate aquifer crops out at topographically low positions. Thus, as shown in Figure 4.11 and Plate 4.1, discharge occurs at various points along almost the whole length of the basin, as does recharge. Determining the discharge point for water entering the system at any given point is difficult (and perhaps impossible). Flow paths are not as clearly definable as they are at Ash Meadows, and, while they could be longer than 50 km, in some instances they could be much shorter. The gentle regional hydraulic gradients mean that minor tectonic or climatic disturbances could significantly alter local flow directions, change the discharge point, and markedly alter groundwater residence time.

4.2 Snake River Plain, Idaho

4.2.1 Introduction

The vadose zone beneath the Snake River Plain is locally more than 300 m deep; however, a detailed literature review and discussions with hydrologists who have studied the area have failed to provide many details of soil moisture movement in the vadose zone.

A groundwater model which appears to adequately simulate the underlying aquifer assumes a steady downward infiltration rate of approximately 3 cm/yr through the vadose zone. This downward flow is believed to be largely through fractures in the basalt rocks. Mathematical treatment of infiltration has not yet been attempted.

In this section, the geology and hydrology are described for the Snake River Plain east of Hagerman, Idaho (Figure 4.13). The Snake River Plain, a geologically young lava plateau, is an eastward extension of the larger Columbia River flood basalt province (Figure 4.2).

The subject area is approximately 320 km long, 80 to 120 km wide, and extends northeastward from Hagerman (Figure 4.13). The plain is relatively flat, sloping gently from its northeast tip towards its western end. The flat topography is locally interrupted by buttes and volcanic features such as lava cones and domes. Mountains border the plain on the northwest and southeast flanks. The Snake River flows mostly along the southeast margin of the plain and forms the southwest boundary of the subject area where it turns north through the Hagerman Valley. Tributaries flow into the Snake River

from the south and southeast. (Most of the streams entering the plain from the north flow intermittently and do not reach the river, because they sink into the ground a short distance from the mountains and recharge the groundwater system.) The plain is underlain by the Snake River Plain aquifer, a thick section of highly permeable basalt, called the Snake River group which is a principal natural resource of the area.

Most of the Snake River Plain receives less than 25 cm of precipitation per year (Mundorff et al., 1964). For instance, the Idaho National Engineering Laboratory, located in the north central part of the plain, receives an average of approximately 21 cm/yr, mostly in the spring (ERDA, 1977). The highland margins of the plain receive as much as 50 cm of precipitation or more per year (Mundorff et al., 1964). Pan evaporation, measured at relatively few localities, is approximately 90 cm/yr (ERDA, 1977).

The surface water and groundwater of the Snake River Plain are widely used for irrigation and domestic purposes. The Snake River has been dammed and diverted in several places to irrigate a strip of land up to 24 km wide on either side of the river (Mundorff et al., 1964). In 1959, a year for which data are readily available, 8.6×10^9 m³ of water were diverted to irrigate slightly more than 4×10^5 hectares (one million acres) of land (Mundorff et al., 1964). Also in 1959, 9×10^4 hectares (225,000 acres) were irrigated wholly or in part by 10^9 m³ of pumped groundwater (Mundorff et al., 1964).

The extensive use of water on the Snake River Plain, dating from the late 1800's, has led to an accumulation of hydrologic data. Also, the establishment of the Idaho National Engineering Laboratory (INEL, formerly the National Reactor Testing Station, NRTS) in 1952 has led to a detailed investigation of the geology and hydrology near the site (Figure 4.13). Geologic and hydrologic investigations have been performed mostly in connection with the disposal of radioactive wastes generated by the facility. Wastes are disposed of by burial in shallow trenches (solid waste), seepage from surface ponds (liquid waste), and well injection to the Snake River Plain aquifer (liquid waste). In an effort to monitor and predict the distribution of radionuclides released to the subsurface, boreholes, observation wells, and aquifer test wells have been drilled. The resulting data have been used to formulate a finite-difference model of groundwater flow and radionuclide transport beneath INEL.

4.2.2 Geology

4.2.2.1 Introduction

The Snake River Plain is a geologic basin filled with young lava flows and sediments. The most recent episode of volcanism occurred about 1,600 years ago, at Craters of the Moon National Monument (Nace et al., 1975). The stratigraphy at INEL, which is typical of the whole plain, is summarized in Table 4.3.

4.2.2.2 Stratigraphy

Pre-Pleistocene Rocks - Paleozoic sedimentary rocks form the mountains surrounding the Snake River Plain, and presumably, they underlie the plain. The depth to the Paleozoic rocks under the plain is unknown; but is greater than 600 m in the central part (Nace et al., 1975). Similarly, Tertiary volcanic rocks, which crop out only at some buttes and at the margins of the plain, underlie the younger volcanic rocks. The older volcanic rocks are relatively impermeable, and most groundwater is apparently found above them (Nace et al., 1975).

Young Basalts and Associated Sediments - The Snake River group of basalts and associated sedimentary interbeds fill the structural depression to a depth of greater than 450 m (ERDA, 1977; Walker, 1964). The deepest test hole at INEL, about 457 m, did not penetrate the bottom of the unit (Nace et al., 1975). Most geologic information comes from depths less than 300 m, as most wells do not penetrate more than 30 m or so below the water table (Walker, 1964). However, a U. S. Geological Survey project currently in progress is scheduled to include drilling and sampling the entire Snake River group (Robertson, 1978).

A brief description of the deposition of the Snake River group is helpful in understanding the hydrology of the unit. Lava welled up from vents and spread many kilometers over gentle slopes. Individual lava flows range in thickness from 3 to 20 m and generally contain vesicles and vertical cooling fractures. The top of each flow is commonly a very irregular, porous surface, so the contact zone between successive flows is commonly very permeable.

Locally, lava flows are rubbly, and, where hot lava encountered water, pillow lavas occur. These deposits are commonly highly permeable. Among the larger depositional

features are hollow lava tubes, formed when still-molten lava drained from beneath the crust of a partly-solidified lava flow. These tubes may be as large as 16 m in diameter and many hundreds of meters long (Stearns et al., 1938).

Thousands of years elapsed between successive flows over an area, and during these periods, wind-blown and water-laid sediments were deposited on the lava surfaces, to be covered by later flows. The sedimentary interbeds are commonly thin and discontinuous, although two are known from deep drilling to be as thick as 100 m or more and to extend for many kilometers.

The Snake River originally followed a course down the center of the plain. Successive eruptions of lava pushed the river progressively southward by damming and filling ancestral canyons cut by the Snake River. The filled and buried canyons may be up to 160 m wide and 80 km long (Stearns et al., 1938). The material filling them is commonly rubbly and more permeable than the normal lavas of the Snake River Plain (Stearns et al., 1938, see below).

The ancestral Snake River originally received tributaries from both north and south (Stearns et al., 1938). However, in the course of volcanic activity, a lava mound, which is still topographically expressed today, was built up, forming a trough along the northwestern margin of the plain in the vicinity of INEL (Stearns et al., 1938; Walker, 1964). This mound cut off the tributaries flowing into the plain from the north, and water from these streams now travels to sinks and playas along the northern margin of INEL where it recharges the groundwater system or evaporates.

The dominant mode of groundwater flow in the basalt is along the megascopic voids described above. The eruption and cooling features giving rise to these voids are related to individual lava flow horizons, and, as a result, the horizontal permeability of the entire aquifer is greater than vertical permeability. The anisotropy is enhanced by fine-grained, relatively impervious sedimentary interbeds which tend to inhibit vertical groundwater flow. The voids in the basalt are irregularly distributed on a local scale, but the assumption of large-scale aquifer homogeneity appears to be valid.

Barraclough et al. (1976) report that the horizontal hydraulic conductivity of the basalt at INEL, determined from field tests is from 0.03 to 3 cm/sec. The vertical

hydraulic conductivity of cores of sedimentary material from interbeds, on the other hand, is 2×10^{-10} to 3×10^{-3} cm/sec (Barracough et al., 1976).

Mundorff et al. (1964) describe 33 field hydrologic tests performed on the aquifer. These tests do not provide results that are representative of the whole unit, as the wells are concentrated in a few areas of the plain and are not drilled very far into the aquifer. Transmissivity ranges from 10^4 to 10^6 m²/day (Barracough and Jensen, 1976) and varies both between flows and within individual flows. However, the variations are sufficiently small that over a large area, the assumption of aquifer homogeneity is valid for practical purposes. The larger the area of interest, the better the assumption (Mundorff et al., 1964; Robertson, 1974). The average transmissivity is about 5×10^4 m²/day. The average storage coefficient determined from 18 tests is approximately 0.04, indicating unconfined conditions, and the effective porosity is 5 to 10% (Mundorff et al., 1964; Robertson et al., 1974). The permeability of 24 core samples ranged from 1.9×10^{-8} to 4.3×10^{-5} cm/sec, and averaged 6.7×10^{-6} cm/sec (Mundorff et al., 1964). Total porosity of 83 core samples averaged about 18% (Mundorff et al., 1964).

Surficial Deposits - The basalt is bare or thinly-covered over 10 to 20% of the Snake River Plain (Mundorff, 1964). Over much of the plain, the covering of surficial deposits is only two to three meters deep and is mostly wind-deposited silty material called loess (Nace et al., 1975). Basalt locally protrudes through this covering where the surface undulates. Locally, the surficial deposits are water-laid and reach a thickness of as much as 100 to 130 m. The details of the sediments are described in Table 3. Since the present runoff from the mountains is insufficient to have deposited all the water-laid sediments, their deposition is believed to have occurred during the last glaciation of the mountains to the north (Walker, 1964). Current deposition is restricted mostly to stream channels and playas (Walker, 1964).

Very few measurements of the hydraulic conductivity of the surficial materials have been made because they are, in most locations, unsaturated. The properties are thought to be similar to those of the sedimentary interbeds reported above (Barracough et al., 1976). The infiltration rate of surficial water in playa sediments is about 10^{-5} cm/sec and in the coarser sediments of the river alluvium is about 10^{-3} cm/sec (Barracough et al., 1976b).

Infiltration beneath outcrops of the fractured basalt has not been measured.

4.2.2.3 Structure

The Snake River group and the surficial sediments have been minimally deformed. The minor faulting and warping that does occur does not significantly affect groundwater movement.

4.2.3 Hydrology

4.2.3.1 Introduction

The Snake River group is the only regional aquifer underlying the plain. Plate 4.2 shows the pattern of flow in the Snake River Plain aquifer, formulated from all available data. Groundwater flow is generally southwest, parallel to the Snake River. Discharge occurs where the Snake River turns to cross the flow paths and has eroded into the aquifer.

Surficial sediments generally have low permeability and probably reduce infiltration of rainfall by storing it for later evaporation. Few details are available concerning the hydrology of surficial deposits.

4.2.3.2 Recharge

The total annual recharge to the Snake River Plain aquifer is about $8 \times 10^9 \text{ m}^3/\text{yr}$ (Mundorff et al., 1964; ERDA, 1977). Of this amount, $1.9 \times 10^9 \text{ m}^3/\text{yr}$ discharges at various points in and near the American Falls Reservoir, and never reaches the major discharge area between Twin Falls and Bliss (Plate 4.2). Thus, net recharge to the aquifer is approximately $6.1 \times 10^9 \text{ m}^3/\text{yr}$ (Mundorff et al., 1964). Natural recharge, prior to irrigation was probably about $3.6 \times 10^9 \text{ m}^3/\text{yr}$ (Mundorff et al., 1964).

Recharge occurs by irrigation, under streams, and along the fringes of the plain where runoff from the mountains and underflow from tributary valleys enter the aquifer. About 10% of the recharge is estimated to be from rain falling on the plain and percolating downward through the vadose zone (Mundorff et al., 1964).

Several estimates have been made of recharge due to infiltration from rainfall, but data are generally lacking for this important parameter. Most investiga-

tions conclude that an average of 2 to 3 cm of water per year reach the water table over the whole plain (Mundorff et al., 1964). One study, in which the plain was divided into four areas, shows that small areas of higher than average precipitation contribute most of the recharge whereas over approximately three-fourths of the plain, less than 1.5 cm/yr reaches the aquifer (Mundorff et al., 1964). The results of the study are shown below:

Recharge To Snake Plain Aquifer

Area	Average Annual			
	Area	Precipitation	Recharge	Recharge
	(sq mi)	(in.)	(in.)	(acre-ft)
Central part of plain (precipitation generally less than 10 inches)	6,100	9	0.3	100,000
Craters of the Moon.	900	13	3.5	170,000
Big Bend Ridge area.	400	18	6.0	130,000
South Side of plain (Lake Walcott-American Falls area and northeast along southeast side of the Snake River).	<u>1,000</u>	<u>12</u>	<u>2.0</u>	<u>100,000</u>
Total	8,400	--	---	500,000

As discussed in Section 3.0, the surficial material is important in determining the infiltration of precipitation. Where more than several meters of fine-grained sediments cover the basalt, net infiltration may be zero; where basalt or coarse alluvium forms the surface, 15 cm of water may reach the water table.

4.2.3.3 Discharge

The Snake River Plain aquifer discharges principally along the Snake River where it turns north and cuts across the aquifer (Figure 4.13 and Plate 4.2). Discharge there is now 6.1×10^9 m³/yr; before the advent of irrigation, it probably was about 3.6×10^9 m³/yr (Mundorff et al., 1964). The Snake River has eroded a canyon 15 to 30 m below the base of the aquifer in the discharge area, and large springs emerge from the northeast canyon wall (Mundorff et al., 1964). Many of the springs are localized in spring alcoves (Stearns et al., 1938) (Figure 4.14). The alcoves result from the cutting, by the present course of the river, of buried ancestral canyons, filled with permeable basalt. These ancestral canyons may be conduits for groundwater flow over the whole plain as well as loci for

discharge, but their occurrence and extent away from the modern canyon have not been determined.

The long history of water use on the Snake River Plain has led to the accumulation of discharge records of major springs and gage records of the Snake River dating back to the early 1900's (Mundorff et al., 1964). Since the aquifer discharges only to the river, and since river flow rates can be measured relatively accurately, the aquifer discharge data are believed to be accurate to $\pm 5\%$. Recharge estimates generally are based on measured discharge rates (Mundorff et al., 1964).

4.2.3.4 Groundwater Flow

Plate 4.2 is a generalized groundwater flow net for the Snake River Plain aquifer. Many investigators feel that most groundwater circulation occurs within the upper 160 m of the saturated zone (ERDA, 1977; Robertson et al., 1974).

Groundwater flows in a southwest direction with flow velocities that average about 5×10^{-3} cm/sec under INEL (Barraclough et al., 1967a; Robertson, 1974). All water, except for that originating from a strip of land along the river above Minidoka Dam, discharges at springs between Twin Falls and Bliss (Plate 4.2). Thus, water originating under INEL flows down the whole length of the plain, a minimum path length of 160 km, and takes 60 to 300 years to reach the discharge area. Reliable tritium age dating of the discharging groundwater is not possible because of the numerous and widely dispersed sources of recharge on the plain.

When irrigation is started each year, the increased recharge under the particular project establishes a groundwater mound in the aquifer which travels at approximately 0.8 km/day and which is observed up to 80 km or more from the irrigation site (Barraclough et al., 1967a; Thomas, 1978).

The hydraulic gradient in the aquifer is small, averaging about 0.002 under the entire plain (Barraclough and Jensen, 1976). Under INEL, the average gradient is 0.001, and in places is as small as 0.0001 (ERDA, 1977; Barraclough and Jensen, 1976). Gradients are locally steep where tributary valleys enter the plain. Water flowing into the plain cascades where it leaves valley-fill sediments and enters the more permeable Snake River basalts (Barraclough et al., 1976a).

4.2.3.5 The Unsaturated Zone

The Snake River Plain aquifer is almost everywhere unconfined, although slight artesian heads are found where sedimentary interbeds locally confine portions of the aquifer (Stearns et al., 1938). Figure 4.15 shows the depth to the water table under INEL and vicinity. The unsaturated zone is thickest just to the southwest of INEL, where the water table is over 300 m deep. However, data from that area are so sparse that the extent of the thick vadose zone cannot be delineated.

The material in the vadose zone is mostly basalt, sedimentary interbeds, and surficial sediments, where they are present.

Unsaturated-zone phenomena that could affect waste disposal operations include (a) infiltration of precipitation through the vadose zone, (b) aquifer responses to atmospheric pressure changes and (c) perched water. The role of certain characteristics of soil and rock in controlling infiltration is discussed in Section 3 of this report. Section 4.2.3.2 describes groundwater recharge mechanisms and gives estimates of the amount of water percolating through the unsaturated zone.

Atmospheric pressure changes cause two effects in the aquifer: the water table may fluctuate as much as a foot in response to pressure changes (Mundorff et al., 1964); and the aquifer is observed to "breathe" or expel moist air with decreases in pressure. This latter effect has an unknown influence on the mechanism of unsaturated flow in the vadose zone.

The interbedded and surficial sedimentary deposits can intercept and perch water infiltrating from the surface. Perched water bodies occur underneath areas with a substantial supply of water, (e.g., irrigation tracts, terminal sinks of rivers, and seepage ponds for the disposal of liquid wastes at INEL). Mud lake (Plate 4.2) is a body of water perched on surficial lake beds (Stearns et al., 1939).

The perched water bodies fluctuate in size in response to the short-term supply of water. For instance, four days after the 1965-1966 flood of the Big Lost River, the water level rose in a well tapping perched water 52 m below the surface and 69 m to one side of the river (Barracough et al., 1967b). Approximately four months after affecting the perched water body, the increased recharge was noticed in the regional aquifer 40 to 50 m below (Barracough et al., 1967a).

Under the liquid waste seepage ponds at INEL, two perched water bodies have formed. One is perched on top of the basalt at a depth of about 16 m and the other is perched on a sedimentary interbed at a depth of about 50 m. The regional water table in the area is about 140 m deep (ERDA, 1977). The laboratory monitors the dimensions of the perched water bodies and radionuclide concentrations in both the perched water and regional aquifer below (see, for example, Barraclough et al., 1967a, b; Robertson et al., 1974; Morris et al., 1963, 1965).

4.2.4 Climate Change and Groundwater Levels

The normal seasonal fluctuation of water levels is 0.6 to 1.5 m (ERDA, 1977). The fluctuations are greatest near irrigation projects where water is seasonally applied, and diminish with distance from a project. Long-term changes in water levels have occurred locally due to continued irrigation or pumping. The configuration of the water table prior to extensive irrigation is known at only a few places, and where that information exists the water table is observed to have risen about 20 m since irrigation began (Mundorff et al., 1964).

The possibility of a significant rise in groundwater levels due to climatic change is difficult to assess. There is little evidence on the plain indicating past climatic parameters or the amount of recharge from the surrounding mountains during past periods of wetter climate. There are no immediately obvious geological "safety valves," as there are in the Ash Meadows groundwater system discussed in Section 4.1.2.4, so that any rise of the water table far from the current discharge point is not likely to be attenuated. Analysis of the hydrologic data associated with specific irrigation projects might give some insight on the possible behavior of the aquifer under different climatic regimes. For a first approximation, it might be assumed that a rainfall rate double that of the present climate would double the flow rate through the aquifer. If the transmissivity were assumed constant, the gradient would have to double according to Darcy's Law. If the gradient were to increase from .002 to .004 over a flow path of 160 km, the water table would rise approximately several hundred meters, virtually eliminating the unsaturated zone at INEL.

Flow nets (Mundorff et al., 1964; Skibitzke and da Costa, 1962), and a finite-difference model have been formulated for the Snake River Plain aquifer. The first two kinds of models have been used for the whole plain. Finite-difference modeling has been used at INEL in an effort to understand and predict the migration of radionuclides.

Nearly 30 years of waste disposal at INEL has led to an accumulation of data on the distribution of radionuclides disposed of by burial, seepage ponds, or well injection. The finite-difference model is two-dimensional and accounts for convective transport, flow divergence, two-dimensional hydraulic dispersion, radioactive decay, and sorption for radionuclides migrating from the seepage ponds and injection well (Robertson, 1974). Currently, the radioactive waste plume down-gradient from the injection well includes tritium, strontium, and minor amounts of iodine 129 and plutonium (ERDA, 1977). None of the many other radioactive species has been detected even in the closest monitoring wells (Robertson, 1974; ERDA, 1977). Tritium has migrated 9 km down-gradient, and strontium only 2.4 km (Figures 4.16, 4.17; Robertson, 1974). Apparently, as deduced from the shape of the plumes, horizontal dispersion is nearly 90° (Robertson, 1974). Vertical dispersivity is assumed negligible, based on a small amount of data and the stratified nature of the aquifer (Robertson, 1974). Only tritium is found in the regional water table below the seepage ponds although many other species have been released to the ponds (Robertson, 1974). The waste plumes and changes in the groundwater gradients due to infiltration from the flooding Big Lost River, are well-simulated by the model (Robertson, 1974, 1978). Few new data have been acquired since formulation of the model, so tests of its predictive power have yet to be made (Figure 4.18; Robertson, 1978).

4.3 Columbia River Plateau, Washington

4.3.1 Introduction

The Columbia River Plateau, like the Snake River Plain, is underlain by a thick sequence of flood basalts, although it is older and more extensively covered by sediments than the Snake River Plain. The national laboratory at the Hanford Reservation, like INEL on the Snake River Plain has collected much of the available geologic and hydrologic information about the plateau. The water table at Hanford, located at a depth of about 100 m, probably is not deep enough to be suitable for a vadose-zone repository. However, the area is of interest because thicker vadose zones may occur elsewhere on the Columbia Plateau, and because Hanford is the site of extensive investigation of unsaturated flow and of deep infiltration of precipitation (Isaacson et al., 1974; Brownell et al., 1975; Last et al., 1976). Liquid and solid nuclear wastes are disposed of in the vadose zone, and their subsurface distribution is monitored. Modeling techniques have been applied to explain and predict the transport of radionuclides.

The Hanford Reservation (hereinafter termed Hanford) is a Department of Energy facility used since 1944 for the production and purification of plutonium. The geology and hydrology of the area is well-summarized by the National Academy of Sciences (1978):

"Geologically, the Hanford area is part of the Pasco Basin (see Figure 4.19, a broad area of low relief bordered by hills of basalt (of which Rattlesnake Hills, Yakima Ridge, and Umtanum Ridge are representative). The basalt is part of the huge accumulation of lava flows called the Columbia River basalt, which covers much of eastern Washington and Oregon. Lava flows underlie not only the hills but the whole of the Pasco Basin; the total thickness is unknown, but a deep well has shown it to be more than 3,000 meters. Overlying the basalt in the Basin is a sequence of sediments ranging in grain size from clay to gravel called the Ringold formation. Above the Ringold are local deposits of wind-blown silt (Palouse formation), then sands and gravels deposited by melt-water from glaciers of the Pleistocene Epoch. The glaciofluvial deposits form most of the surface of the Reservation, except for a 5-kilometer-wide belt of wind-blown sand. Groundwater moves fairly readily through much of the Ringold and glaciofluvial material, and the water table (the upper surface of the zone in the sediments that is saturated with water) stands far below the land surface over much of the Reservation. In most places, the water table is not far from the uneven contact between the glaciofluvial material and the Ringold sediments, and the thickness of the unsaturated (vadose) zone ranges up to 100 meters (emphasis added). Production and waste-management operations, especially the percolation of large volumes of low level waste water, have caused the water table to rise under parts of the Reservation, at places more than 25 meters. In the basalt beneath the Ringold formation, most aquifers are confined because some of the basalt flows are relatively impermeable."

The Hanford area is in the rainshadow of the Cascade Range, and receives an average annual precipitation of only 16 cm, about 40% of which falls in winter. Precipitation increases with altitude, and the surrounding hills receive up to 38 cm/yr (ERDA, 1975).

4.3.2 Geology

4.3.2.1 Stratigraphy

The stratigraphy of the Hanford area is summarized in Table 4.4 and the subsurface distribution of the stratigraphic units in Figure 4.20.

The Columbia River Basalt Group - The Columbia Plateau is underlain by more than 3,000 m of basalt flows (ERDA, 1975). The flows were extruded from fissures and were unaccompanied by violent explosion. Flows exposed at the surface are three to 45 m thick and some extend for tens of kilometers (National Academy of Sciences, 1978). Flows as thick as 120 m have been confirmed by drilling (National Academy of Sciences, 1978). The top of each flow may be slightly-to-moderately vesicular, but the vesicles are mostly unconnected and do not contribute to effective porosity. The upper and lower portions of each flow commonly contain columnar cooling fractures, while the central portions of thick flows generally are only slightly fractured. These fractures, while open in basalts near the surface, appear to be closed at depth (National Academy of Sciences, 1978). The basalt is most permeable at the contact between successive flows, where rubbly zones are commonly developed.

Interbeds of sediments separate the basalt flows and become increasingly thick upwards in the section. These sediments include clay, silt, sand, gravel, and some volcanic debris (National Academy of Sciences, 1978).

Few hydrologic tests have been performed on the Columbia River basalt near Hanford. One well was used to investigate the variation of hydraulic conductivity of the basalt with depth, and 28 values were obtained from injection and swabbing tests (National Academy of Sciences, 1978). The most permeable zones were weathered or highly-fractured flows and the rubbly contact zones between flows. Hydraulic conductivity in this material is of the order of 10^{-3} cm/sec (LaSala and Doty, 1971). The vesicular, unjointed basalt has a hydraulic conductivity of about 10^{-6} cm/sec and the dense basalt in deep zones where fractures are apparently healed also has a hydraulic conductivity of about 10^{-6} cm/sec (LaSala and Doty, 1971). The sedimentary interbeds, generally fine-grained, are commonly less permeable than the zones of fractured basalt. Some interbeds may be very permeable, however; one coarse-grained sedimentary zone about 30 m thick had a hydraulic conductivity of 10^{-3} cm/sec (ERDA, 1977; LaSala and Doty, 1971).

A summary of the data obtained from several other wells appears in Table 4.5. The wells are among the approximately 90 that penetrate into the Columbia River basalt. Most of the wells are drilled only into the upper portion of the unit. Table 4.5 shows that while the hydraulic conductivity of individual basalt layers can be low, the transmissivity of a thick section containing several basalt flows can be high.

The effective porosity of the unit is unknown because no tracer tests have been performed. From well-logging measurements, total porosity has been determined to be between 1 and 10% (ERDA, 1975).

Surficial Sedimentary Deposits - A locally thick sequence of sedimentary material has been deposited on top of the Columbia River basalts. The two principal units are the Ringold formation and the glaciofluvial and fluvial deposits (Table 4.4).

The Ringold formation has been arbitrarily divided at its type locality into a lower clay member, a gravel or conglomerate member and an upper sand-silt member; but this scheme is not applicable over the whole Pasco basin, because, as is typical for glacial deposits, the various sediment types are irregularly distributed. The Ringold formation is described in more detail in Table 4.4.

The hydraulic conductivity of the Ringold formation, determined by pumping tests, ranges from about 3×10^{-4} to 7×10^{-2} cm/sec (ERDA, 1975). More data on the properties of the Ringold formation are presented in Table 4.6.

Above the Ringold formation lie the glaciofluvial and fluvial sediments, predominantly coarse-grained material deposited by a series of floods originating from ice-dammed lakes upstream during Pleistocene time. Depositional features include flood-scoured, gravel-filled channels buried by later deposits which serve as conduits for groundwater flow (ERDA, 1975). These sediments are the most permeable deposits at Hanford, having a hydraulic conductivity of about 1 cm/sec. The Ringold and glaciofluvial sediments are collectively termed the "unconfined aquifer," and the results of aquifer tests are presented in Table 4.6. The calculated distribution of hydraulic conductivity derived from existing data is shown in Figure 4.21.

Other sedimentary units which play a lesser role in the movement of groundwater include the Palouse soils, fine-grained, wind-blown sediments locally lying between the Ringold formation and the glaciofluvial deposits, the Touchet fine-grained lakebed deposits, and volcanic ash beds. These lie above the water table and do not influence the saturated-flow regime directly.

4.3.2.2 Structure

Since their eruption, the basalt flows have been gently deformed into a series of anticlines and synclines which form the framework of the present topography. The Ringold sediments filling the Pasco Basin are only mildly warped, and the younger sediments have barely been deformed (ERDA, 1975). Faulting has been minimal, and, in general, geologic structure has a minor effect on groundwater movement.

4.3.3 Hydrology

4.3.3.1 Introduction

There are two groundwater systems recognized at the Hanford Reservation, an unconfined system and a lower confined system. The unconfined system comprises most of the surficial sediments and is well-known, while the confined system comprises the basalts and locally the lower Ringold formation, and is less well-known.

4.3.3.2 The Groundwater System

An unconfined groundwater system occurs in the Ringold and younger sediments. Runoff from the hills surrounding the Pasco basin and infiltration from rivers recharge the aquifer. Precipitation falling directly on the thick sediments may evaporate before infiltrating to the water table (Isaacson et al., 1974; Brownell et al., 1975).

Artificial recharge is supplied by waste disposal operations at the Hanford facilities, including infiltration of liquids from cribs, trenches, and seepage ponds. These operations have substantially altered the groundwater regime; Figure 4.22, a reconstruction of the water table before artificial recharged, shows that the table sloped gently and fairly uniformly towards the Columbia River with an average gradient of about 0.001. Figure 4.23, a map of the water table in 1975, shows that two mounds have built up beneath disposal areas, 23 m and 6 m above the natural position of the water table. Groundwater gradients have increased, and groundwater has risen in wells as far as 26 km from the disposal operations (ERDA, 1975).

Figure 4.24 shows groundwater flow paths for liquid originating at the disposal sites. The velocity of groundwater flow averages about 50 m/day (ERDA, 1975; Newcomb et al., 1972).

Monitoring wells have been used to trace the migration from disposal sites of radionuclides in the unconfined aquifer (see Figure 4.25). Some of the flow from cribs and ponds near the Columbia River discharges to the river 8 to 240 m away within three to eight days (ERDA, 1975). Up to at least 1973, dilution at the Columbia River was relied upon to lower the concentration of some radionuclides from one disposal area to permissible levels (ERDA, 1975).

The confined groundwater system at Hanford includes the Columbia River basalt group and the lower Ringold formation where clay beds act as aquitards. This system is, in general, poorly known, but each permeable zone is recognized

as a discrete leaky aquifer (National Academy of Sciences, 1978). A map of the piezometric levels in the uppermost confined aquifer appears in Figure 4.26. The horizontal component of the gradient over most of the Reservation is about 0.0002 to 0.001.

Recharge to the confined groundwater system is derived from precipitation on the highland regions of the terrain (National Academy of Sciences, 1978), especially where the basalt crops out. The attitude of the basalt affects the recharge of the system; since the basalt aquifers are layered between impermeable zones, flat-lying basalt sections experience slow recharge while tilted and broken basalt sections on anticlinal ridges expose the permeable zones for rapid recharge (Newcomb et al., 1972; Newcomb, 1961, 1969). Recharge is also derived from irrigation on lands to the east and north of the Hanford Reservation (National Academy of Sciences, 1978). The groundwater discharges in a zone beneath the generalized course of the Columbia River (National Academy of Sciences, 1978).

Geotechnical evidence shows that: (1) water currently in the confined aquifer beneath the Hanford Reservation probably entered the aquifer during a climate colder than at present, and (2) water from one zone entered the aquifer about 13,000 years ago. Thus, water movement in this zone appears to be slow.

Over most of the Hanford Reservation, the piezometric level of the confined aquifer is higher than the water table by 0.3 to 10 m (NAS, 1978). There are few measurements of the variation of head with depth in the confined aquifer, but those that exist suggest an upward gradient several times the magnitude of the horizontal gradient (NAS, 1978).

4.3.3.3 Groundwater Models

The history of groundwater modeling at Hanford is described in ERDA (1975):

"Initially, simple graphical flow net analyses and analog models were applied and evaluated. In the early 1960's a program was initiated to develop the capability to predict transport of contaminants in Hanford sediments. Early models developed to predict groundwater movement under steady state conditions required many compromising assumptions to simulate the system. The models were not able to simulate transient conditions or waste transport and thus were not effective tools for managing a groundwater basin. In the late

1960's an intensive effort was initiated to develop a more refined model that would provide transient analysis of contaminant movement.

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"To date, models for describing groundwater and contaminant movement in the unsaturated zone and the unconfined aquifer have been developed and tested against analytical solutions and limited field data. The testing provided confidence in the mathematical methods and the approach being used. Verification with field data showed areas where data coverage was inadequate and pointed up the need for development of a technique to compute uncertainties in the output based on known uncertainties in the input data. Additional data and work are needed to reduce the uncertainty in the model predictions.

"In general, the Hanford Reservation modeling is divided into three categories: 1) unsaturated-zone analysis capability, 2) saturated unconfined groundwater analysis capability, and 3) transport capability. The unsaturated and saturated portions of the groundwater system were considered separately in order to develop feasible models. The transport phenomena were assumed to be affected by, but not to affect, the convective flow patterns.

"The model development effort is separated into three major categories: data models, hydraulic models, and contaminant transport models. Data models calculate from a relatively small number of field measurements the input characteristics required for operation of the hydraulic and transport models. The hydraulic models predict the groundwater velocities in the unsaturated and saturated sediments. The transport model combines the groundwater movement calculated by the hydraulic models with soil-waste reactions to predict the contaminant movement rate. The groundwater convective transport is simulated independent of waste movement.... A major data model, the TIR, has been developed for calculating the hydraulic conductivity distribution in highly heterogeneous aquifers where characterization by field measurement alone would be prohibitive. The method is based on the numerical integration of the Boussinesq equation for the transmissivity along instantaneous streamtubes of flow. Testing of the computer program on a synthetic surface identified a set of control parameters that resulted in a maximum computation error of + 5%.... (Details of the model assumption limitations, error, and sensitivity analyses are contained in Garlock et al., 1974.)

"The basic hydraulic models are the Variable Thickness Transient Groundwater Model (VTT) and the Partially-Saturated Transient Flow Model (PST). The VTT Model was developed to predict the changing height of the water table (phreatic surface) throughout the unconfined aquifer underlying the Han-

ford Reservation. It provides for the simulation of two-dimensional flow in unconfined aquifers. The model utilizes the nonlinear transient Boussinesq equation with the appropriate initial and boundary conditions. The heterogeneous permeability distribution input can be calculated by the TIR. A successive line over-relaxation technique with unequal time steps is used for numerical solution.

"The PST Model includes a mathematical description and computer program to simulate transient Darcian transport of a fluid in heterogeneous partially-saturated porous media (Reisenauer, 1972). The objective of the model is to assess movement of groundwater in the unsaturated zone. The model involves the solution of the basic two-dimensional equation of flow in the unsaturated region (Reisenauer, 1972).

"A transport model (Baca et al., 1973; Ahlstrom and Baca, 1974) has been developed for use with the PST and VTT Models for predicting movement and distribution of contaminants in groundwater systems. Based on the two-dimensional form of the general diffusion-convection equation, the model considers variable dispersion coefficients, variable media thickness and sink and/or source terms. The model has two major components: the macroion segment handles dissolved minerals that are typically present in groundwater systems, and the microion segment predicts movement of contaminants which are normally present in trace quantities (in comparison to the naturally-occurring materials). The model encompasses the compound problem of multi-component transport with simultaneous chemical reactions and includes the general transport mechanisms of advection, dispersion, and sorption, as well as radioactive decay. Each segment can be run as a separate model."

4.3.3.4 The Unsaturated Zone

The unsaturated, or vadose zone, reaches a maximum depth of about 100 m at the Hanford Reservation. It consists primarily of widely graded silts, sands, and gravels, although at some locations, basalt lies within the unsaturated zone (Figure 4.27).

Because of the potential consequences of radionuclide migration to the Columbia River, the unsaturated zone at Hanford is being studied intensively. Several years ago, two lysimeters (3 m-diameter, instrumented pipes buried 20 m into the soil, to monitor vertical movement of soil moisture) were constructed, and during the past five years the data have indicated that a large percentage, if not all, of the rainfall that infiltrates the upper 12 m of the soil eventually evaporates.

4.4 Copper Mine Area, Arizona

4.4.1 Introduction

Within the Colorado Plateau province of southwestern United States, deep water tables can occur where there is a combination of deep drainage by canyons, low recharge caused by the arid climate, and a thick section of relatively permeable rocks. The Copper Mine area (Figure 4.28), a small portion of the Colorado Plateau, demonstrates the profound influence that local geologic structure can have in maintaining deep, unsaturated zones in rock. The Navajo sandstone, widespread in the surrounding area, is in many places an excellent aquifer. In the Copper Mine area, however, the sandstone is sandwiched between rocks that do not readily transmit water, and it is located adjacent to and well above the Colorado River on the limb of an anticline (a convex upward fold). Drainage is directly into the Colorado River, and it is unlikely that the groundwater level could rise appreciably, because, as shown in the cross section in Figure 4.29, the elevation of the groundwater table is strongly controlled by the depth of the canyon of the Colorado River.

The Copper Mine area is located near the Utah border and lies within the Navajo Indian Reservation. The region is arid and is composed of elevated plateaus of roughly horizontal sedimentary rocks which have been deeply dissected by the Colorado River and its tributaries. While the area may not be a suitable repository site because of its location within the reservation, it provides an excellent example of how a thick, unsaturated zone of wide, horizontal extent can be maintained in rock under certain geologic and topographic conditions.

The climate of the Copper Mine area is influenced strongly by its altitude and the surrounding mountain barriers. The mean annual precipitation is about 20 cm and the mean annual temperature is about 12°C. Summer precipitation supplies 50 to 65 percent of the annual total and is sporadic, whereas winter precipitation is generally distributed evenly. Summer temperatures higher than 38°C (100°F) are common. Prevailing winds are southwesterly. The distribution and orientation of dunes and other eolian deposits indicate that the wind movement in the area during the past million or more years (at least during times of eolian deposition) was from the southwest and generally similar to the present wind pattern (Cooley et al., 1969).

4.4.2 Geology

4.4.2.1 Stratigraphy

The Copper Mine area is located on the Kaibito Plateau under which lies over 1,300 meters of sedimentary rocks (sandstones, shales and limestones) resting upon an igneous and metamorphic basement (Cooley, et al., 1969). A summary of the stratigraphy and certain characteristics of the stratigraphic units is presented in Table 4.7.

The sedimentary rocks have been classified according to their water supply capability. The term "multiple aquifer system" is used to describe one or more aquifers which are separated from other aquifers by aquicludes (rocks which do not transmit water). A specific multiple-aquifer system is designated by the initial letter of the major aquifer of the system.

The sedimentary rocks of the Kaibito Plateau have been divided into four groups which, from upper (younger) to lower (older) are described below.

1. The N multiple-aquifer system is named for the Navajo sandstone, the major aquifer of the system. It is capped over the northern part of the plateau by about 25 m of essentially impermeable red siltstones of the Carmel formation, which act as an umbrella to prevent precipitation from percolating into the N-multiple aquifer system. Sandstone beds form a minor part of the Carmel formation whereas the siltstone units range in thickness from 1.5 to 6 m (Harshbarger et al., 1957).

The main aquifer of the system is a 244 m thick section of Navajo sandstone. Although the Navajo sandstone is a regional aquifer, it is largely unsaturated in the Copper Mine area. It is composed of medium- to fine-grained sand and in most places is bonded with a weak calcareous cement. The formation is remarkably uniform throughout the Colorado Plateau region.

In its lower portions, the Navajo sandstone is interbedded with about 40 meters of the impermeable Kayenta formation and is isolated at the bottom from the underlying multiple aquifer system by about 500 meters of impermeable rocks, the silty sandstones and sandy siltstones of the Moenave, Chinle, and Moenkopi formations. In its typical occurrence, the Kayenta consists of a fine-grained sandstone interbedded with mudstone (Harshbarger, 1957). The sandy units of the Kayenta are con-

nected hydraulically with the Navajo sandstone and form part of the N multiple-aquifer whereas the mudstone units act as confining beds (McClymonds, 1961). The silty sandstones and sandy siltstones of the underlying Moenave formation are not waterbearing and are interbedded with the overlying Kayenta to form the bottom confining aquiclude for the N multiple-aquifer system. In this location, the Moenave would act to keep water from leaking upward into the unsaturated zone of the Navajo sandstone if the underlying aquifer were saturated.

2. The Moenkopi and Chinle formations are mainly aquicludes (Cooley et al., 1969). They comprise mudstones and siltstones which together are about 380 m thick as measured in sections 4 miles northwest of the Copper Mine area (Cooley et al., 1964). They act as a further barrier to prevent the transfer of water between multiple-aquifer systems.

3. The C multiple-aquifer system is named for the Coconino sandstone, a rock composed of well-cemented, wind-blown sand. Other members of the system which yield small amounts of water in this area are the Kaibab limestone and the upper part of the Supai formation. The Coconino sandstone is saturated under part of the Kaibito Plateau but is unsaturated in the Copper Mine area. The C multiple-aquifer system, including the relatively impermeable basal siltstone beds of the Supai formation, is about 300 m thick in the Copper Mine area. The siltstone beds form the lower confining bed for the C multiple-aquifer system. The closest outcrop of the Supai formation to the Copper Mine area is in the gorge of the Colorado River about 25 km to the west. These rocks lie below the level of the Colorado River in the region of the buried anticline at Copper Mine, and they do not affect directly the hydrology of the unsaturated zone at the Copper Mine area.

4. The Redwall limestone is an important aquifer yielding large amounts of water to the Little Colorado River at Blue Springs, approximately 53 km south of the Copper Mine area. It is too deeply buried to affect the area of this study and is mentioned only to complete the description of aquifers in the total stratigraphic column above the basement rocks.

4.4.2.2 Structure

The Kaibito Plateau is located in the south-central part of the Colorado Plateau province, which is characterized by nearly horizontal strata, high elevations and spectacular canyons (Penneman, 1931). As shown in Figure 4.28, the Copper Mine area lies on the limb of the elongate Preston Mesa anticline where it crosses the north part of the Kaibito Plateau (McClymonds, 1961).

The province has been generally stable for about 600 million years since late Precambrian time. It was warped into gentle folds about 70 million years ago during late Cretaceous time and was uplifted and faulted about 10 million years ago during late Tertiary and Quaternary time (Harshbarger et al., 1957).

The regional uplift set the stage for erosion, and rivers have dissected the area into an intricate pattern of canyons. To the east of the Kaibito Plateau, the Echo Cliffs and Marble Gorge of the Colorado River expose some 1,000 m of the stratigraphic sequence down to the Redwall limestone (Cooley et al., 1969).

The Preston Mesa-Mt. Beautiful anticline can be traced for over 240 kilometers to the southwest from the Copper Mine area (Cooley et al., 1969) (Figure 4.28). In the area of interest, the dips of the anticline and the associated synclines to the east and west are less than 3 degrees (McClymonds, 1961). These low dips exemplify the very gentle nature of the folding in the Colorado Plateau.

There are several faults at Mormon Ridge on the anticline crest south of the Copper Mine. The displacement on the faults does not exceed 15 meters (Harshbarger et al., 1957). In general, the faults in the Copper Mine area are too isolated and displacements are too small to influence groundwater flow (Cooley et al., 1969).

4.4.3 Hydrology

4.4.3.1 Groundwater Recharge

Recharge to aquifers in the area is from precipitation, ephemeral streams or interformational leakage of groundwater. Direct recharge is controlled by the rock permeability, geologic structure, characteristics of surficial deposits and by the duration, type and amount of precipitation (Cooley et al., 1969). As shown in Figure 4.29 the Navajo sandstone is capped for the most part by the relatively impermeable Carmel formation in the Copper Mine area. The cap extends for about 14 km on the surface over an interval where the elevation decreases about 275 m toward the north. Where it occurs, the Carmel retards greatly the infiltration of rainfall into the underlying Navajo sandstone, and what water does infiltrate the upper few meters of the Carmel formation is believed to be lost to the atmosphere by evapotranspiration.

In this area, it is unlikely that recharge would occur from an underlying aquifer since, as is shown in Figure 4.30, the C multiple-aquifer below the Navajo sandstone is not saturated in this area. Furthermore, even if the C multiple-aquifer system were saturated, it is separated from the Navajo formation by about 400 m of relatively impermeable rocks, the Chinle and Moenkopi formations.

4.4.3.2 Groundwater Flow and Discharge

Movement of groundwater in the Copper Mine area is toward the Colorado River (Cooley et al., 1969). Figure 4.30 shows the water level contours in the N multiple-aquifer system.

The average hydraulic gradient in the Navajo sandstone from the northern boundary of the Copper Mine area to the Colorado River, as determined from the water-level contours of Figure 4.30, is approximately 0.03. Based on field measurements, the transmissivity of the Navajo sandstone ranges from about 5 to 50 m^2/day (Cooley et al., 1969). These values of transmissivity and a hydraulic gradient of 0.03 indicate that the flow to the Colorado River per meter of width of the Navajo sandstone (where saturated) ranges from about 0.2 to 2 m^3/day . Groundwater flow does not occur directly in the study area, however.

The length of groundwater flow is approximately 15 km, and commonly discharge flows into the Colorado River from springs associated with rock fractures near the base of the Navajo sandstone.

As shown in Table 4.8 the permeability of the Navajo sandstone as measured from cores ranges from about 1.4×10^{-4} to 2.3×10^{-2} cm/sec. Assuming an effective porosity of 0.30 the flow velocity under a hydraulic gradient of 0.03 would have an approximate range of 4 to 700 m/yr. Therefore, the residence time of water in the aquifer, if it were to enter below the Copper Mine area, would be of the order of 20 to 4,000 years.

4.4.3.3 The Unsaturated Zone

To the north and northeast of the Copper Mine area, the Glen Canyon of the Colorado River and the tributary Navajo Canyon expose the Navajo sandstone and provide drainage down to the river elevations.

As shown in Figure 4.29, a zone of unsaturated Navajo sandstone extends approximately 10 km eastward from the Colorado River. The thickness of the unsaturated zone at the Copper Mine area is about 250 m. As shown in Figure 4.30, the zone of unsaturated rock has an areal extent, centered at the Copper Mine area, that is roughly circular with a diameter of about 25 km (an area of approximately 500 km²). No studies have been made of the flow of water in the unsaturated zone.

As can be seen from Figure 4.29, the groundwater level in the Navajo sandstone is strongly controlled by the elevation of the Colorado River, and a major change in the level of the river would be required to raise the water table significantly. Because there is no evidence that this deep drainage will be significantly altered by tectonic activity, the extensive unsaturated zone at the Copper Mine area will, in all probability, continue to exist for the design lifetime of a radioactive waste repository.

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PLATE 4.1. DELINEATION OF GROUND-WATER FLOW SYSTEMS.

CENTER FOR WATER RESOURCES RESEARCH
UNIVERSITY OF NEVADA

STATE OF NEVADA

Scale 1:1,000,000

DELINEATION OF GROUND-WATER FLOW SYSTEMS

Area of ground-water discharge

Area and direction of transfer of water from one ground-water system to another (may involve either surface or subsurface flow)

Approximate area of ground-water discharge

Direction movement of ground-water

Approximate area of significant moisture for recharge and runoff

Approximate position of system boundary

25

Index number of flow system

Uncertain flow system boundary

(AFTER WIFFLIN, 1967)

BASE MAP FROM U.S. GEOLOGICAL SURVEY, 1963 EDITION

WHITE RIVER
GROUNDWATER SYSTEM

NEVADA

UNITED STATES DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

PREPARED IN
U.S. BUREAU

PREPARED IN COOPERATION WITH THE
U.S. BUREAU OF RECLAMATION

WATER-SUPPLY PAPER 1846
PLATE I

2

GENERALIZED MAP SHOWING PHYSICAL AND
OF THE SNAKE RIVER PLAIN EAST

PLATE 4.2
(After Mundorff, 1967)

Hydrology modified from Mundorff, Crosthwaite, and Kilburn (1964)

ING PHYSICAL AND HYDROLOGIC FEATURES SNAKE RIVER PLAIN EAST OF BLISS, IDAHO

255-995 O - 67 (In pocket)

PLATE 4.2
Mundorff, 1967

4

TABLE 4.1 - SUMMARY OF STRATIGRAPHY AND HYDROLOGIC PARAMETERS OF GEOLOGIC MATERIALS AT NEVADA TEST SITE

Age		Hydrogeologic Unit		Maximum Thickness At NTS (m)	Lithology	Dominant Flow Characteristics	Hydraulic Characteristics
Cenozoic	Recent to Quaternary or Tertiary	Valley-fill Aquifer		1,160	Discontinuous interbeds of coarse to fine grained sediments including alluvial fan, lakebed, river, and mudflow deposits. Partial cemented with caliche and locally deeply cracked.	Interstitial flow, except locally fracture flow in such features as playa cracks.	Transmissivity: 30 to 164 m ² /day.
	Tertiary	Volcanic Aquifer	Lava-flow Aquifer	762	Lava flows, with vertical cooling fractures.	Fracture flow in macroscopic voids.	
			Welded Tuff Aquifer	635	Non-welded to completely welded ash-fall tuff; highly fractured in the middle of each flow.	Fracture flow.	Transmissivity: 2.5 to 850 m ² /day.
			Bedded Tuff Aquifer	106	Ash-fall tuff and tuff reworked by running water; soft, fine to coarse grained; no fractures.	Interstitial flow.	Transmissivity: 16 m ² /day Hydraulic conductivity: 3×10^{-3} to 1.6×10^{-1} m/day.
		Tuff Aquitard		1,300	Ash fall and ash flow tuffs with associated sediments; fractures unconnected; extensively zeolitized.	Interstitial flow.	Hydraulic conductivity: 10^{-3} to 10^{-4} m/day
Paleozoic	Permian to Pennsylvanian	Upper Carbonate Aquifer	1,100	Highly-fractured limestone.	Fracture flow.	Similar to lower carbonate aquifer.	
		Upper Clastic Aquitard	2,400	Conglomerate, argillite, and quartzite; fractures believed to be sealed by plastic deformation.	'Interstitial' flow.	Transmissivity: 0.01 to 0.1 m ² /day.	
	Devonian to Cambrian	Lower Carbonate Aquifer	5,000	Limestone, dolomite, minor clastic rocks; highly-fractured.	Fracture flow.	Low interstitial hydraulic conductivity: 4×10^{-4} m/day. Transmissivity of formation (includes fracture flow): 16 m ² /day to 500+ m ² /day.	
	Cambrian	Lower Clastic Aquitard	2,800	Quartzite, shale, siltstone, sandstone; fractures believed to be sealed by plastic deformation.	Interstitial flow.	Interstitial hydraulic conductivity: 3×10^{-8} to 4×10^{-6} m/day.	
Precambrian							

TABLE 4.2 - INTERSTITIAL POROSITY AND HYDRAULIC CONDUCTIVITY
OF CORES FROM ROCK UNITS AT NTS
(From Winograd and Thordarson, 1975)

	<u>Number of</u> <u>Samples</u>	<u>Range</u>	<u>Mean</u>
Tuff Aquitard (zeolitized tuff)			
Total porosity (%) ^a	34	19.8 - 48.3	37.7
Effective porosity (%) ^b	-	-	-
Hydraulic conductivity (m/day) ^c	34	2×10^{-6} - 2×10^{-2}	2×10^{-3}
Tuff Aquitard (argillaceous tuff and argillaceous sediments)			
Total porosity (%) ^a	-	-	-
Effective porosity (%) ^b	32	1.8 - 21.6	11.0
Hydraulic conductivity (m/day) ^c	38	8×10^{-8} - 2×10^{-2}	8×10^{-4}
Upper Clastic Aquitard			
Total porosity (%) ^a	22	2.0 - 18.3	7.6
Effective porosity (%) ^b	22	0.6 - 15.1	4.2
Hydraulic conductivity (m/day) ^c	-	-	-
Lower Carbonate Aquifer			
Total porosity (%) ^a	16	0.4 - 12.4	5.4
Effective porosity (%) ^b	25	0.0 - 9.0	2.3
Hydraulic conductivity (m/day) ^{c,d}	13	8×10^{-7} - 4×10^{-3}	4×10^{-4}
Lower Clastic Aquitard			
Total porosity (%) ^a	23	0.2 - 10.0	3.8
Effective porosity (%) ^b	20	0.6 - 5.0	1.9
Hydraulic conductivity (m/day) ^c	18	3×10^{-8} - 4×10^{-6}	4×10^{-7}

^a Determined from grain and bulk densities.

^b Determined by mercury injection or water saturation.

^c Conductivity to Denver, Colorado, tap water.

^d Most hydraulic conductivity in the carbonate aquifer is derived from fractures, not interstitial voids. Thus, these values understate formational hydraulic conductivity.

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TABLE 1.—Summary of principal characteristics and stratigraphic relations of geologic units

Geologic time			Lithologic unit	Thickness (ft)	Physical character	Areal distribution	Origin
Era	Period or system	Series					
Cenozoic	Quaternary	Holocene	Windblown deposits	0-30	Fine sand and silt	Mantles large areas in various parts of the NRTS	Loose transported from distant sources. Dune sand derived from local sources
			Playa deposits	0-10	White to light-tan clay, silt, and sand. Chiefly poorly sorted clayey, sandy silt. Some is quite calcareous	Covers about 22 square miles within the NRTS, in Big Lost River and Birch Creek playas and numerous scattered small playas	Sediments derived chiefly from local unconsolidated older deposits
			Slopewash and talus	0-25	Slopewash is chiefly reworked windblown sediments, minor amount of detrital local. Talus of West Twin Butte is blocky, black to brown, generally dense basalt. Talus of East Twin Butte consists of fragments of welded tuff, pumice, and miscellaneous volcanic rocks, ranging in size from sand grains to boulders	Slopewash occupies numerous depressions and sidelopes. Talus and slopewash occur together around the flanks of East Twin Butte and on the northern flank of West Twin Butte	Slopewash is derived from windblown materials, local products of weathering and erosion. Talus is derived from adjacent rock outcrops
			Alluvium of the Big Lost River	0-120	Consists chiefly of pebbles and rubble gravel with varying amounts of silt and sand matrix. Coarser in the southern part of the station, progressively finer grained northward, becoming fine sand and silt	Covers about 50 square miles within the NRTS, in a belt up to several miles wide on either side of the river	Derived principally from mountains in the headwaters area of the Big Lost River outside of the Station
			Alluvium of Birch Creek	0-32	Sandy pebble and cobble gravel, about 40 percent by weight of gravel and 20 percent sand and silt matrix. Particles are rounded and are similar in composition to those in the alluvium of the Big Lost River. Thirtysix foot thickness penetrated at test well 7N-31E-34dd1	Covers about 15 square miles in the extreme northern part of the NRTS, north of the Birch Creek playa	Derived chiefly from the drainage area of Birch Creek outside the NRTS, on the eastern slope of the Lemhi Range and the western slope of the Beaverhead Range
			Alluvial-tan deposits	0-10	Noncompacted boulder and rubble gravel in a matrix of fine materials ranging in size from clay to coarse sand	Covers about 16 square miles within the NRTS adjacent to foothill slopes near the western boundary	Derived from the mountains, foot hills, and basalt ridges adjacent to and on the NRTS
			Black basalt	0-20	Dark gray to black pahoehoe. Texture ranges from aphanitic to porphyritic. Vesicles, crevices, joints and other fractures, and open contacts between layers are common	Covers about 5 square miles in the extreme south-central part of the NRTS	Extruded from vent near Cedar Butte
			Lake beds and associated leach and bar deposits	0-140	Sandy and clayey light-colored silt, with a few layers of relatively pure sand, silt, and clay	Covers about 60 square miles in the northern NRTS. Thickness of 137 feet was drilled in test hole 6N-33E-26dd1	Sediments derived principally from south slopes of the Beaverhead, Centennial, and Lost River Ranges, transported by the Big Lost River, Birch Creek and intermittent local streams
		Holocene Pleistocene	Snake River Group (undifferentiated)	0-1,500	Chiefly ash-colored basalt. Texture ranges from glassy to porphyritic. Gray and black are the most common colors. Vesicles, crevices, joints, fractures, open contacts, and other voids are common. Basalt similar to that at surface was drilled at depth of nearly 1,500 feet in test well 4N-30E-66b1. Locally includes interflow beds of windblown, lacustrine, and alluvial sediments	Covers about 570 square miles within the NRTS, with outcrop in all parts of the station	Extruded from cones and fissures, some within and some outside the NRTS
			Unconformity				
Paleozoic	Tertiary	Fluviene ^a Suberene	Basalt	0-900	Basalt, blocky, black to brown, generally dense. Thickness at West Twin Butte is about 900 feet	Covers about 1 square mile in West Twin Butte	Extruded from depth in outcrop area
		Lower Tertiary	Volcanic rocks (undifferentiated)	0-800	Welded tuff, ash, pumice, and various types of flow rocks. Exposed thickness at East Butte is about 800 feet	Covers about 10 square miles at East Twin Butte and Thome Point	Flows originated from vents at or near the outcrop areas. Pyroclastic materials probably from same or related vents
	Cretaceous Tertiary	Unconformity					
			Paleozoic rocks	1,500	Primarily limestone. May include minor amounts of sandstone, siltstone, and conglomerate. Estimated exposed thickness in this vicinity exceeds 1,500 feet	Coverages about 2 square miles in northeastern part of NRTS	Deposited in an ancient sea

Table 4.3 - GENERALIZED GEOLOGIC FRAMEWORK OF THE NATIONAL REACTOR TESTING STATION, IDAHO

(After Nace, et al., 1975)

TABLE 4.4 - GEOLOGICAL HISTORY OF PASCO BASIN
(After ERDA, 1975)

ERA	SYSTEM	SERIES	GEOLOGIC UNIT	MATERIAL
CENOZOIC	QUATERNARY	HOLOCENE	DUMPS AND EOLIAN SEDIMENTS (0-40 FEET THICK)	SANDS, INCREASINGLY FINER AND QUARTZ-RICH TO THE NORTHEAST
			ALLUVIUM, COLLUVIUM, LANDSLIDES (0-100 FEET THICK)	UNSORTED RUBBLE AND DEBRIS, LOCALLY INTERFINGER WITH RINGOLD FORMATION AND PASCO GRAVELS
			PASCO GRAVELS AND THE TOUCHET BEDS (0-400 FEET THICK)	SANDS AND GRAVELS OCCURRING AS GLACIAL FLOOD DEPOSITS. COMMONLY ROUGHLY GRADED, UNCONSOLIDATED BUT HIGHLY COMPACT.
			PALOUSE SOILS (0-30 FEET THICK)	WIND-TRANSPORTED AND DEPOSITED SILT, LOCALLY WEATHERED TO CLAY
	PLEISTOCENE		RINGOLD	WELL-BEDDED FLUVIAL AND FLOOD-PLAIN SILTS, SANDS AND GRAVEL, POORLY SORTED, COMPACT BUT VARIABLY CEMENTED. BASAL PORTION LARGELY SILT AND CLAY OF HIGHLY VARIABLE THICKNESS. REMAINDER OF FORMATION IS INTERBEDDED GRAVEL, SAND AND SILT, GENTLY DEFORMED.
			FORMATION (0-1200 FEET THICK)	
	TERTIARY	PLIOCENE	ELLENBURG (20-200 FEET THICK)	VOLCANICLASTIC ROCKS AND THEIR WEATHERING PRODUCTS, LARGELY CLAYS, GRADES INTO AND INTERFINGERS WITH RINGOLD FORMATION SEDIMENTS.
			FORMATION	
		MIOCENE	YAKIMA BASALT FORMATION (PROB. 2500 FEET THICK)	BASALTIC LAVAS WITH INTERBEDDED STREAM SEDIMENTS IN UPPER PART, LOCALLY FOLDED AND FAULTED.
			PICTURE GORGE FORMATION (EQUIVALENT ?) (PROB. 1500 FEET THICK)	BASALTIC LAVAS
		OLIGOCENE		
MESOZOIC		Eocene		
		PALEOCENE		
			ROCKS OF UNCERTAIN AGE, TYPE AND STRUCTURE	PROBABLY METASEDIMENTS AND METAVOLCANICS INTRUDED BY GRANITIC ROCKS

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TABLE 4.5 - CONFINED AQUIFER PUMP TEST RESULTS
COLUMBIA RIVER BASALT
(After ERDA, 1975)

<u>Well Number</u>	<u>Transmissivity</u> <u>m²/day</u>	<u>Hydraulic Conductivity</u> <u>m/day</u>
199-B3-2	0.33	0.03
199-H4-2	0.21	<0.03
299-W11-2	9	0.03
399-5-2	0.56	<0.03
699-S18-E2	1.2	0.4
699-S11-E12	3.3	0.3
699-S6-E4-C	140-220	2.8-4.6
699-S1-7-B	150-200	3-4
699-10-E12	20	6
699-14-E6	40-70	2-3
699-14-38	37	40
699-20-E12	33	2.8
699-20-E5	40	0.4
699-24-1	8	0.3
699-28-40	0.45	0.1
699-34-88	2.2	0.03
699-84-35	0.4	0.1

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TABLE 4.6 - AVERAGE FIELD HYDRAULIC CONDUCTIVITY
(M /DAY) MEASUREMENTS
(After ERDA, 1975)

<u>Tested</u>	<u>Pumping Tests</u>	<u>Specific Capacity Tests</u>	<u>Tracer Tests</u>	<u>Cyclic Fluctuations</u>	<u>Gradient Method</u>
Glaciofluvial	370-3,700	400-5,200	2,400	670-2,300	-
Glaciofluvial and Ringold	37-200	40-160	-	40-240	-
Ringold (including clays)	0.3-60	2-12	-	6-18	4-12

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TABLE 4.7 STRATIGRAPHY AT THE NORTH END OF HAIKITE PLATEAU
(Based on data from Cooley et al., 1964; Cooley et al., 1964; McClymonds, 1961)

Graphic Log	Stratigraphic Unit or Multiple-aquifer System	Thickness of Unit Beneath Northern Kaibito Plateau Meters	System	Stratigraphic Unit	Lithology	Water-bearing Characteristics
	N Multiple-aquifer System	25	Jurassic	Carmel formation	Siltstone and some sandstone.	Yields water to a few small springs in northeastern Coconino County; forms upper confining bed of the N multiple-aquifer system.
		244	Jurassic and Triassic(?)	Navajo sandstone	Sandstone; thickens northwestward.	Unit is the chief aquifer of the N multiple-aquifer system; yields small to medium amounts of water principally in the northern part of the Navajo Indian Reservation; base of formation is conspicuous spring zone.
		40	Triassic(?)	Kayenta formation	Chiefly a sandstone in the northern part of the Navajo Indian Reservation; chiefly a siltstone in southwestern part of the reservation.	Yields water to a few springs; tongues of the Navajo sandstone in the upper part of the formation yield small amounts of water to wells near Tuba City; a few of these wells flow.
		112	Triassic(?)	Moencopse formation	Silt, sandstone and sandy siltstone.	Not water bearing.
				Wingate sandstone	Lukachukai Member is a sandstone. Rock Point Member is a siltstone and silty sandstone.	Yields small and moderate amounts of water to springs and wells in northern Apache and Navajo Counties. Yields some water to springs; unit generally does not yield sufficient water for drilled wells.
	Aquiclude	377	Triassic	Chinle formation	Alternating shaly units with sandstone beds.	Shaly units are not water bearing.
	C Multiple-aquifer System and Basal Aquiclude	300	Permian	Moenkopi formation	Siltstone and some sandstone.	Yields a small amount of water to a few wells.
				Kaibab limestone	Limestone and some limy sandstone.	Yields small amounts of water to a few wells near Leupp.
				Toroweap formation	Chiefly sandstone.	Usually not water bearing in the Navajo country.
				Coconino sandstone	Sandstone.	Unit is the chief aquifer of the C multiple-aquifer system; yields small to large amounts of water to wells in southwestern part of Navajo country.
				Hermit shale	Chiefly sandy siltstone.	Not water bearing.
			Pennsylvanian	Supai formation	Chiefly siltstone and sandstone.	Uppermost sandstone beds of formation interconnect hydrologically with the Coconino sandstone.
	Redwall Limestone (aquifer)	150	Mississippian	Redwall limestone	Limestone.	Yields large amounts of water to Blue Spring in canyon of the Little Colorado River.
		—	Cambrian	Moav limestone	Limestone.	Yields small amounts of water to a few springs in Marble and Grand Canyons; most of the water comes into the formation from the Redwall limestone.
				Bright Angel shale	Shale.	Not water bearing.
				Tapeats sandstone	Sandstone and some conglomerate; partly quartzitic.	Yields small amounts of water to springs in Grand Canyon.
	Basement Rocks	—	Precambrian	—	Granite	—

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Key
 Aquifer
 Aquiclude

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TABLE 4.8 - RANGE OF THE HYDRAULIC PROPERTIES OF THE
NAVAJO SANDSTONE AQUIFER IN THE NAVAJO
AND HOPI RESERVATIONS
(After Cooley et al., 1969, Table 7)

Rock unit	Navajo sandstone
Median grain size	0.18 mm
Porosity (lab)	25 to 35% (24 tests)
Hydraulic conductivity (lab)	1.4×10^{-4} to 2.3×10^{-2} cm/sec (24 tests)
Transmissivity (field)	5 to 47 m ² /day (4 tests)

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boundary between exterior and interior
drainage (outline of Great Basin)

Figure 4.1 - Outline and topography of the Great Basin
(from Fenneman, 1931)

Figure 4.2 - Index map of Nevada Test Site and vicinity, showing Ash Meadows groundwater system (after Winograd and Thordarson, 1975)

Relief = a few thousand meters or less
 Distance = 20 to 40 km between ranges

- Valley-fill aquifer and thin alluvial cover on highlands - interstitial flow
- Volcanic aquifer - flow through interconnected fractures
- Volcanic aquitard - fractures discontinuous - interstitial flow
- Carbonate aquifer - fracture flow
- Clastic aquitard - interstitial flow

Figure 4.3- Schematic cross section of a typical topographic basin showing discontinuous valley-fill and volcanic aquitard, and generally continuous carbonate aquifer.

Figure 4.5 - Schematic cross section perpendicular to Basin and Range topography of regional flow system showing effect of topographically low discharge point and gentle hydraulic gradient on the establishment of the deep water table. Barrier to groundwater flow causes reduction in depth to water by increasing gradient.

Note how the dam effect of the groundwater barrier causes a reduction in depth to the groundwater table.

Modified from Thordarson (1965, fig 4)

EXPLANATION

- Welded-tuff aquifer
- Bedded-tuff aquifer
- Tuff aquitard
- Lower carbonate aquifer

Test Hole

FRACTURE TYPES

- ① En echelon joints or faults, some contain water, others empty
- ② Faults, closed above tuff aquitard-carbonate aquifer contact, fully and partly saturated
- ③ Fault, open at bottom, empty
- ④ Fault, with pinch-and-swell structure, partly saturated

Water-filled fractures are black

Figure 4.6 - Diagrammatic cross section of Rainier Mesa showing depth to the regional water table and perched and semi-perched groundwater. Vertical exaggeration about 2 times (from Winograd and Thordarson, 1975).

Figure 4.8 - Schematic cross section of a topographic basin showing typical shallow groundwater flow system including upward flow under valley floors.

Figure 4.9 - The distribution of thick carbonate rock units in Nevada (the carbonate rock province; from Mifflin, 1968).

Figure 4.10- Sketch map of the distribution of ash flow tuff in the central Great Basin (from Winograd, 1971, after Cook {1960}).

Figure 4.11 - Index map of and generalized groundwater flow pattern in the White River groundwater system (from Eakin, 1966).

Figure 4.12 - Diagrammatic profile showing relation of water level to land surface along axis of the White River groundwater system (from Eakin, 1966).

Figure 4.13 - Index map of Idaho showing the Snake River Plain and the Idaho National Engineering Laboratory (from ERDA, 1977).

Figure 4.14 - Diagram showing the geometry and formation of spring alcoves discharging the Snake River Plain aquifer (from Stearns et al., 1938).

- 19 odl - well
 △ Stream gage
 --- INEL
 —250— Water depth in feet below the surface .
 Dashed where unsure.

0 1 2 miles

Figure 4.15 - Contour map of the depth to the water table under part of the Snake River Plain where a) the vadose zone is thickest and b) there is adequate data for contouring (from Mundorff *et al.*, 1964).

Figure 4.16 - Sketch of the waste disposal area showing distribution of tritium in the Snake River Plain aquifer in 1970 (from ERDA, 1977).

Figure 4.17 - Sketch of waste disposal area showing distribution of Strontium-90 in the Snake River Plain aquifer in 1970 (from ERDA, 1977).

Figure 4.18 - Distribution of tritium in the Snake River Plain aquifer in the year 2000 as predicted by the digital model, assuming disposal continues at the current rate (from ERDA, 1977).

Figure 4.19 - Index map of the Hanford Area, Washington State, showing the Pasco Basin and the Hanford Reservation (from National Academy of Sciences, 1978).

Figure 4.20 - Geologic cross section of the Hanford Reservation from east to west showing the distribution of subsurface materials and various positions of the water table (from ERDA, 1975).

Figure 4.21- Map of the distribution of hydraulic conductivity in the unconfined aquifer under the Hanford Reservation (from ERDA, 1975).

Figure 4.22 - Reconstruction of the water table under the Hanford Reservation as it probably occurred in 1944, before the start of Hanford operations (from ERDA, 1975).

Figure 4.23 - Contour map of the water table under the Hanford Reservation in 1975 (from ERDA, 1975).

Figure 4.24 - Map of the 1972 water table under Hanford including groundwater flow lines (from ERDA, 1975).
 Note: Convergence points of streamlines are questionable.

Figure 4.25 - Map of the Hanford Reservation showing the waste plume of tritium migrating in the unconfined aquifer (from ERDA, 1975).

Figure 4.26 - Map of the piezometric level in the uppermost confined aquifer in 1970 (from ERDA, 1975).

(a)

(b)

Figure 4.27 - Simplified geologic cross sections of the Hanford Reservation showing the water table and material in the vadose zone (from ERDA, 1975).

Figure 4.28 - Hydrologic basins and chief structural features of the Colorado Plateau in northeastern Arizona. The Copper Mine area, located within the Navajo Indian Reservation, is on the axis of the Preston Mesa anticline. (Cooley, et al, 1969).

Figure 4.29 - Section of buried anticline. The zone of interest in the unsaturated Navajo sandstone is shielded from precipitation by the Carmel formation and is freely drained by the deep canyons of the Colorado River. (McClymonds, 1961).

5. CHARACTERISTICS OF A VADOSE ZONE REPOSITORY

5.1 Introduction

Successful demonstration of geologic disposal of radioactive waste in a particular medium requires that two conditions be met: (1) that the medium exist at at least one geographic location, with sufficient areal extent such that reasonable criteria relative to population density and political boundaries could be met and (2) that the effects of pertinent geological and hydrological processes and of the activities associated with site exploration and waste emplacement are understood to the point that the containment and isolation of the radionuclides for the required time span could be assured.

Implicit in this second condition is the requirement that the characteristics of the medium which affect containment be measurable and that their effects can be bounded so that numerical estimates of potential release of radionuclides to the biosphere are possible.

Several of the geologic and hydrologic factors that characterize the vadose-zone-repository (VZR) concept are described in the following section. These factors are presented in two categories: those that appear advantageous for the containment of radioactive waste and those that do not. Section 5.3 outlines the analytic and numerical techniques available for modeling the transport of radionuclides from a VZR. Conclusions regarding the VZR concept are given in Section 5.4 and recommendations for additional study are presented in Section 5.5.

5.2 Factors Affecting the VZR Concept

5.2.1 Factors Favoring the VZR Concept

Large areas with deep water tables can be found in several parts of the United States. The Great Basin area of Nevada, which contains the Nevada Test Site (NTS), has two groundwater systems where water tables are as deep as 1,000 m over areas of many square kilometers. The Copper Mine area of the Colorado Plateau in Arizona typifies the type of geologic setting where deep drainage by canyons lowers the water table to depths of several hundred meters over areas of several hundred square kilometers. The water table at the Hanford Reservation, which is located on a portion of the Columbia River Plateau in Washington, reaches a maximum depth of about 100 m, and at the Idaho National Engineering Laboratory (INEL) of the Snake River Plain, the water table is locally as deep as 300 m.

Soil or rock with interstitial porosity constitutes the vadose zone in many arid areas. Studies indicate that downward movement of rainfall in arid areas is greatly retarded, if not eliminated, where soil or rock with interstitial porosity forms the upper 10 to 20 m of the vadose zone. Rainfall impinging on the surface of a porous medium is drawn into it by capillary and gravitational forces but it is held near the surface in pore storage after an approximate balance is reached between flow due to gravity, capillary forces and thermal gradients. Studies at Hanford suggest that, in that region, rainfall is held in the upper 12 m of soil, from which it is slowly lost by evapotranspiration. On the other hand, few data are available on infiltration through rock in which fracture permeability dominates over interstitial permeability.

Moisture flow can be extremely slow in media with interstitial porosity at low water contents. As a first approximation, the velocity of flow under gravity through porous media can be found by:

$$v = \frac{K(\theta) i}{\theta}$$

where v = velocity of flow

$K(\theta)$ = hydraulic conductivity, a function of θ

i = gradient

θ = volumetric water content.

There are no hydraulic conductivity measurements of unsaturated soil or rock with interstitial porosity at the low water contents typically found in arid regions. However, since hydraulic conductivity decreases with water content, an upper bound to the actual hydraulic conductivity of soils in arid areas can be estimated by using laboratory values determined for loam soils at the lowest water contents used in the tests (typically 10% by volume as opposed to typical field values of about 6% by volume). Typically, laboratory values of hydraulic conductivity of such soils are of the order of 10^{-11} cm/sec or less⁽¹⁾. For gravity drainage the vertical downward gradient (i) is one. However, downward flow would actually be slower than predicted using a gradient of unity because moisture movement caused by the geothermal gradient is opposite to that caused by gravity. Assuming a volumetric water content of 0.06, the velocity of flow in this instance would be about 1 m in 20,000 yr. This represents a maximum rate of radionuclide transport since sorption or chemical retardation has been neglected.

(1) Extrapolation of published hydraulic conductivity vs water content curves would give $K = 10^{-13}$ cm/sec or less at water contents of 6%. The higher value of 10^{-11} cm/sec is used for conservativeness.

The flow of moisture in an unsaturated medium is down thermal gradients. As shown in Section 3 moisture in unsaturated porous media, when subjected to a thermal gradient, flows toward lower temperatures. This has two consequences for a VZR: (1) moisture will flow away from a depository as long as the waste emits sufficient thermal energy (and if the depository is not inundated), and (2) moisture will tend to flow upward, mostly in the vapor phase, under the geothermal gradient of about $3^{\circ}\text{C}/100\text{m}$. Both of these phenomena provide an additional barrier to both the entry of water into the depository and the transport of radionuclides down to the water table.

Soil moisture does not flow into voids in an unsaturated medium with interstitial porosity. The outflow principle of Richards (1950) states that where capillary moisture tension is high (such as in the soils at depth in arid areas), water will not enter larger pores from smaller pores. Therefore, in the case of downward movement of moisture in unsaturated media water will not flow from a finer-grained medium into a coarser-grained medium until the upper stratum becomes saturated. The consequence of this phenomenon to the VZR concept is shown in studies by Frind et al. (1976), who demonstrate that potential downward infiltration can be circumvented around waste storage rooms by backfilling the rooms with a gravel that is coarser than the surrounding soil.

Areas with low population densities and deep water tables are available on large tracts of land in the western United States. There are several areas in the western United States which are so arid and where the water tables are so deep that irrigation and development are discouraged. In these areas of essentially desert terrain the population densities are near zero. In addition, much of the areas consists of large tracts of Federal land.

5.2.2 Factors Adversely Affecting the VZR Concept

Many arid areas with deep water tables are characterized by vadose zones composed of fractured rock that extends to the ground surface. In certain areas, such as INEL and in the mountains of the Great Basin of Nevada, rainfall is inferred or observed to infiltrate rapidly downward through fractured rock. Such inflow above a VZR could reduce the effectiveness of the vadose zone in containing radionuclides. Specifically, observations at Rainier Mesa at NTS suggest that rainfall readily percolates through several hundred feet of rock to the water table: inflow occurs in tunnels which are located in fractured rock 300 m below the rock surface and peaks only a few months after the

corresponding peak in precipitation. At INEL, a model which adequately simulates the groundwater hydrology is based on the assumption that 10% of the annual rainfall, some 2 to 3 cm, percolates down through fractured basalt to the water table. Therefore, unsaturated zones in rock may be less suitable if the vadose zone is considered a primary barrier to radionuclide migration. This phenomenon may eliminate a large class of sites with deep water tables because much additional field study is required to characterize movement of rainfall through fractured rock.

Rock that is overlain by an adequate thickness of soil (about 10 to 20 meters), however, would have the same lack of infiltration which is believed to exist in thick soil layers in arid climates.

Geologic faults and open fissures are exposed at the ground surface in certain arid areas. Recent faulting is observed at the ground surface in the alluvium in the vicinity of NTS, suggesting that relative movements at depth occur in the vadose zone. Most of the observed movement at this particular location appears to be associated with the underground nuclear test blasts. In addition, large cracks exist in the playa deposits and substantial quantities of surface water drain into these fissures following the infrequent rain storms. Both the faults in the alluvium and the fissures represent potential threats to the containment characteristics of an improperly-sited VZR.

Future climatic conditions are uncertain and of potential consequence. Climatologists, while stressing the cyclical nature of weather, are unable to predict, with certainty, the timing and duration of future temperatures and rates of precipitation and evaporation for an area. Inundation of an area above a VZR, such as might occur during a future period of increased rainfall and/or reduced evaporation, could have substantial consequence to the containment characteristics of a VZR. This is because the media between the ground surface and the water table could experience downward flow at the rate of the saturated hydraulic conductivity divided by the effective porosity (the hydraulic gradient would be unity).

The theory of unsaturated flow at water contents found at depth in arid areas has not been verified. Only recently have numerical techniques been devised for solving the equations of flow at low water contents (below about 10% by volume). The techniques have been calibrated with field data and have been used successfully to predict moisture movement for short time periods, but they have not been verified for materials with very low water contents at depths of hundreds of meters. Furthermore, there are no measurements of hydraulic conductivity at the low water contents found at depth in arid areas. Most measurements are made at water contents signifi-

cantly higher than those found in arid areas, giving values that may be several orders of magnitude higher than those that exist in-situ. These points may be moot, however, since as shown in Section 5.2.1, simple, conservative, analytical computations indicate that flow in the unsaturated zone may be of the order of 1 m in 20,000 years.

VZR sites generally are not as deep as potential sites in other geologic media. The typical depth of the water table in the four areas studied ranges from 250 m to 500 m. Assuming that a repository would be located 100 m above the water table to accommodate any fluctuations due to future climatic changes, the depository might be approximately 150 m to 400 m below the ground surface, compared with an 800 m depth at the Waste Isolation Pilot Plant (WIPP) in southeastern New Mexico.

Heat flow from a VZR has not been studied. Concern is expressed over the ability of soils or rock with interstitial porosity to transfer the heat generated by radioactive waste away from the canister area. This concern depends in large part upon 1) the rate of heat generation of the wastes at the time of emplacement, and 2) the thermal conductivity of the host medium. This topic may require additional study as the thermal characteristics of waste packaging schemes are developed.

5.3 Modeling a Vadose Zone Repository

The containment ability of a radioactive waste repository is commonly analyzed by subdividing the entire system into a number of discrete components, called "barriers," which impede the transport of radionuclides and which are connected in series and/or parallel networks to simulate the conditions that would exist under various escape scenarios.

One possible series of containment barriers of a VZR might be the following:

- Thermal barrier surrounding the waste
- Dissolution barrier
- Vadose zone transport barrier
- Aquifer transport barrier
- Human water usage patterns barrier

This section discusses the information required to model the first, third, and fourth barriers above. If each barrier is to be modeled as a hydrologic system, the following four aspects of the system must be known:

Flow equations
Geometry of the system
Hydraulic parameters of the materials
Boundary and initial conditions

Each of the three barriers is analyzed in the following sections with regard to the status of knowledge of the information required for modeling.

5.3.1 Thermal Barrier Surrounding the Waste

The solute transport equation for unsaturated porous media at low water contents under non-isothermal conditions has only recently been solved by Baca and King (1978), and while additional studies are in progress, there has been only minimal verification of the solution to date. The equations and model are discussed in more detail in Section 3; however, it is our opinion that, at the present time, the model is in the developmental phase. If future verification studies are successful, the model may become an important tool in analyzing solute transport under the above conditions. At present, however, there is no accepted model that can be used to compute rates of radionuclide transport under the thermal and moisture regime that would exist around a VZR.

Geometric considerations for radionuclide transport modeling in the vadose zone are relatively simple compared with modeling in the phreatic, or saturated zone. Gradients are essentially vertical, except in the immediate vicinity of the waste canisters, so one-dimensional models are adequate to represent the physics of flow in-situ. For detailed modeling of waste dissolution, two- or three-dimensional modeling would be required.

The hydraulic parameters remain problematic. There are no published values of hydraulic conductivity for porous media at water contents below 10% by volume, whereas the water contents of Hanford soils are consistently below 7% in water content. Previously, there has been little interest in making these measurements because for most practical agricultural or engineering problems, water contents are higher than 10%. Published hydraulic conductivity vs. water content data, such as shown in Appendix B, can be extrapolated to low water contents, but the uncertainties may be several orders of magnitude, and there are no published data for the soil or rock of the four areas represented in this study. Diffusivity parameters due to thermal gradients are even less understood.

Because moisture flow under non-isothermal conditions is coupled with heat flow, the boundary conditions for a repository model would include time-varying temperatures at the ground surface and at the waste canisters in addition to

the fixed moisture content at the water table and the time-varying moisture contents near the ground surface. Initial and boundary conditions were not studied in detail because other limitations to unsaturated-flow modeling restrict the application of current theory to vadose zone repositories.

Previous experiments have demonstrated that the flow of moisture under thermal gradients is toward lower temperatures. While, at present, it is not possible to predict with confidence the rate of flow, it is probably sufficient to know that flow will be away from the waste as long as the surrounding soil is at a relatively dry water content. Additional study is needed to determine if the heat generated after the fission products have decayed is sufficient to keep moisture from migrating down toward the waste under gravitational flow.

5.3.2 Vadose Zone Transport Barrier

If further studies show that heat production after decay of the fission products is insufficient to prevent moisture from contacting the waste, dissolution would require analysis. Studies have shown that much of the movement of moisture at low water contents occurs in the vapor phase, and therefore cannot transport dissolved radionuclides. Modeling of radionuclide transport under these conditions is influenced by the same constraints noted in Section 5.3.1 above; namely, that the transport theory has not been verified, and that the hydraulic parameters have not been measured.

The velocity of moisture migration was found in Section 5.2.1 to be less than 1m in 10,000 yrs for soils with volumetric water contents below about 10%, based on the available published data for the hydraulic conductivity of loam soils. At this rate of moisture movement, flow modeling would be of academic interest only. Other phenomena, such as climatic changes or sudden inundation would be of greater importance for establishing the feasibility of the VZR concept. The various accident scenarios should be modeled so that the consequences can be considered in light of the probabilities of occurrence.

5.3.3 Aquifer Transport Barrier

To analyze the consequences of radionuclides reaching the water table below a VZR, transport could be modeled in a manner similar to that currently employed in the Lawrence Livermore Laboratory waste transport code. Ground-water flow is commonly subject to complex boundary conditions

and geometries and is more reliably analyzed on a site specific basis. Two-dimensional groundwater models are currently used at both INEL and the Hanford Reservation, but no groundwater modeling has been attempted at NTS or at the Copper Mine area of the Colorado Plateau.

Borg et al. (1976) studied the status of knowledge regarding groundwater modeling at NTS and concluded that, because of complex aquifer geometry, boundary conditions, and lack of parameter data, one-dimensional modeling was the greatest degree of sophistication warranted at NTS. No data or interpretations from this study contradict these conclusions, which state: "...currently only crude predictions of radionuclide transport at NTS can be made. Reliable and convincing predictions await further evaluation of the parameters that control transport in the NTS area" (Borg et al. 1976).

There has been no groundwater modeling in the vicinity of the Copper Mine area although the relatively simple geology, discussed in Section 4.4, suggests that first order, one- or two-dimensional modeling could be performed if additional field data are obtained regarding the geometry and hydraulic properties of the Navajo sandstone aquifer.

5.4 Summary and Conclusions

Large, relatively unpopulated areas with deep water tables exist in arid regions in the western United States. Such areas, where underlain by soil or rock with interstitial porosity, warrant additional study with regard to radioactive waste disposal, because moisture flow through some arid site vadose zones is so slow that adequate containment could be achieved. Furthermore, what limited flow does exist would probably not enter the canister area for two reasons: first, moisture flows down thermal gradients (i.e., away from sources of heat such as radioactive waste), and, secondly, moisture will not flow from the host medium into the canister area if the depository is left open or backfilled with a material, such as coarse gravel, that has larger pore spaces than the host medium. These factors suggest that the vadose zone can provide an effective barrier to radionuclide migration from a radioactive waste repository.

The major uncertainty regarding the vadose zone repository concept concerns the potential for increased rainfall or decreased evaporation, in which case water might percolate downward through the canister area to the water table. Rainfall stability through the past several millenia of recorded history, however, suggests that climatic conditions are unlikely to change significantly during the several hundred years required for containment of the fission products in HLW.

At present, a lack of both verification of theory and parameter-measurement technology makes it impossible to model mathematically the transport of radionuclides from a vadose-zone repository. Conservative analytic computations show that moisture would flow with a velocity of less than one meter in 10 years under the conditions likely to exist in arid-site vadose zones consisting of materials with interstitial porosity. Mathematical modeling should be performed, however, to analyze the consequences of accident scenarios such as sudden inundation of the ground surface.

5.f Recommendations

Considerable additional research is required before the engineering feasibility of the VZR concept can be established. The topics listed below represent those that we feel are fundamental to further characterization of the VZR concept. The topics are listed in relative order of importance, and the answers to some questions may obviate the importance of others; for example, if the critical containment period is the 300 to 600 years required for decay of the fission products of HLW, the question of future climatic conditions reduces to assuming that a major climatic change will not occur within that time period. Similarly, if waste cannot be dissolved or leached by predominantly vapor flow through coarse backfill immediately surrounding the waste canisters, then the question of unsaturated flow rate becomes irrelevant.

The first five topics are generic and would therefore apply to any potential site:

1. Required time for containment and isolation

What is the critical period for containment? for isolation?

2. Future Climate

What is the maximum credible future rainfall, and will the maximum occur during the critical period for containment?

What will be the rate of infiltration during future periods of increased rainfall and/or decreased evaporation?

3. Geochemistry

Can waste canisters dissolve if embedded in a moist but unsaturated medium?

What values of the distribution coefficients (K_d) are appropriate for unsaturated soil and rock at various water contents?

4. Heat Flow

Can heat produced by the waste be dissipated adequately through the coarse backfill that may be required between the canisters and the host medium?

What will be the characteristics of the thermal regime in the rock or soil surrounding the waste?

5. Hydrology - Unsaturated

How fast, and in what direction does moisture move in arid site vadose zones with very low water contents under non-isothermal conditions?

Can radionuclides be transported by predominantly vapor flow, which may be found in arid site vadose zones?

If moisture flow in the vadose zone is significant, what are the relevant flow equations and parametric values? How can the parameter values be measured?

What are the characteristics of moisture flow through unsaturated, fractured rock, if the repository is to be constructed in rock?

What is the potential for migration of gaseous radionuclides in the unsaturated zone?

The last two topics are site specific and would require additional detailed study at particular areas of interest.

6. Hydrology - Saturated

How stable will the present water table elevation be in future periods of increased precipitation or reduced evaporation?

What are the relevant geometries, parameter values, and boundary conditions for saturated groundwater flow?

7. Geology

What is the likelihood of faults, glaciers, earth cracks, or other features that could short-circuit the vadose-zone barrier?

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APPENDIX A
TABULATED SUMMARY OF LITERATURE REVIEW

APPENDIX A

This appendix consists of a summary of most of the literature which was reviewed for this study. The information is categorized into the following subject areas:

Table 1 Theory of Moisture Flow in Unsaturated Porous Media Under Isothermal Conditions

Table 2 Theory of Moisture Flow in Unsaturated Porous Media Under Non-isothermal Conditions

Table 3 Unsaturated Flow Experiments

Table 4 Measurement of Parameters Pertinent to Moisture Flow in Unsaturated Soils

Table 5 Additional References - General Discussions of Unsaturated Flow

In these tables, several symbols are used which are defined as follows:

D	Diffusivity (l^2t^{-1})
h	Depth of water ponded on ground surface (l)
K	Hydraulic Conductivity (lt^{-1})
K_o	Saturated Hydraulic Conductivity (lt^{-1})
q	Moisture flux ($l^3l^{-2}t^{-1}$) or (lt^{-1})
R	Rain Intensity (lt^{-1})
T	Temperature
t	Time (t)
u	Rate of advance of wet front (lt^{-1})
v_o	Rate of infiltration ($l^3l^{-2}t^{-1}$) or (lt^{-1})

x, z	Distance, depth (l)
1D	One-dimensional Analysis
2D	Two-dimensional Analysis
α	Experimental constant
β	Experimental constant
θ	Volumetric water content ($l^3 l^{-3}$)
θ_n	Volumetric water content at $t = 0$ ($l^3 l^{-3}$)
θ_o	Volumetric water content at saturation ($l^3 l^{-3}$)
ϕ	Total potential per unit volume ($\Psi + \Psi_g$) ($ml^{-1}t^{-2}$)
Ψ	Matric potential per unit volume ($ml^{-1}t^{-2}$)
Ψ_g	Gravitational potential per unit volume ($ml^{-1}t^{-2}$)

Ψ -based equations are of the form:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{\partial \Psi}{\partial z} \right) + \frac{\partial K}{\partial z}$$

θ -based equations are of the form:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left(D \frac{\partial \theta}{\partial z} \right) + \frac{\partial K}{\partial z}$$

TABLE 1 - THEORY OF MOISTURE FLOW IN UNSATURATED
ISOTHERMAL POROUS MEDIA

Page 1 of 2

Useful Data

Author, Date	Type of Equation	Special Assumptions+	Type of Solution	Applications	Useful Data
Klute, 1952a	1-D, Q-based		Boltzman transformation, iterative	Sand column example	
Klute, 1952b	1-D, Q-based	$K = f(\theta)$ defined	Boltzman transformation, iterative	1/2 to 1/4 mm sand	θ vs x
Philip, 1954	1-D, Q-based	Empirical approximations	Algebraic approximation		
Youngs, 1957	1-D, Q-based		Series approximation-finite difference	Slate dust and glass beads-infiltration experiment	
Philip, 1957a	1-D, Q-based	Short time span	Power series approximation, iterative	Numerical example for Yolo light clay-infiltration	θ vs x at several times
Philip, 1957b	1-D, Q-based	Examines limit as $t \rightarrow \infty$	Geometric, finite difference	Numerical example for Yolo light clay-infiltration	v_o vs t as $t \rightarrow \infty$, $v_o = K_o$
Philip, 1957d	Empirical		Algebraic approximation	Numerical example for Yolo light clay-infiltration	
Takagi, 1960	1-D, Q-based		Algebraic approximation		
Scott and Hanks, 1962	1-D, Q-based	$D(\theta) = ae^{B\theta}$	Taylor power series		
Hanks and Bowers, 1962	1-D, Q-based	D and K are given approximate functions	Finite difference	Loam soil - measurements compared with Scott, 1962 and Philip, 1957	v_o vs t , θ vs $x/\sqrt{t}$ $\partial \theta / \partial x$ for various layer sequences
Ashcroft et al., 1963	1-D, Q-based		Finite difference	Several soil types	θ vs x
Rubin and Steinhardt, 1963	1-D, Q-based		Finite difference		
Rubin, 1966	1-D, Q-based		Finite difference	Examples for Rehovot Sand	
Staple, 1966	1-D, Q-based	Accounts for hysteresis	Finite difference	-2 mm silt loam	ψ vs θ , θ vs Z
Jensen and Hanks, 1967	1-D, ψ -based	$K = f(\theta)$ from curvilinear interpolation	Same as Hanks and Bowers, 1962	Comparison of several solutions	
Remson et al., 1967	1-D, Q-based	Heterogeneous, hysteretic soil	Finite difference	Drainage of sandy loam	θ vs Z
Freeze, 1969	1-D, ψ -based	Hysteretic	Same as Rubin, 1963	Several hypothetical examples	ψ vs θ , ψ vs K influencing factors

+ Otherwise assumed non-hysteretic, isothermal, isotropic, homogeneous, $h=0$, air escapes readily

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TABLE 1 - THEORY OF MOISTURE FLOW IN UNSATURATED
ISOTHERMAL POROUS MEDIA

Page 2 of 2

Author, Date	Type of Equation	Special Assumptions*	Type of Solution	Applications	Useful Data
Hanks <u>et al.</u> , 1969	1-D, θ -based	Hysteretic	Same as Hanks and Bowers, 1962	Nunn clay loam	
Cisler, 1972	1-D, ψ -based	Anisotropic			
Parlange, 1972	1-D, θ -based		Singular perturbation	Infiltration from a furrow	
Yong and Warkentin, 1972	1-D, θ -based	Swelling soil	Analytical error function	Clay and glass beads	θ vs $x/\sqrt{t}$, u vs $\sqrt{t}$
Reichardt <u>et al.</u> , 1972	1-D, θ -based	Layered soil	Finite difference	Clay, loam, silt loam, sandy loam	θ vs x , ψ vs x various layer sequences
Neuman, 1973	1-D, ψ -based		Finite element	Transient flow through a dam, and into a ditch	
Aylor and Parlange, 1973	1-D, θ -based	Layered soil, empirical approximations	Method of Parlange, 1972		
Giesel <u>et al.</u> , 1973	1-D, ψ -based	Hysteretic	Finite difference	Comparison with reported experiments	ψ vs θ , K vs θ , θ vs z , ψ vs z
Parlange, 1975	1-D, θ -based	$C = a\theta^{B_0}$	Analytical approximation (iterative)		
Wind and Van Doorne, 1975	1-D, ψ -based	$K = K_0 e^{a\psi}$, wet soil	Digital model with approximate equation		
Frind <u>et al.</u> , 1976	1-D, ψ -based	$K(\theta) = a\theta^B$	Finite element	Wick effect - repository	Sketch of repository with wick layer
Parlange and Babu, 1976	1-D, θ -based	Optimization - 2nd order approximation			
Perrins and Watson, 1977	2-D, ψ -based	Hysteretic	Finite difference	No. 17 Ottawa sand, sandy loam	Perspective plots of θ vs x vs z

* Otherwise assumed non-hysteretic, isothermal, isotropic, homogeneous, $h \cdot 0$, air escapes readily

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**TABLE 2 - THEORY OF MOISTURE FLOW IN UNSATURATED
POROUS MEDIA DUE TO THERMAL GRADIENTS**

Page 1 of 3

Author, Date	Type of Analysis	Special Assumptions	Type of Solution	Applications	Useful Data	Notes
Gurr et al.,	Experimental	-	-	Salt tracer in soil column	$\bar{z}$ vs x from heat	Water flows from hot to cold in vapor form and is accompanied by a lesser reverse flow in liquid form
Rollins, et al., 1954	Gaseous diffusion	Tortuosity factor	-	Silty loam lab columns	q vs ∇T , $\bar{Q}$ vs x from heat	
Taylor and Cavazza	Experimental	-	-	Air gap experiment	-	
Philip and De Vries, 1957	Moisture dependent diffusion, combined heat and mass transfer equations	Unique Ψ vs $\bar{Q}$	-	-	-	Flow process theory is proposed
Kuzmak and Sereda, 1958	Experimental	-	-	Air gap, salt tracer, streaming potential experiments	-	Flow is in vapor form only
Woodside and Kuzmak, 1958	Experimental	-	-	Large-scale pore analysis	-	Temperature-gradient in pores is much greater than across sample
Taylor and Cary, 1960	Thermodynamics of irreversible processes	Onsagers reciprocity, Phenomenological laws	Equations are linear	Thermal, electrical and salt gradient experiments	-	
Cary and Taylor, 1962a	Thermodynamics of irreversible processes	Onsagers reciprocity, Phenomenological coefficients, vapor transfer only	Equations are linear	Millville loam	Vapor flux vs temperature gradient	Lab test required for coefficients
Cary and Taylor, 1962b	Thermodynamics of irreversible processes with modification for liquid flow	Onsagers reciprocity, Phenomenological coefficients near equilibrium, steady state. Coefficients are independent of T , D & D pure water, no other gradients present	Equations are linear	Millville loam	Moisture flux vs temperature gradient	Lab tests required for coefficients
Jackson, 1963	Experimental	-	-	Loam, silt and clay	$\bar{Q}$ vs x , D vs $\bar{Q}$	-

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TABLE 2 - THEORY OF MOISTURE FLOW IN UNSATURATED
POROUS MEDIA DUE TO THERMAL GRADIENTS

Page 2 of 3

<u>Author, Date</u>	<u>Type of Analysis</u>	<u>Special Assumptions</u>	<u>Type of Solution</u>	<u>Applications</u>	<u>Useful Data</u>	<u>Notes</u>
Cary, 1963	Thermodynamics of irreversible processes	Onsagers reciprocity, Phenomenological coefficients,	Equations are linear	Vapor diffusion experiment to check Onsager's theory	-	Lab tests required for coefficients
Taylor and Cary, 1964	Thermodynamics of irreversible processes	Onsagers reciprocity, Phenomenological coefficients, θ vs $\ln T$ is linear, ψ vs $\ln T$ is linear, soil is inert, system is near equilibrium, flow can be described linearly, all fluxes are accounted for, if diffusion form - no hysteresis	ψ -based or θ -based equations	-	-	Lab tests required for coefficients
Cary, 1965	Thermodynamics of irreversible processes	See Taylor and Cary, 1964	Equations are linear	Columbia loam	Moisture flux vs temperature gradient	Lab tests required for coefficients
Cary, 1966	Diffusion based equation	See Taylor and Cary, 1964	Equations are linear	Diurnal heat fluctuations in field soil, moisture flow to freezing zone	-	Good discussion of theory and applications
Hadas, 1968	Experimental	-	Cary-Taylor equations, Philip and De Vries equations	Gilat silty loam, sinusoidal temperature fluctuations	Table comparing predictions with measurements	Both theories underestimated flow. Cary-Taylor gave best prediction
Cassel et al., 1969	Experimental	-	Cary-Taylor equations, Philip-De Vries equations	Columbia fine sandy loam under constant ψ	Table comparing predictions with measurements	Philip-DeVries gave good prediction
Moench and Evans, 1970	Experimental	-	Measurement of thermal conductivity and diffusivity	Sandy loam soil with cylindrical heat source	Tabulated values vs water content	

TABLE 2 - THEORY OF MOISTURE FLOW IN UNSATURATED
POROUS MEDIA DUE TO THERMAL GRADIENTS

Page 3 of 3

<u>Author, Date</u>	<u>Type of Analysis</u>	<u>Special Assumptions</u>	<u>Type of Solution</u>	<u>Applications</u>	<u>Useful Data</u>	<u>Notes</u>
Jury, 1973	Combined Taylor-Cary and Philip-DeVries	Onsagers reciprocity, uniform gas pressure, ideal gas law, moisture flow is chiefly liquid	Orthogonal Collocation	Analysis tested on medium sand	ψ vs Q , K vs ψ , K vs Q , D vs ψ	Lab tests required to obtain coefficients, difficult to measure coefficients at low Q
Kay and Groenevelt, 1974	Combined Taylor-Cary and Philip-DeVries with Claapeyron equation	Horizontal flow, salt-free water, no electric potential, rigid soil matrix, vapor is at atmospheric pressure	New equations for coupled flow including ice lenses	Frozen and unfrozen soils	-	-

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TABLE 3 - UNSATURATED FLOW EXPERIMENTS

Author, Date	Type of Experiment	Type of Soil	Theory Applied	Conclusions	Page 1 of 2
					Useful Data
Youngs, 1958a	Redistribution	Slate dust and Bal-lotini glass beads	-	Diffusion theory has limitations	-
Youngs, 1958b	Redistribution	Slate dust and Bal-lotini glass beads	-	Coarse soils exhibit sharp wetting front	θ vs z with time
Remson et al., 1959	Infiltration	Sandy soil in field Seabrook, N. J.	-	Water will not reach groundwater table until soil column is at field capacity. Each rain may establish a separate wetting front	Table of water penetration vs rainfall
Fenimore, 1964	Monitoring of waste burial	Sandy clays	-	Migration of waste was generally about 2 ft in 10 yrs. One area, subject to a perched water table, showed 7 ft migration	-
Rubin et al., 1964	Infiltration	Rehovot Sand	Rubin and Steinhart, 1963 (θ -based, finite difference)	Low intensity rain infiltration can be predicted accurately	ψ vs θ , u vs t , u vs R
Gupta and Staple, 1964	Infiltration	Grenville silt loam	Philip, 1957 (θ -based, iterative) and explicit finite difference approximation	Philip theory underpredicted u finite difference theory with provision for high K near surface gave good prediction	ψ vs θ , θ vs z , D vs θ , K vs θ
Horton and Hawkins, 1965	Infiltration flow path	Sand and glass beads, loam and sand	-	Larger pores empty first, flow is by downward displacement of water initially held in soil pores. Entrapped air serves as a barrier to infiltration	-
Biswas et al., 1966	Redistribution	Hanford sand, Columbia silt loam, Aiken clay loam	ψ -based equations	Rate of redistribution is dependent on initial depth of wetting and ψ vs θ relation of soil. Type of soil markedly effected the shape of redistribution curves	θ vs z for 3 soils with time
Schmaltz and Polzer, 1969	Infiltration, redistribution of tritium in rain	Silt loam to silty clay loam	Diffusion theory plus convection to predict tritium spreading	Prediction by theory was within an order of magnitude	$D = 0.9 \times 10^{-6}$ to 9×10^{-6} cm ² /sec $u = 3$ cm/month

TABLE 3 - UNSATURATED FLOW EXPERIMENTS

Page 2 of 2

<u>Author, Date</u>	<u>Type of Experiment</u>	<u>Type of Soil</u>	<u>Theory Applied</u>	<u>Conclusions</u>	<u>Useful Data</u>
Staple, 1969	Redistribution	Uplands sand, custer loam, Rideau clay	Staple, 1966 (θ -based explicit F.D.) Hanks and Bowers, 1962 (θ -based implicit F.D.)	Good agreement with theory. D and K can be determined from wetting profile. θ depends on depth wetted and K vs ψ relations	ψ vs θ , r vs θ , K vs θ , (scanning curves), θ vs z with time
Freeze and Banner, 1970	Field and lab, ground-water recharge	Medium sand in lab, sandy loam to sandy gravel in field	Freeze, 1969	Mathematical prediction is good but need to know functional relationship among ψ , K and θ	Table of recharge information
Vachaud et al., 1972	Infiltration, drainage	Layered fine and coarse sand	Two-phase immiscible fluids (Buckley and Leverett, 1942)	Development of water profiles is strongly dependent on air flow	θ vs z with time (effect of air pressure)

**TABLE 4a - MEASUREMENT OF PARAMETERS PERTINENT TO MOISTURE
FLOW IN UNSATURATED SOIL - LABORATORY**

Page 1 of 1

<u>Author, Date</u>	<u>Parameters Measured</u>	<u>Method</u>	<u>Type of Soil Tested</u>	<u>Useful Data and Notes</u>
Bruce and Klute, 1956	Diffusivity	Infiltration to horizontal column. $\theta(x)$ gravimetrically in sections.	Compacted samples of glass beads, fine sand, silty fine sand	Using Boltzmann transformation, determined $D(\theta)$. Diffusivity peaked at moisture content less than saturation.
Nielsen et al., 1960	Hydraulic conductivity	Steady state flow under potentials up to 100 cm of water. Measured Q vs ψ on pressure apparatus.	Undisturbed sample of loess and glacial till	Compared experimental results with K vs ψ curves predicted by the method of Childs and Collis-George (CCG) and the method of Marshall. The CCG method showed good agreement with experimental data.
Elrick and Bowman, 1964	Hydraulic conductivity	Steady state and transient tests in pressure cell. Moisture content obtained at equilibrium by weighing sample.	Compacted samples of loam	Found hysteresis in Q vs ψ curves but not in $K(\theta)$ curves.
Jackson, 1964	Diffusivity	Horizontal cylinder divided in 1 cm sections in diffusion chamber under constant vapor pressure. Measured Q at different times.	Compacted samples of loam and silty clay	Shown linearity of Q vs $x t^{-1/2}$ and peak in $D(\theta)$ curve at θ less than saturation.
Youngs, 1964	Hydraulic conductivity	Infiltration to a vertical column. Measured Q gravimetrically.	Compacted samples of slate dust.	-
Jackson et al., 1965	Hydraulic conductivity	Steady state desorption and sorption. Measured Q and potential.	Compacted samples of fine sand	Compared results with prediction by several other methods. Millington and Quirk method showed best fit when matching factor used.
Watson, 1966	Hydraulic conductivity	Transient method. Moisture content measured by gamma ray attenuation during desorption.	Compacted samples of fine sand	Instantaneous profile method.
Topp and Miller, 1966	Hydraulic conductivity	Transient test on small rectangular sample. Moisture content measured by gamma ray attenuation.	Glass beads	Large hysteresis observed in Q vs ψ and K vs ψ curves but little in K vs θ curves. Found independent domain theory inapplicable for predicting scanning curves.
Watson, 1967	Hydraulic conductivity	Method of Watson (1966).	Compacted samples of fine sand	Finite difference technique of predicting water profiles during desorption compared favorably with measured profiles.

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TABLE 4a - MEASUREMENT OF PARAMETERS PERTINENT TO MOISTURE
FLOW IN UNSATURATED SOIL - LABORATORY

Page 2 of 3

<u>Author, Date</u>	<u>Parameters Measured</u>	<u>Method</u>	<u>Type of Soil Tested</u>	<u>Useful Data and Notes</u>
Weeks and Richards, 1967	Hydraulic conductivity	Desorption of horizontal column.	Compacted and undisturbed samples of loam	Developed empirical equations for K and D under transient flow conditions. Required knowledge of Ψ vs θ .
Kunze et al., 1968	Hydraulic conductivity	-	-	Compared measured K with calculated K and found that Millington and Quirk method gives best results if a matching factor is used. Requires knowledge of K_0 and θ vs Ψ .
Hillel and Gardner, 1970	Hydraulic conductivity, diffusivity	Measured infiltration under transient and steady state conditions to soil with surface crust.	Compacted samples of loam	Determined weighted mean diffusivity and hydraulic conductivity as a function of θ . By varying resistance in crust, different sections can be developed.
Green and Cory, 1971	Hydraulic conductivity	-	-	Compared, measured and calculated $K(\theta)$ curves. Found that Marshall and Millington and Quirk methods gave similar results.
Topp, 1971	Hydraulic conductivity	Sorption and desorption of vertical columns subjected to differential air pressure.	Compacted samples of clay loam and silt loam	$K(\theta)$ curves showed little hysteresis but for clay loam there was scatter due to rearrangement of sample. Scanning θ vs Ψ curves were measured, and independent domain theory found inapplicable.
Vachaud and Thony, 1971	Hydraulic conductivity	Sorption and desorption of vertical column.	Compacted samples of sand	Concluded hysteresis in $K(\theta)$ can be neglected.
Bruce, 1972	Hydraulic conductivity	θ vs Ψ and diffusivity with pressure plate apparatus.	Compacted and undisturbed samples of loam	Calculates $K(\theta)$ from θ vs Ψ curves using reversal methods. Found that Millington and Quirk and Marshall theories accurate for coarse-grained materials but difficult to use for fine-grained materials. Found large difference in θ - Ψ curves derived with undisturbed and reconstituted samples.
Haridasan and Jensen, 1972	Hydraulic conductivity	Steady state flow experiments in pressure cell. Transient flow experiments in pressure plate apparatus.	Compacted samples of two silt loams	Performed tests at 15° , 25° , and 35°C and compared steady state and transient methods. Transient method sensitive to temperature fluctuations. Steady state method and K vs Ψ not affected by temperature.

TABLE 4a - MEASUREMENT OF PARAMETERS PERTINENT TO MOISTURE
FLOW IN UNSATURATED SOIL - LABORATORY

Page 3 of 3

<u>Author, Date</u>	<u>Parameters Measured</u>	<u>Method</u>	<u>Type of Soil Tested</u>	<u>Useful Data and Notes</u>
Elzeftawy and Mansell, 1975	Hydraulic con- ductivity	Steady state and transient determinations in pressure cell apparatus.	Compacted and un- disturbed samples of fine sand	In transient method, θ vs γ curve depends on rate at which equilibrium is attained during drainage.
Mehuys <u>et al.</u> , 1975	Hydraulic con- ductivity	Transient outflow method	Compacted samples of three sands with and without +2 mm fraction	Calculated K using Weeks and Richards (1967) method. Pre- sence of +2 mm material did not affect K vs γ curve, but must be accounted for when determining θ for K vs θ curves.
Gilham <u>et al.</u> , 1976	Hydraulic con- ductivity	Instantaneous profile method on vertical columns.	Compacted samples of sand.	Determined scanning curves.
Alami <u>et al.</u> , 1976	Hydraulic con- ductivity	Centrifuge	Undisturbed samples of loam	Compaction of sample may occur during centrifugation.

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**TABLE 4b - MEASUREMENT OF PARAMETERS PERTINENT TO MOISTURE
FLOW IN UNSATURATED MEDIA - FIELD**

<u>Author, Date</u>	<u>Parameters Measured</u>	<u>Method</u>	<u>Type of Soil Tested</u>	<u>Useful Data and Notes</u>
Bouwer, 1964	Hydraulic conductivity	Double tube infiltrometer	-	Determined horizontal and vertical hydraulic conductivity.
Rose et al., 1965	Hydraulic conductivity	Measured moisture content with neutron moderation equipment after irrigation. θ - ψ curves obtained in lab.	Sandy loam grading into clay	-
Gardner, 1970	Diffusivity	-	-	Developed an equation for calculating D from in-situ tests in which potential was measured with depth. Best suited for moisture contents near saturation.
Roulier et al., 1972	Hydraulic conductivity	In field - measured potential with tensiometers and moisture content with neutron probe.	Sandy loam	Field θ vs h curves agreed well with lab curves from 30 cm cores but not 6 cm cores. Found transient flow method of Richards and Weeks (1967) better to estimate K in field than Millington and Quirk theory using θ - h curves.
Baker et al., 1974	Hydraulic conductivity	-	-	In field, best method to install tensiometers is in a soil slurry in a borehole. Best way to measure θ in field is with neutron probe. If not possible, then can get θ from θ - h curves derived in lab on long cores. When instantaneous profile method is used in the field, there may be problems if the strata are not horizontal and if there are layers which have a lower permeability than the other layers.
Baker, 1977	Hydraulic conductivity	Infiltration to a cylindrical pedestal of soil capped with a crust of gypsum.	Medium sand and silty clay loam	Determined K vs h curve by using crusts with different amounts of gypsum. Diameter and length of pedestal affects results.

TABLE 5 - ADDITIONAL REFERENCES
FOR GENERAL DISCUSSIONS
OF UNSATURATED FLOW

This Table contains a list of references with a text book type treatment of the general aspects of unsaturated flow. For the reader interested in an overview of the entire subject of unsaturated flow, but more detail than presented in this report, the references listed below are recommended. Hillel (1971) and Yong and Warkentin (1975) were the most useful in the preparation of this report.

1. Bear et al, 1968
2. Childs, 1969
3. Hillel, 1971
4. Philip, 1958
5. Philip, 1970
6. Swartzendruber, 1963
7. Yong and Warkentin, 1966
8. Yong and Warkentin, 1975

APPENDIX B

SOIL-WATER RELATIONSHIPS

APPENDIX B

Solution of the general flow equation presented in the text requires a knowledge of the water retention and conductivity (or diffusivity) relationships of the porous medium in question. Empirical information on such relationships has been obtained by several investigators and this appendix contains a compendium of data which have been reported in the literature. (A tabulated summary of the literature is presented in Appendix A.) For discussion purposes, the data are divided into the following four categories:

1. Water Content vs Soil Moisture Potential
2. Water Content vs Hydraulic Conductivity
3. Water Content vs Diffusivity
4. Hydraulic Conductivity vs Soil Moisture Potential

Each of these relationships is discussed below. In these discussions, mention is often made to data which have been taken from a particular reference. For convenience, copies of the actual table or graph as presented by the original author are included in this appendix and are referenced in this text by page number and figure letter (for example, B8A refers to Figure A on page B8).

Water Content vs Soil Moisture Potential

The relationship between soil moisture potential (Ψ) and volumetric water content (θ), frequently called a retention curve, is fundamental to the solution of the flow equation since flow under isothermal conditions is caused by a potential gradient. The Ψ vs θ curves presented herein illustrate several characteristics of soil water retention:

1. In general, soil water potential becomes increasingly negative as the volumetric water content decreases due to the increasing influence of adsorptive and capillary forces. At the very low water contents found in arid site vadose zones, the soil water potential may be so low that gravity may have little influence on the movement of moisture.

2. At low water contents, the slope of Ψ vs θ curves become very steep, indicating that large changes in potential result from small changes in water content. The steepness of this curve depends on soil type: coarse-grained soils exhibit much steeper curves than do fine-grained soils, since the larger pores empty first and the remaining water is held by capillary and adsorptive forces as discontinuous "islands" of water at soil particle contacts (pgs. B6 through B12).
3. Sandy soils exhibit much lower water contents than clay soils at a specific potential. Expressed in a different way, clays will exhibit a much lower potential (more negative) than sands when both soils are at the same water content, (B6B, B8C, B9AC, B10A, B11AB). This implies the field capacity of a sand (see Section 3.2.3 in text) will be much lower and should be more well defined than in clay soils.
4. The relationship between potential and water content is not unique: one value of potential may correspond to several values of water content. The manifestation of this phenomenon, called hysteresis, is that the curve followed by a particular soil at a particular initial water content will differ in desorption from that in sorption, (B6C, B9A, B12, B13, B14, B15, B16). These data also show that a soil will not follow the main wetting or drying curve if it is initially at a water content other than 0 or 100% saturation, but will follow another curve (called a scanning curve) from its initial position toward one of the main curves (B13 through B16).
5. Typically, remolding causes the soil to retain more moisture at a particular potential than if it were undisturbed (B9B). Therefore, testing core samples in the lab may not give an accurate representation of in-situ soil behavior because of the disturbance inherent in the sampling process.
6. As the clay content of a soil increases, the water content corresponding to a particular potential also increases (B17A).

Additional water retention (Ψ vs θ) curves are presented on pages B18 through B21.

Water Content vs Hydraulic Conductivity

One important characteristic of unsaturated flow is that the hydraulic conductivity (K) is not a constant (as in saturated flow) but varies with water content (θ). In addition, the relationship between K and θ is nonlinear. Since the flow rate depends upon the hydraulic conductivity, this relationship must be known in order to solve the flow equation.

The K vs θ curves presented herein illustrate the following characteristics of hydraulic conductivity:

1. In general, hydraulic conductivity decreases as the water content decreases. Mehuys et al. (1975) indicate that in the low water content range a 10% change in water content would change the conductivity by $3\frac{1}{2}$ to $4\frac{1}{2}$ orders of magnitude (B22). Very few experimental data exist for very low water contents (i.e., 5% by volume), such as may be found at arid site vadose zones. As will be discussed later, one can more easily measure diffusivity at low θ and then calculate K indirectly (B23).
2. The hydraulic conductivity of soil at very low water contents becomes very small (B23 through B32).
3. The $K(\theta)$ relationship is less hysteretic than $\Psi(\theta)$ (B27CD, B29C, B33C).
4. Predictions of $K(\theta)$ from other soil properties, such as pore-size distribution, are usually close to measured values on sands but fail for more clayey soils (B33AB, B34, B35).
5. The effect of gravel particles (called stones in the literature) on conductivity of soils was examined by Mehuys et al. (1975), B22. The actual gradation of the soil was not given, only that from 10% to 20% of the particles were greater than 9.5 mm and 60% to 70% were less than 2 mm. They concluded that gravel has a definite effect on $K(\theta)$ but when water content determinations were corrected for the presence of gravel, there was good agreement between tests with and without gravel particles. This analysis may be useful when determining the relationship of K vs θ of desert soils which contain gravel.

Water Content vs Diffusivity

The mathematical and experimental treatment of unsaturated flow can be simplified by using equations which are analogous to those of heat diffusion. Such equations, called θ -based herein, require measurement of soil-water diffusivity, D , which varies with water content. The diffusivity function is much easier to determine at low water contents than the hydraulic conductivity. Hence, diffusivity can be used either to solve the flow equation directly or to extend the K vs θ relationship into the low water content range. However, due to the difficulty of measuring small flows, there are not enough data available to verify the applicability of currently proposed flow equations at very low water contents.

The D vs θ curves presented herein illustrate the following characteristics of diffusivity:

1. At high water contents (i.e., approaching saturation) diffusivity becomes indeterminate since it tends to infinity (B36, B37).
2. Diffusion, for the most part, is less sensitive to changes in water content than is the hydraulic conductivity (B36, B37, B38). For example, the data of Gupta and Staple (1964) indicate that a 10% change in water content will result in less than an order of magnitude change in diffusivity (rather than the $3\frac{1}{2}$ to $4\frac{1}{2}$ orders of magnitude change in hydraulic conductivity mentioned previously), (B36A).
3. In the dry range of some soils, there exists a water content at which diffusion reaches a maximum value (B38ABC). This point is discussed further in Appendix E.

At water contents below field capacity, the diffusion coefficient is much easier to measure than is the hydraulic conductivity coefficient. Also, several empirical relationships between D and θ which have been proposed aid in computer modeling of unsaturated, nonhysteretic flow under isothermal conditions.

Hydraulic Conductivity vs Potential

Typically, the relationship between hydraulic conductivity, K , and potential, Ψ , is determined indirectly from K vs θ curves and θ vs Ψ curves. If hydraulic conductivity can be measured

directly as a function of matric potential, the highly hysteretic θ vs ψ need not be used to calculate K vs ψ . However, $K(\psi)$, itself, is subject to hysteresis but to a much lesser degree. In addition, the slope of K vs ψ is often very steep, hence K may vary several orders of magnitude within the potential range of general interest.

Nonetheless, several researchers have reported K vs ψ some of which are reproduced herein (B39, B40, B41C). These data serve to illustrate several characteristics of unsaturated flow:

1. In general, conductivity decreases significantly as potential decreases (becomes more negative).
2. For coarse-grained soils one can approximate K vs ψ by a procedure which is based on grain-size distribution (B41A).
3. At very negative potentials (hundreds of bars) a clay soil may transmit moisture more rapidly than a sandy soil (B41C).
4. The data of Mchuys et al. (1975) indicate that the presence of gravel in a soil does not seem to affect the unsaturated conductivity (as compared to the same soil minus the gravel portion), (B39).

A Table 1—Texture and water retention

Soil classification	Mechanical analysis, %			Field capacity, % 2/3 atm	Wilting point, % 15 atm
	Sand	Silt	Clay		
Uplands sand	88.8	7.0	3.4	6.2	3.7
Castor loam	44.8	44.4	10.8	13.8	3.7
Ridgely clay	22.8	38.9	37.9	28.6	17.1

* Particle size in mm.

Table 2—Aggregate size distribution

Soil classification	Dry sieving analysis, %					
	< 0.063*	0.063-0.150	0.150-0.250	0.250-0.425	0.425-0.750	> 0.750
Uplands sand	1.3	28.9	48.8	18.8	4.4	4.4
Castor loam	18.9	9.8	5.0	3.2	14.8	35.8
Ridgely clay	28.4	22.0	15.2	8.6	12.6	7.2

* Sieve opening in mm.

Staple, 1969

B Fig. 1—Soil water tension ($-\psi$) for wetting vs water content.

Staple, 1969

C Fig. 6—Drying scanning curves for Castor loam.

Staple, 1969

D Fig. 5—Drying scanning curves for Uplands sand.

Staple, 1969

A Figure 3 Wetting curves for fine sand and silt loam
Frind et al, 1976

B Fig. 2. Moisture characteristics of five soils (after Rijtema, 1969).
Wind, 1972

NOTE: Thermistor psychrometer will measure osmotic pressure, the tensiometer will not; hence, the notation used by Kay and Low, 1970.

C Fig. 4—Moisture characteristic curve for Brockton silt loam obtained by a pressure-plate apparatus (reported as suction) and by the thermistor psychrometer (reported as total suction).
Kay & Low, 1970

B-8

A Fig. 1. Relationship between pressure potential and water content for a loam soil sample. Childs & George, 1948

B Fig. 1. Relationship between pressure potential and shrinkage curve for a loam clay block. Childs & George, 1948

C FIG. 1. RELATION OF pF, PRESSURE POTENTIAL, AND WATER CONTENT AT 20°C. FOR UNHEATED LOAM SAMPLES AND A WASHED FINE SAND Gurr et al, 1952

A Fig. 4.8. Representative water-retention curves. Relationship between potential and water content for different soils.

Yong & Warkentin, 1975

B Fig. 4.9. Effect of remoulding on water retention of a clay soil.

Yong & Warkentin, 1975

Fig. 4.12. Calculated and measured water retention for kaolinite-glass bead mixtures.

C

Yong & Warkentin, 1975

D Fig. 10—Dimensionless water content as a function of scaled pressure head. Shaded area represents data for Hanford (MF), Hanford (FF), Hanford sl, Panoche cl, Yolo cl, and Yolo sl. The straight line satisfies $h^* = -3.6 \times 10^4 \exp(-8.06 \theta)$.

Reichardt et al, 1972

E Fig. 4.15 Hysteresis and scanning curves in water retention.

Yong & Warkentin, 1975

BULK DENSITIES

of disturbed samples and method of

Pressure & Temperature	Sample depth, ft.		
	1	2	4
Yorton sandy loam	1.38	1.42	1.40
Meola sand	1.50	1.51	1.50
Whitesand sandy gravel	1.55	1.56	1.61

SATURATED HYDRAULIC CONDUCTIVITIES

Exp. method	Constant head	Fall permeameter
Yorton sandy loam		7.2×10^{-3}
Meola sand	0.26	0.01
Whitesand sandy gravel		0.60

- At 30ft depth at Site B
- At 22ft depth at Site 2 miles from B
- At 16ft depth at Site C

A Fig. 7. Functional relationships between pressure head ψ and moisture content.

Freeze & Banner, 1970

B-11

A FIG. 1. MOISTURE RELEASE DATA TAKEN FROM THE CURVES IN FIGURE 3 BUT PLOTTED ON DIFFERENT SCALE.

L.A. Richards, 1949
FIELD CAPACITY

B FIG. 2. MOISTURE RELEASE CURVES FOR FIVE SOILS OVER 0.0 ATMOSPHERE TENSION RANGE.

L.A. Richards, 1949

C FIG. 3. MOISTURE RELEASE CURVE FOR CHINO CLAY AS OBTAINED ON POROUS MEMBRANE.

L.A. Richards, 1949

Ground-water Recharge and Discharge

A Fig. 2. Functional relationships between soil moisture-tension head (ψ), moisture content (θ) and hydraulic conductivity (K) for Del Monte sand, Rehovot sand, and Greenville silt loam. Freeze, 1969

B Figure 4.—Moisture content vs. suction curves, desorption and absorption, for Ida silt loam used in theory calculations of infiltration. Green et al, 1964

C Fig. 1. Relations between ψ and θ and between K and θ of the used sand (experimental data from Vachaud and Thony (1971)).

Giesel et al, 1973

Bear et al, 1968

D Fig. 9. Hyet-resis loops (after Poulavassili (1962)).

A Fig. 1 Hydrologic characteristics of no. 17 sand
Perrins & Watson, 1977

B Fig. 2 Hydrologic characteristics of Rubicon sandy loam.
Perrins & Watson, 1977

C FIG. 3—Soil water characteristic curves for Yolo sl, Panoche cl, Hanford (MF), Hanford sl, Hanford (FF), and Yolo cl. Dimensionless water content $\theta = (\theta - \theta_u) / (\theta_l - \theta_u)$.

Reichardt, 1972

Fig. 4.14. Soil-water-energy characteristic surface. (Yong et al., 1971)

D Yong & Warkentin, 1975

A Fig. 2—Experimental data for water content-pressure head hysteresis. A single main branch loop and rewet scanning curves from three tensiometers (1, 3, 4). The scanning curves resulted from three rewet-redry cycles. The filled square symbols represent the rewet phase of one cycle, the filled triangle and circle symbols a second cycle, the open symbols a third cycle.

Topp, 1969

B Fig. 3—Experimental data for water content-pressure head hysteresis. The same main branch loop as Fig. 2 and the redry scanning curves from the three tensiometers. The filled and open symbols represent the redry portions of two separate rewet-redry cycles.

Topp, 1969

Fig. 7—Comparison between experimental (short dashed) and theoretical (long dashed) redry (θ - h) scanning curves. Theoretical curves resulted from applying the independent domain theory to redry scanning curves.

Topp, 1969

Fig. 8—Comparison between experimental (short dashed) and theoretical (long dashed) rewet (θ - h) scanning curves. Theoretical curves resulted from applying the independent domain theory to rewet scanning curves.

Topp, 1969

A Fig. 2—Soil moisture absorption curve for Greenville silt loam; open circles from fritted-glass funnels; dots from tensiometers in diffusivity experiments; open and solid triangles from capillary rise and conductivity experiments, respectively.

Gupta & Staple, 1964

B Fig. 3—Water content-pressure head data for wetting scanning curves as determined at position 15 in the column.

Gillham et al, 1976

C Fig. 4—Water content-pressure head data for drying scanning curves as determined at position 15 in the column.

Gillham et al, 1976

D Fig. 5—Water content-pressure head relations, main drying and main wetting curves for two positions.

Gillham et al, 1976

A Fig. 4.11. Water content, w , of Ca-kaolinite-glass bead mixtures at various suction values given in bars on drying run. (De Jong and Warkentin, 1965, *Soil Sci.*, © Williams and Wilkins Co., Baltimore.) Yong and Warkentin, 1975

Figure 5.2 Moisture Potential vs Water Content

Jury, 1973

Fig. 1. Moisture characteristic curves for 50- to 500-mesh sand measured on the same columns as was the capillarity data in Figures 2 and 3.

Jackson et al, 1965

Fig. 5. Desorption moisture characteristic curves for Adelfanto loam and Pachappa loam.

Jackson et al, 1965

Figure 1—The dependence of Rehovot sand's moisture content and hydraulic conductivity upon matric suction. The calculated curves are based on equations [1] and [2]. Rubin et al, 1964

Fig. 1—Moisture retention characteristics of Plainfield sand for 0-25-cm, 25-60-cm, 60-100-cm depths.

Black et al, 1969

Fig. 1—Soil water content, by weight, as a function of soil water suction determined from sieved and core samples of Ap horizon of Cecil loamy coarse sand.

Bruce, 1972

Fig. 2—Soil water content, by weight, as a function of soil water suction determined from sieved and core samples of B2 horizon from Cecil loamy coarse sand.

Bruce, 1972

Fig. 4—Soil water desorption curves for principal horizons of Cecil loamy sand.

Fig. 3—Soil water content on a volume basis as a function of soil water suction determined from sieved and core samples of Ap, B2, and B3 horizons.

Bruce, 1972

HYDRAULIC CONDUCTIVITY

VS

WATER CONTENT

A Fig. 6—Relation between hydraulic conductivity and volumetric water content for Rock Valley gravelly loamy sand with and without stones. Curves represent the average of replicates. Mehuis *et al.*, 1975

B Fig. 7—Relation between hydraulic conductivity and volumetric water content for Tubac gravelly sandy loam with and without stones. Mehuis *et al.*, 1975

C Fig. 8—Relation between hydraulic conductivity and volumetric water content for Rillito gravelly sandy loam with and without stones. Mehuis *et al.*, 1975

Figure 5.4 Hydraulic Conductivity vs Water Content
Jury, 1973.

A FIG. 4. The relationship between the hydraulic conductivity at 20°C. and moisture content of a slate dust: (open circles) suction method; (closed circles), pressure method; (half-closed circles) measurements reported by Youngs (7), and (open squares) measurements by the method of Childs and Collis-George (1). (Circles and full line thus indicate wetting values, and squares and broken line, draining values.) Youngs, 1964

B FIG. 9.6. Unsaturated hydraulic conductivity K as a function of moisture concentration C with stable aggregates of 0.25-0.5 mm diameter. Bear et al., 1968.

C FIG. 7. The water content-instantaneous hydraulic conductivity relation showing the computed values. Watson, 1966.

Fig. 3a
Beit Netofa clay

Fig. 3b
Silt Mont Cenis

Fig. 3c
Granville silt loam

Fig. 3d
Ida silt loam (0-15 cm)

Fig. 3e
Indio loam

Fig. 3f
Guelph loam

Fig. 3g
Radeau clay loam

Fig. 3h
Caribou silt loam

Fig. 3i
Adelanto loam

Fig. 3. Computed relative hydraulic conductivities for 16 soils using four models: Averjanov's (Y symbols; see (1)), WG's (pluses; see (28)), MQ's (solid triangles; see (29)), and the proposed model (solid circles; see (27)). The measured curve is shown by solid squares. Y. Maulem, 1976

$$K_r = K/K_0$$

K_0 = Saturated Hydraulic Conductivity

For complete description of soil types, see Maulem, 1976

Fig. 3j

Screened River sand

Fig. 3k

Volcanic sand

Fig. 3l

Molongo River sand

Fig. 3m

Sand fraction (150-300 μ m)

Fig. 3n

Sand fraction (0.5-0.25 mm)

Fig. 3o

Aggregated glass beads

Fig. 3p

Monodispersed glass beads

Y. Maulem, 1976

A Fig. 3—Hydraulic conductivity K for wetting vs water content. Solid lines from measurements of D and K . Dashed lines computed from redistribution moisture profiles. Staple, 1969

B Fig. 4—Hysteresis as measured in the hydraulic conductivity-water content relationship. Topp, 1969

C Fig. 4—The hydraulic conductivity-water content relationship for four levels during the initial rewet phase of the soil column. The solid lines bound all data measured during succeeding rewet-redry cycles. Topp, 1969

D Fig. 5—The decrease in hydraulic conductivity associated with the initial redrying of the sample. Topp, 1969

Fig. 5—Hydraulic conductivity as a function of water content calculated by several procedures from the plotted desorption data of the sand in comparison to measured values.

Bruce, 1972

Fig. 6—Hydraulic conductivity as a function of water content calculated by several procedures from desorption data for graded sand (upper curves) and a fine sand (lower curves) in comparison to their measured hydraulic conductivity.

Bruce, 1972

Fig. 7—Hydraulic conductivity of Ap horizon core samples of Cecil loamy sand calculated by several procedures over a range of water contents in comparison to measured values.

Bruce, 1972

Fig. 8—Hydraulic conductivity of B2 horizon core samples of Cecil loamy sand calculated by several procedures over a range of water contents in comparison to measured values.

Bruce, 1972

A Fig. 9—Hydraulic conductivity of B3 horizon core samples of Cecil loamy sand calculated by several procedures over a range of water contents in comparison to measured values. Bruce, 1972

B Fig. 10—Hydraulic conductivity calculated by Marshall (m) method from three versions of the desorption curve between 0 and 15 mbars for B2 horizon core samples of Cecil loamy sand in comparison to measured values. Bruce, 1972

C Fig. 6—Hydraulic conductivity-water content relationship at position 15. Gillham et al., 1976

B-30

Fig. 1—Calculated hydraulic conductivities for Guelph loam using three methods of computation compared with experimental data from Elrick and Bowman (1964, their Fig. 3). Calculated and experimental data were matched at θ_s .

A Green and Corey, 1971

Fig. 2—Calculated hydraulic conductivities for monodispersed glass beads using three methods of computation compared with experimental data from Topp and Miller (1966, their Fig. 2c). Calculated and experimental data were matched at θ_s . Green and Corey, 1971

Fig. 3—Calculated hydraulic conductivities for Rubicon sandy loam (desorption characteristic) using three methods of computation compared with experimental data from Topp (1969, his Fig. 4). Calculated and experimental data were matched at θ_s . The experimental data are the upper and lower limits obtained along a soil column during desorption studies. Green and Corey, 1971

Fig. 4—Calculated hydraulic conductivities for Rubicon sandy loam (absorption characteristic) using three methods of computation compared with experimental data from Topp (1969, his Fig. 4). Calculated and experimental data were matched at θ_s . The experimental data are the upper and lower limits obtained along a soil column during absorption studies. Green and Corey, 1971

1—Comparison of measured and calculated conductivities for aggregated glass beads using 2, 8, 32, 128, and 512 pressure classes and no matching factor. Kunze et al., 1968

1968 A

B Fig. 2—Comparison of measured and calculated conductivities for monodispersed glass beads using 2, 8, 32, 128, and 512 pressure classes and no matching factor.

Kunze et al., 1968

C Fig. 4—Comparison of measured and calculated conductivity for Ida soil. Kunze et al., 1968

D Fig. 5—Comparison of measured and calculated conductivity for Webster soil. Kunze et al., 1968

A Fig. 6—Comparison of measured and calculated conductivity for Colo soil. Kunze et al., 1968

B Fig. 5—Variation of capillary conductivity coefficients with moisture content for Granville silt loam; squares and circles computed from diffusivity; cross and triangles from steady flow experiments. Gupta and Staple, 1964

A Fig. 2. Comparison of measured and calculated values of hydraulic conductivity of 50- to 300-micron sand for desorption.

Jackson et al., 1965

B Fig. 3. Comparison of measured and calculated rates of hydraulic conductivity of 50- to 300-micron sand for sorption. Jackson et al., 1965

C Fig. 4. Hydraulic conductivity of Garph loam; the line was calculated from Figure 2 of Elrick & Bowman (1961) and the symbols are reproduced from their Figure 3.

Jackson et al., 1965

D Fig. 6. Hydraulic conductivity of Adelanto and Pachappa loam, measured and calculated non characteristic curves.

Jackson et al., 1965

Figure 3—Calculated values of capillary conductivity for four depths of Ida soil plotted against measured values. Different symbols indicate different soil moisture suctions. Solid symbols are used for values calculated by Marshall's method; open symbols for Childs and Collis-George's (CCG) method. If the theories were correct and the experimental measurements accurate, the plotted points should lie on the 45° dashed line. The arrows indicate where a fitting factor m in the CCG method was determined. Nielsen *et al.*, 1960 A

Figure 4—Same as figure 3, except for Monona soil. B Nielsen *et al.*, 1960

Figure 5—Same as figure 3, except that the soil is Floyd and an extra ordinate scale has been added at the right of each subfigure to apply to the lower (CCG) curves. Nielsen *et al.*, 1960

Table 1—Textural percentages; also, water content θ vs. capillary conductivity K for five values of suction on core samples from several depths of four Iowa soils. Each value in the table is the average of four samples.

Soil and depth (inches)	Textural percentages			Water content ($\frac{\text{cm}^3}{\text{cm}^3}$) - capillary conductivity ($\frac{\text{cm}}{\text{hr}}$)											
	Clay	Silt	Sand	θ_1	θ_2	θ_3	θ_4	θ_5	θ_6	θ_7	θ_8	θ_9	θ_{10}	θ_{11}	θ_{12}
Ida 24	20	80	15	0.211	7.39	0.464	3.07	0.441	3.15	0.403	0.25	0.387	0.81		
Ida 30	23	82	15	0.200	6.79	0.481	2.14	0.419	0.50	0.395	0.34	0.375	0.75		
Ida 36	22	82	18	0.199	6.99	0.481	4.23	0.439	1.33	0.439	0.30	0.405	0.85		
Ida 42	18	86	18	0.205	2.08	0.679	2.35	0.480	1.78	0.443	0.39	0.435	0.87		
Monona 12	20	80	15	0.172	11.10	0.402	1.89	0.351	0.30	0.346	0.29	0.322	3.43		
Monona 24	27	59	14	0.325	10.53	0.441	1.88	0.401	0.22	0.378	0.18	0.351	3.79		
Monona 36	22	82	15	0.325	7.77	0.477	0.06	0.442	0.12	0.412	0.27	0.385	0.16		
Monona 48	20	84	16	0.309	6.10	0.483	2.66	0.458	1.70	0.433	0.36	0.408	2.38		
Floyd 12	24	42	34	0.155	1.02	0.608	0.0050	0.381	0.0024	0.390	0.0018	0.374	5.812		
Floyd 24	25	45	30	0.249	20.88	0.309	0.030	0.294	0.0085	0.294	0.0019	0.279	0.0009		
Floyd 36	23	12	67	0.249	0.42	0.611	0.255	0.0019	0.258	0.0006	0.255	0.0002	0.0002		
Floyd 48	20	21	49	0.212	0.22	0.303	0.0044	0.325	0.00078	0.289	0.00032	0.280	0.0002		
Webster 12	25	35	32	0.208	32.77	0.408	0.380	0.0025	0.387	0.00049	0.361	0.00015	0.00015		
Webster 24	28	31	31	0.187	18.21	0.384	0.0920	0.308	0.00023	0.350	0.00003	0.358	0.00002		
Webster 36	28	73	21	0.288	7.38	0.334	0.81	0.223	0.00872	0.323	0.00015	0.280	0.0001		
Webster 48	24	16	60	0.248	8.71	0.350	0.488	0.330	0.078	0.330	0.0000	0.330	0.0001		

* Suction (K) indicates cm. water suction applied.
† Average of only two determinations.

Nielsen *et al.*, 1960

Figure 6—Same as figure 5, except for Webster soil. Nielsen *et al.*, 1960

Fig. 5—Calculated hydraulic conductivities for Ida silt loam using three methods of computation compared with experimental data from Green et al. (1964, their Fig. 3, 0–15 cm depth). Calculated and experimental data were matched at θ_s . Green and Corey, 1971 A

B Fig. 6—Calculated hydraulic conductivities for Panoche clay loam ("6 to 12-inch depth") using three methods of computation compared with experimental data from Nielsen et al. (1964, their Fig. 11). Calculated and experimental data were matched at $\theta = 0.35 \text{ cm}^3/\text{cm}^3$. Green and Corey, 1971

C Fig. 7—Calculated hydraulic conductivities for Panoche clay loam ("12 to 18-inch depth") using three methods of computation compared with experimental data from Nielsen et al. (1964, their Fig. 11). Calculated and experimental data were matched at $\theta = 0.35 \text{ cm}^3/\text{cm}^3$. Green and Corey, 1971

D Fig. 8—Calculated hydraulic conductivities for Panoche clay loam ("18 to 24-inch depth") using three methods of computation compared with experimental data from Nielsen et al. (1964, their Fig. 11). Calculated and experimental data were matched at $\theta = 0.35 \text{ cm}^3/\text{cm}^3$. Green and Corey, 1971

DIFFUSIVITY
VS
WATER CONTENT

B-36

A
Fig. 4—Variation of diffusion coefficients with moisture content for Grenville silt loam. Gupta and Staple, 1964.

B
Fig. 3—Soil-water diffusivity as a function of volumetric water content for the 0-25-cm depth of Plainfield sand. Black et al, 1969.

HILLEL AND GARDNER

C
Fig. 4. Relation of diffusivity to water content for Gilat loam. Hillel and Gardner, 1970

A 1. Relation of moisture content, w , to pressure head, h , for a soil.
Childs and George, 1948

B Table 1—Relation of moisture content, θ , to moisture diffusivity D , and pressure head, h .

θ Vol. fraction	Sandy loam		Gravy silt loam	
	D Cm. ² /sec.	h Cm. of H ₂ O	D Cm. ² /sec.	h Cm. of H ₂ O
0.05	1.45×10^{-4}	-6975	-	-
0.06	1.49×10^{-4}	-5165	-	-
0.07	1.88×10^{-4}	-2120	-	-
0.08	3.40×10^{-4}	-1255	-	-
0.09	7.94×10^{-4}	-659	-	-
0.10	8.26×10^{-4}	-417	-	-
0.11	1.30×10^{-3}	-362	-	-
0.12	1.36×10^{-3}	-330	-	-
0.13	1.37×10^{-3}	-289	-	-
0.14	1.44×10^{-3}	-275	-	-
0.15	1.51×10^{-3}	-231	-	-
0.16	2.07×10^{-3}	-209	-	-
0.17	2.35×10^{-3}	-197	-	-
0.18	2.58×10^{-3}	-168	1.6×10^{-4}	-7685
0.19	3.28×10^{-3}	-151	3.6×10^{-4}	-5635
0.20	5.75×10^{-3}	-131	5.6×10^{-4}	-4025
0.21	1.04×10^{-2}	-120	7.6×10^{-4}	-3190
0.22	1.35×10^{-2}	-106	9.5×10^{-4}	-2675
0.23	1.51×10^{-2}	-92	1.28×10^{-3}	-2175
0.24	1.59×10^{-2}	-78	1.92×10^{-3}	-1675
0.25	6.05×10^{-2}	-63	2.29×10^{-3}	-1175
0.26	4.78×10^{-2}	-61	3.61×10^{-3}	-915
0.27	5.05×10^{-2}	-58	4.72×10^{-3}	-665
0.28	5.14×10^{-2}	-53	5.31×10^{-3}	-535
0.29	5.43×10^{-2}	-47	7.35×10^{-3}	-433
0.30	5.73×10^{-2}	-43	1.02×10^{-2}	-331
0.31	5.32×10^{-2}	-39	1.15×10^{-2}	-259
0.32	5.82×10^{-2}	-34	1.22×10^{-2}	-212
0.33	7.52×10^{-2}	-30	1.50×10^{-2}	-175
0.34	8.89×10^{-2}	-26	1.78×10^{-2}	-143
0.35	1.25×10^{-1}	-22	2.12×10^{-2}	-116
0.36	1.74×10^{-1}	-18	2.74×10^{-2}	-94
0.37	4.50×10^{-1}	-14	2.93×10^{-2}	-75
0.38	4.83×10^{-1}	-10	3.32×10^{-2}	-59
0.39	4.07×10^{-1}	-6	3.41×10^{-2}	-45
0.40	4.42×10^{-1}	-3	3.85×10^{-2}	-36
0.41	4.18×10^{-1}	0	4.25×10^{-2}	-28
0.42	-	-	4.32×10^{-2}	-21
0.43	-	-	4.77×10^{-2}	-15
0.44	-	-	4.92×10^{-2}	-10
0.45	-	-	4.86×10^{-2}	-5
0.46	-	-	4.86×10^{-2}	0

Ranks and Bowers, 1962

A Fig. 3—Diffusion coefficients for Adelanto loam at three sets of boundary and initial conditions. Each symbol indicates the average of two measurements. R. D. Jackson, 1964.

B Fig. 4—Diffusion coefficients for Pine silty clay, as a function of gravimetric water content. R. D. Jackson, 1964.

C Fig. 5—Diffusion coefficients for Pachappa loam as a function of gravimetric water content. R. D. Jackson, 1964.

D Figure 6—Distance to the wetting front vs. the square root of time for Pachappa loam, porosity 0.500; Adelanto loam, porosity 0.487; and Pine silty clay, porosity 0.510. The different symbols on each line are data for duplicate columns. Jackson, 1963.

E Figure 1—Soil-water diffusivity vs. relative water content at three porosities: A—Adelanto loam; B—Pachappa Loam and C—Pine silty clay. The broken lines are the least squares fit to the data at the lowest porosity. Jackson, 1963.

HYDRAULIC CONDUCTIVITY

VS

SOIL MOISTURE POTENTIAL

Fig. 3—Hydraulic conductivity as a function of matric potential for stony and non-stony samples of Rock Valley gravelly loamy sand. Curves represent the average of replicates. Mehuijs *et al.*, 1975

Fig. 4—Hydraulic conductivity as a function of matric potential for stony and non-stony samples of Tubac gravelly sandy loam. Mehuijs *et al.*, 1975

Fig. 5—Hydraulic conductivity as a function of matric potential for stony and non-stony samples of Rillito gravelly sandy loam. Mehuijs *et al.*, 1975

Fig. 6—Capillary conductivity as a function of soil suction for the 25-60 cm depth of Plainfield sand. Black *et al.*, 1969

A Figure 5.3 Hydraulic Conductivity vs Moisture Potential
Jury, 1973

A Fig. 1. Examples of $k(\psi_s)$ relations for five soils (after Rijtema, 1969). Wind, 1972

B Fig. 9.3. Unsaturation hydraulic conductivity for aggregates of different sizes. Bear et al., 1968

C Fig. 5.4. Dependence of conductivity on suction in soils of different texture (log scale). Hillel, 1971

APPENDIX C
INFILTRATION

APPENDIX C

Most research into the behavior of moisture in unsaturated porous media has been related to the infiltration of rain to the soil profile. For convenience of discussion, the data obtained from the literature are divided into four categories:

1. Water Content or Potential vs Depth
2. Rate of Infiltration and Wet Front Advance
3. Effect of Layering
4. Other Factors Which Influence Infiltration

Water Content or Potential vs Depth

Both laboratory and field experiments have been performed wherein water was applied to the soil surface in the form of rain and measurements were made of water content, θ , and soil water potential, ψ , as a function of depth. Typically, a moisture profile such as noted by Bodman & Colman (1943), will develop (C7C). The curves of water content vs depth illustrate the following:

1. Infiltration into a homogeneous soil profile can be characterized by the following zones which are listed in order of increasing depth:

Saturated zone - the surface few centimeters become saturated and will not increase in vertical extent as long as the rain intensity, R , is less than the saturated hydraulic conductivity, K_o , of the soil.

Transmission zone - the water content decreases rapidly with depth until it reaches a value somewhere between field capacity and saturation and remains relatively constant as the transmission zone increases in vertical extent with the advancement of the wetting zone.

Wetting zone - the soil moisture content gradually decreases with depth in the wetting zone wherein the moisture content of the soil increases rapidly with time as infiltration continues.

Wet front - the wetting zone stops abruptly at the wet front which marks the lower extent of infiltration and is characterized by a very steep gradient of moisture content.

2. The shape of the water content vs depth curve varies with soil type (C8 to C14). Generally, the coarser the material, the steeper the gradient of moisture content with depth at the wet front.
3. The amount of water applied to the surface will affect the depth to which the wet front will advance after time, t (C8, C9, C10, C11, C13B, C15, C16). In general, the higher the application rate, the further the wet front will advance in a given time period.

Examination of the potential vs depth indicates the following:

1. Initially, infiltration induces a very steep gradient of potential with depth which, with time, approaches a gradient of 1 (i.e. gravity becomes the only remaining driving force; (C13A, C14, C17, C18).
2. As with the water content profiles, potential profiles vary with soil type (C13A, C17).

Rate of Infiltration and Wet Front Advance

The rate of infiltration, v_o , is the quantity of water entering the soil profile through a unit area at the soil surface per unit time.

The rate of infiltration, v_o , vs time (for a constant surface supply of water) is illustrated by the data in Figs. C19A, C20A, and C20B. These data indicate that with time, the infiltration rate, v_o , approaches a constant value equal to the saturated hydraulic conductivity, K_o . Concurrently, the total amount of water entering the profile, i , (called cumulative infiltration) approaches a linear relationship with time (C19A and C20C).

One may also consider the rate at which the soil profile is wetted. This rate can be expressed as the rate of advance of the wet front, u , which, due to the storage ability of the soil, is usually different from v_o . Figures C19B, C21A, C21B,

C21C, C22A and C22B indicate that the advance of the wet front in a stable (i.e., non-swelling) soil is proportional to the square root of time. In swelling soils, such as clay, the relationship is not linear; rather, as time increases, the rate of advance of the wet front decreases (C22B and C22C).

It is important to note that these data are for relatively short time spans, a limited range of soil types, and shallow depths of infiltration.

If these data can be extrapolated to longer times, to greater depths, and to other soils, the aforementioned phenomena have potential significance in the feasibility of storing nuclear waste in the vadose zone. For example, if surface water is ponded for an extended period of time, a saturation zone may be created which extends from the surface, through the repository, down to the water table. Once the entire soil column is saturated, water will drain down to the groundwater table at a rate⁽¹⁾ equal to the saturated hydraulic conductivity, K_0 , assuming a hydraulic gradient of one (Philip, 1957f).

Effect of Layering

Adjacent layers of soils with different $\Psi(0)$ relationships (i.e., different grain size distributions) will cause discontinuities in water content vs depth (or distance from source) curves (C23AB, C24, C25). However, at equilibrium, the Ψ vs Z or Ψ vs X curves are continuous across the boundaries (in accordance with the Second Law of thermodynamics).

Examination of Figure C23 reveals several characteristics of unsaturated flow through layered porous media:

1. Flow rate is controlled by the layer with the lowest hydraulic conductivity, regardless of the sequence of layers in the soil profile. This point is illustrated by Fig. C23C in that, no matter which layer is on top, the infiltration rate curve joins that of Geary or the cumulative infiltration curve parallels that of Geary (the soil with the lower hydraulic conductivity).

(1) The term rate, with units of l/t , equals the volume of water that moves through a unit area per unit time. Rate, therefore, corresponds to seepage velocity. The actual flow velocity equals the seepage velocity divided by the effective porosity.

2. When the finer soil overlays the coarser soil (i.e., Geary over Sarpy), the wet front stops at the boundary until the water content is increased enough to increase the potential (reduce the suction) sufficiently to permit entry of water into the larger pores of the coarser material.
3. Downward unsaturated (i.e., flow under a negative potential) may occur in the soil profile even if an overlying coarser layer is saturated. The 109 min. curve in Fig. C23A shows a positive potential in the Sarpy (coarser) layer but a negative potential in the Geary (finer) layer slightly below the boundary between the two soils.
4. At very negative values of potential (B41C) a fine-grained soil will transmit water in the unsaturated state faster than a coarse-grained soil. Figs. C24A and C25AD illustrates this phenomenon by showing where the wet front would have been if the soil were homogeneous rather than layered. In Fig. C24A the wet front of 121 min. has penetrated further into the FF soil than it would have if the entire column were homogeneous MF soil. Conversely, the 196 min. curve in Fig. C25A has not penetrated as far into the MF soil as it would have if the column were homogeneous FF soil.

These characteristics are advantageous in the design of vadose zone repositories. For example, if the waste were placed in a cavern backfilled with coarse material (e.g., crushed stone) overlain by layers of successively finer soil, the layers above the coarse repository backfill would act as "wicks" (Richards, 1950) causing the water to flow around the repository rather than through it (C33A).

Other Factors Which Influence Infiltration

Other factors, in addition to layering, such as initial water content, θ_n , depth of surface ponding, h , and rain intensity, R , may influence the characteristics of infiltration. These factors are discussed here.

1. Initial water content (Philip, 1957e)

- infiltration rate, v , increases as initial water content, θ_n , decreases (C26A).
- cumulative infiltration (total amount of water entering soil profile), i , increases as θ_n decreases (C26B).

- the "long time" moisture profile will be shallower as θ_n decreases (C26C).
 - the rate of wet front advance, u , will decrease as θ_n increases (C27D).
2. Depth of water ponded on surface, h , (Philip, 1957e, f)
- initially, v_o increases with h but as time increases, the effect of h diminishes and v_o approaches K_o (C27A).
 - initially, i increases with h but as time increases, the curves of i vs t become parallel since v_o approaches K_o as t increases (C27B).
 - for $h > 0$, x_o (the depth of saturation) increases without limit; hence, the effect of h on x_o persists with time (in contrast with the diminishing effect with time of h on v_o) (C27C).
 - in the limit as h increases or as t increases (with $h > 0$) the saturated zone dominates the profile.
 - for a given h and t , the depth of the saturated zone, x_o , decreases as θ_n decreases (Probably because there is greater void volume available for storage).
 - the relative effect of h on v_o decreases as θ_n decreases.
3. Rain intensity, R (C28 BDE)
- the rate of wet front advance, u , increases as rain intensity increases.
 - the depth of infiltration increases as rain duration increases.
4. Miscellaneous
- gravity effects vary with soil type: Hanks and Bowers (1962), C28C, showed gravity affected the coarse soil more than the fine soil.

- the effect of soil type on infiltration to a shallow water table is illustrated by Fig. C29A (the moisture retention curves for the three soils can be found in Appendix B on page B-11). It is useful to note that the infiltration curves are dependent on the moisture characteristics of the soil, i.e., rain intensity, R , relative to the saturated hydraulic conductivity, K_0 .
- the effect of confinement of swelling soils is illustrated by Fig. C30A: as the soil tries to swell, pore spaces are closed and, due to confinement, less water enters the soil; therefore flow rate is reduced.
- the effect of a soil surface crust is illustrated by Figs. C30B and C31A: the presence of a crust (a thin surface layer of soil which has a lower hydraulic conductivity than the bulk portion of the soil profile) rate of infiltration to the soil profile. Similarly, there is less cumulative infiltration with time in soils with a surface crust.
- the data of Freeze (1969), C32B and C, illustrate the effect of groundwater conditions on infiltration through soils in areas with shallow water tables.

A FIG. 2. The moisture profile development during the experiments: (a) suction method with an applied suction of 70 cm. of water, and (b) pressure method with an applied air pressure of 45 cm. of water; numbers near profiles refer to sampling time in hours. Youngs, 1964

B FIG. 7. COMPUTED MOISTURE PROFILES DURING INFILTRATION INTO YOLO LIGHT CLAY FOR LARGE TIMES, (equation 52)
Numbers on each profile represent time in seconds at which profile is realized
J. R. Philip, 1957b

C FIG. 5.34. Water content vs. depth for two times during infiltration. (From Bodman and Coleman, 1943. *Soil Sci. Soc. Am. Proc.* 8.) Yong and Warkentin, 1975 with Revision by GEI, 1978

A Fig. 3. Soil water profiles during the redistribution of water in Hanford sand. The parameter near each curve represents the number of hours or days after the cessation of infiltration. Figures *a* and *b* correspond to surface water applications of 3.8 and 7.3 cm, respectively. Zero water content corresponds to an initial value of 0.005 g/cm^3 before infiltration. Values of soil water content in this figure and subsequent figures are numerically identical to θ , assuming that the density of the soil water is 1 g/cm^3 . Biswas *et al.*, 1966

B Fig. 4. Soil water profiles during the redistribution of water in Hanford sand. The parameter near each curve represents the number of hours or days after the cessation of infiltration. Figures *a* and *b* correspond to surface water applications of 10.1 and 14.3 cm, respectively. Zero water content corresponds to an initial value of 0.005 g/cm^3 before infiltration. Biswas *et al.*, 1966

A Fig. 5. Soil water profiles during redistribution of water in Columbia silt loam. The parameter for each curve represents the number of hours or days after the cessation of infiltration. The profiles stem from a surface water application of 6.1 cm. Zero water content corresponds to an initial value of 0.02 g/cm³ before infiltration. Biswas et al., 1966

B Fig. 6. Soil water profiles during redistribution of water in Columbia silt loam. The parameter for each curve represents the number of hours or days after the cessation of infiltration. Figures a and b correspond to surface water applications of 31.9 and 17.9, respectively. Zero water content corresponds to an initial value of 0.02 g/cm³ before infiltration. Biswas et al., 1966

A Fig. 7. Soil water profiles during redistribution of water in Aiken clay loam. The parameter near each curve represents the number of hours or days after the cessation of infiltration. The profiles stem from a surface water application of 7.1 cm. Zero water content corresponds to an initial value of 0.03 g/cm³ before infiltration. Biswas *et al.*, 1966

B Fig. 8. Soil water profiles during redistribution of water in Aiken clay loam. The parameters near each curve represent the number of hours or days after the cessation of infiltration. Figures a and b correspond to surface water applications of 13.7 and 20.6 cm, respectively. Zero water content corresponds to an initial value of 0.03 g/cm³ before infiltration. Biswas *et al.*, 1966

A FIG. 1.—Distribution of soil moisture in relation to depth of wetting during infiltration into Yolo silt loam. Infiltration times are indicated by numbers on curves. Time for curve 1 = 12.3 minutes; for curve 2 = 76 minutes; for curve 3 = 173 minutes; for curve 4 = 280 minutes; for curve 5 = 615 minutes; for curve 6 = 1,020 minutes.

Bodman and Colman, 1943

B FIG. 2.—Distribution of soil moisture in relation to depth of wetting during infiltration into Yolo sandy loam. Infiltration times are indicated by numbers on curves. Time for curve 1 = 8.6 minutes; for curve 2 = 20.6 minutes; for curve 3 = 33.6 minutes; for curve 4 = 72 minutes; for curve 5 = 125 minutes. Bodman and Colman, 1943

A FIG. 6.—MOISTURE CONTENT AS A FUNCTION OF HEIGHT ABOVE THE DATUM AND OF TIME DURING THE REDISTRIBUTION OF INFILTRATED RAINFALL Remson et al., 1967

A FIG. 2. — Head, depth, and time relationships for Hesperia sandy loam, Yolo loam, and silica flour.
Miller and Richard, 1952

A Fig. 1. Water profiles during constant flux infiltration and redistribution. (a) Water content profiles. (b) Water suction profiles. The lines represent computed data, and the signs represent experimental data: inverted triangles, $t = 0$ hours; open circles, $t = 0.5$ hour; squares, $t = 1$ hour; triangles, $t = 1.5$ hours; solid circles, $t = 2$ hours; solid circles, $t = 3$ hours. Giesel *et al.*, 1973

B Fig. 2. Water profiles during constant head infiltration and redistribution. (a) Water content profiles. (b) Water suction profiles. The lines represent computed data, and the signs represent experimental data: inverted triangles, $t = 0$ sec; open circles, $t = 1000$ sec; squares, $t = 2500$ sec; triangles, $t = 3000$ sec; solid circles, $t = 7500$ sec.

Giesel *et al.*, 1973

A Fig. 8—Soil moisture redistribution profiles for Uplands sand.
Staple, 1969

A Fig. 7—Soil moisture redistribution profiles for Rideau clay.
Staple, 1969

B Fig. 9—Soil moisture redistribution for Castor loam.
Staple, 1969

A Fig. 5 - Distribution of moisture potential in relation to depth of wetting during infiltration into Yolo sandy loam. Infiltration times are indicated by numbers on curves. Time for curve 1 = 86 minutes; for curve 2 = 206 minutes; for curve 3 = 330 minutes; for curve 4 = 72 minutes; for curve 5 = 1.5 minutes.

B Fig. 6 - Distribution of moisture potential in relation to depth of wetting during infiltration into Yolo silt loam. Infiltration times are indicated by numbers on curves. Time for curve 1 = 123 minutes; for curve 2 = 76 minutes; for curve 3 = 173 minutes; for curve 4 = 280 minutes; for curve 5 = 915 minutes; for curve 6 = 1,020 minutes.

Bodman and Colman, 1943

FIG. 6. COMPUTED PROFILES OF "TOTAL POTENTIAL" DURING INFILTRATION INTO YOLO LIGHT CLAY

"Total potential" = $(\psi - z)$. The numbers on each profile represent the time in seconds which the profile is realized. J. R. Philip, 1957c

RATE OF
INFILTRATION

A Figure 2—Comparison of the numerical method with the solution by the method of Scott et al. for infiltration rate and cumulative infiltration for a soil with $D = 1 \text{ cm}^2/\text{min}$.
Hanks and Bowers, 1962

B Fig. 6—Measured values of cumulative infiltration (points) into layers of Hanford (MF) and Hanford (FF). The lines are theoretical values based upon the solution of equation [3]. The straight solid line represents the cumulative infiltration into a homogeneous column of Hanford (MF).
Reichardt et al., 1972

A FIG. 6. COMPUTED INFILTRATION RATE CURVES FOR YOLO LIGHT CLAY

Note change of scales between inset and main diagram; points A and B of inset correspond to A and B of main diagram; the broken curve denotes the solution given by the first four terms of equation (36) of (2); BC represents $\pi_s = K_s$, the asymptote to the infiltration rate curve. J. R. Philip, 1957b

B Figure 5—Comparison of calculated and field-measured infiltration rates for Ida silt loam at three antecedent moisture levels. Green et al., 1964

C Figure 6—Comparison of calculated and field-measured cumulative infiltration for Ida silt loam at three antecedent moisture levels. Green et al., 1964

A Figure 4—Distance to the wetting front vs. the square root of time for three soil materials, each in duplicate at the highest and lowest temperatures, Jackson, 1963

B Fig. 5.11. Wetting front advance, x , vs. square root time, t , for a kaolin clay at three densities Yong and Warkentin, 1975

C Fig. 5.12. Time-wetting front profiles for a silt loam with negative intake at -2 mbars. Yong and Warkentin, 1975

A Fig.5.14. Time-wetting profiles for unsaturated flow into a low density kaolinitic. Volume decrease occurs during infiltration due to collapse of soil on wetting.
Yong and Warkentin, 1975

B Fig.5.13. Wetting front advance vs. square root of time showing influence of volume change and swelling pressure on time rate of advance profile.
Yong and Warkentin, 1975

C Fig.5.15. Time rate of advance of wetting front for unsaturated flow into a swelling soil showing difference in profile if volume change is allowed. In confined flow into the swelling soil, no volume change is allowed. Yong and Warkentin, 1975

INFILTRATION
EFFECT OF LAYERING

A Figure 5—Relation of $h - x$ and $\theta - x$ for infiltration into a layered soil, Sarpy over Geary with boundary at 11 cm. $\theta = 0.05$ for $0 < x < 11$ cm., $\theta = 0.184$ for $11 \text{ cm.} < x < 40$ cm. at $t = 0$. $\theta = 0.41$ for $x = 0$, $t > 0$. Hanks and Bowers, 1962

B Figure 6. Relation of $h - x$ and $\theta - x$ for infiltration into layered soil, Geary over Sarpy with boundary at 11 cm. $\theta = 0.184$ for $0 < x < 11$ cm., $\theta = 0.05$ for $11 \text{ cm.} < x < 40$ cm. at $t = 0$. $\theta = 0.41$ for $x = 0$, $t > 0$. Hanks and Bowers, 1962

h = soil-water potential
 Sarpy = coarser (silt loam)
 Geary = finer (loam)

C Figure 7—Influence of layer sequence on infiltration and cumulative infiltration. Hanks and Bowers, 1962

A Fig. 2—Experimental soil water contents (points), calculated soil water content distributions (large broken lines), and bulk density distributions for horizontal infiltration into a layer of Hanford (MF) followed by a layer of Hanford (FF). Calculations are based upon D^* and λ (Table 1). Small broken lines indicate where the wetting front should have been if the column were homogeneous, consisting only of Hanford (MF). Reichardt et al., 1972

B Fig. 4—Calculated pressure head distributions for horizontal infiltration into a layer of Hanford (MF) followed by a layer of Hanford (FF) corresponding to soil water content distributions given in Fig. 2. Reichardt et al., 1972

Yolo clay loam, Panoche clay loam, and Yolo silty loam were sieved to <0.5 mm, and from Hanford sandy loam three fractions were obtained by dry sieving: (i) sieved to <1.0 mm, called Hanford sandy loam, (ii) sieved through 0.5 mm and retained by 0.25 mm, called Hanford medium fraction (MF), and (iii) sieved to <0.25 mm, called Hanford fine fraction (FF).

A Fig. 3—Experimental soil water contents (points), calculated soil water content distributions (solid lines), and bulk density distributions for horizontal infiltration into a layer of Hanford (FF), followed by a layer of Pancoche sl (MF). Calculations are based upon D^* and λ (Table 1). Small broken lines indicate where the wetting front should have been if the column were homogeneous, consisting only of Hanford (FF). Reichardt et al., 1972

B Fig. 5—Calculated pressure head distributions for horizontal infiltration into a layer of Hanford (FF) followed by a layer of Hanford (MF) corresponding to soil water content distributions given in Fig. 3.

Reichardt et al., 1972

D Fig. 7—Experimental values of soil water contents (points), calculated soil water content distributions (large broken lines), and bulk density distributions for an experiment of horizontal infiltration into a sequence of layers consisting of Hanford sl/Panchoche sl/Hanford sl. Calculations are based upon D^* and λ (Table 1). Small broken lines indicate where the wetting front should have been if the column were homogeneous, consisting only of Hanford sl.

Reichardt et al., 1972

ADDITIONAL FACTORS WHICH
INFLUENCE INFILTRATION

A • 3. INFLUENCE OF INITIAL MOISTURE CONTENT θ_0 ON INFILTRATION RATE i , COMPUTED FOR YOLO LIGHT CLAY

t is the time. Figures on the curves denote values of θ_0 . Curves for $\theta_0 = 0.0620$ and $\theta_0 = 0.1661$ omitted for clarity. Curve for $\theta_0 = 0.4400$ based on interpolation, not on detailed computation. Note that curve for $\theta_0 = 0.4950$ is the line $i = K_s$.

J. R. Philip, 1957e

B FIG. 4. INFLUENCE OF INITIAL MOISTURE CONTENT θ_0 ON CUMULATIVE INFILTRATION L , COMPUTED FOR YOLO LIGHT CLAY

t is the time. Figures on the curves denote values of θ_0 . Curves for $\theta_0 = 0.0620$ and $\theta_0 = 0.1661$ omitted for clarity. Curve for $\theta_0 = 0.4400$ based on interpolation, not on detailed computation. J. R. Philip, 1957e

FIG. 6

C • FIG. 6. INFLUENCE OF INITIAL MOISTURE CONTENT θ_0 ON THE "LONG TIME" MOISTURE PROFILES DURING INFILTRATION, COMPUTED FOR YOLO LIGHT CLAY

θ is the moisture content. x'_∞ denotes depth as time increases without limit, with the origin taken at $\theta = 0.4821$. Figures on the curves denote values of θ_0 .

J. R. Philip, 1957e

x'_∞ is the depth to the wet front

B FIG. 3. Influence of water depth h on cumulative infiltration i computed for Yolo light clay; t is time, and the figures on the curves denote value of h in cm. J. R. Philip, 1957f

FIG. 8

- D FIG. 8. INFLUENCE OF INITIAL MOISTURE CONTENT θ_0 ON "LONG TIME" VELOCITY OF ADVANCE OF THE WET FRONT DURING INFILTRATION u COMPUTED FOR YOLO LIGHT CLAY. Solid curve obtained from the detailed analysis; broken curve based on approximate theory (9).

A FIG. 4. Influence of water depth h on infiltration rate, u , computed for Yolo light clay; t is time, and the figures on the curves denote values of h in cm. J. R. Philip, 1957f

C FIG. 4. Influence of water depth h on the depth of the saturated zone x_s computed for Yolo light clay; t is time, and the figures on the curves denote values of h in cm. (note that $x_s = 0$ for all t when $h = 0$) J. R. Philip, 1957f

A

TABLE 1

Depth of Water Penetration*

Date	Water Application mm	Depth of Penetration mm
Aug. 12, 1955	170	4
Aug. 8, 1955	152	3
Aug. 11-13, 1955	470	10
Aug. 18-19, 1955	152	6
Aug. 19-20, 1955	235	4
Aug. 19, 1956	164	2
Aug. 27-28, 1956	152	10
Aug. 6, 1956	152	10

*Source: Remson et al., 1959

Remson et al., 1959

WATER CONTENT, %

B Figure 2—Influence of rain infiltration duration upon water content profiles. S and C indicate data of Rehovot sand (47 + 1 mm. hr. rain) and Beisan clay aggregates (150 + 1 mm./hr. rain), respectively. The initial water contents are indicated by w_i .
Rubin et al., 1964

C Figure 4—Effect of the gravitational term, G , on θ vs x for Geary silt loam ($\theta = 0.184$ for $x > 0$, $t = 0$ and $\theta = 0.460$ for $x = 0$, $t > 0$) and Sarpy loam ($\theta = 0.05$ for $x > 0$, $t = 0$ and $\theta = 0.410$ for $x = 0$, $t > 0$).
Hanks and Bowers, 1962

D Figure 5—Influence of time and of rain intensity upon the wetting front advance rates during infiltration into air-dry Rehovot sand. Numbers next to the curves indicate rain intensity in mm./hr. Rubin et al., 1964

E Figure 6—Influence of time and of rain intensity upon the wetting front advance rates during infiltration into air-dry Beisan clay aggregates. Numbers next to the curves indicate rain intensity in mm./hr. Rubin et al., 1964

Ground-water Recharge and Discharge

A Fig. 3. Effect of soil type.
Freeze, 1963

B Fig. 4. Effect of intensity and duration of rainfall and subsequent conditions.
Freeze, 1963

Numbers on graphs refer to time since the start of rainfall.
R = rain intensity
Q = groundwater recharge rate

A Fig. 5.16 Comparison of moisture profiles for confined and unconfined flow into an unsaturated swelling soil. (Wong, 1971)

Yong and Warkentin, 1975
TRANSIENT INFILTRATION

B Fig. 3. I vs. $\sqrt{t}$ for uniform and crusted profiles. The I data are corrected for the initial quantity of water sorbed by the crusts. Hillel and Gardner, 1970

A FIG. 1. Time-dependence of cumulative infiltration (A) and of infiltration rate (B) for crusted and uncrusted soil columns. R_{cl} , R_{c2} , R_{c3} are 3.2, 9.1 and 17 days, respectively. Hillel and Gardner, 1970

A Fig. 5. Effect of antecedent soil moisture conditions.
Freeze, 1969

B Fig. 6. Effect of groundwater recharge or discharge rate

C Fig. 7. Effect of the depth to the water table
Freeze, 1969

A Figure 2 Basic geologic environment for waste container
(from Frind et al, 1976)

APPENDIX D

MEASUREMENT OF UNSATURATED FLOW PARAMETERS

APPENDIX D

Introduction

In order to model or define unsaturated flow, the value of hydraulic conductivity or diffusivity must be determined for a range of soil-water potentials or water contents. This appendix contains a discussion of the devices and techniques available to measure these parameters in the laboratory or in the field, together with a discussion of their advantages and limitations.

Devices To Measure Soil-Water Potential

The type of device to use for measuring soil-water potential must be selected on the basis of the required range and type of potential. Some of the more common devices are discussed below and include the tensiometer, pressure plate and pressure membrane, thermocouple psychrometer, thermistor psychrometer, salinity sensor, gypsum block, and heat dissipation block.

Tensiometers measure matric potential down to -0.8 bar under laboratory and field conditions by means of water in a porous ceramic tip placed in contact with the soil. Figure D11A illustrates a typical tensiometer. When the soil changes potential, it exerts a tension or pressure on the water in the ceramic tip and causes a change in an attached manometer, Bourdon gage, or pressure transducer (Yong and Warkentin, 1975). When used in the laboratory, tensiometers are usually installed at several locations in an undisturbed core or a column of compacted soil. In the field, the tensiometer is installed in a borehole filled with a soil slurry in order to ensure a good contact between the tensiometer and the in-situ soil (Baker et al., 1974).

Tensiometers typically respond rapidly to changes in potential and are relatively simple to install. They cannot measure potentials below -0.8 bar and are sensitive to changes in temperature (Watson, 1975). When tensiometers are installed in the field, daily temperature fluctuations can cause volume changes in the tubing and the water in the tubing above ground surface even if the tubing is insulated. In one experiment, researchers had to install an air-conditioned hut around the tubing to minimize the temperature effects (Watson et al., 1975).

Peck and Rabbidge (1969) extended the range of the tensiometer by filling it with an osmotic solution, such as polyethylene glycol, which shifted the zero point to a lower potential (D13B). If solutes cannot pass through the semi-permeable membrane in the tensiometer, then the instrument measures the sum of the matric and osmotic potentials. Changes in temperature affect the response of the tensiometer by about 1.5% per 1°C. Leakage of solute and strain-gage creep result in a drift of the zero point of about 1 to 2% per month.

Pressure-plate and pressure-membrane devices have been used in the laboratory to develop moisture content vs potential curves for undisturbed and compacted samples (e.g., Bruce, 1972; Haridasan and Jensen, 1972). Figure D11B illustrates a typical pressure-plate device and Figure D11C illustrates a typical pressure-membrane device. With these devices, air pressure, applied to a sample placed in a chamber, causes water from the sample to move into a ceramic plate or a cellulose membrane. The water content of the sample at equilibrium must then be determined by one of the techniques discussed in the next section. Ceramic plates can facilitate the development of potentials down to about -15 bars (Yong and Warkentin, 1975) and cellulose membranes achieve potentials down to about -110 bars (Richards, 1949). Plates and membranes are available with various maximum pressures. In general, as the maximum pressure increases, the flow rate through the plate or membrane decreases and the time required to reach equilibrium increases. For example, the flow rate for a 3-bar ceramic plate is about 70 times faster than that for a 15-bar plate (Soil Moisture Equipment Corp. Catalog). At high pressures and for clayey soils, equilibrium may take one to two weeks, a distinct disadvantage since vapor losses may occur or air bubbles may collect under the membrane and plates, either of which affect the measurements. The laboratory samples are small (1 to 2 cm for pressure tests), and it is difficult to obtain undisturbed samples that are representative of the in-situ soil (Yong and Warkentin, 1975).

Thermocouple psychrometers measure the total potential (including osmotic potential) in a soil by measuring the vapor pressure in the soil and relating the vapor pressure to potential. The psychrometer consists of a thermocouple enclosed in a ceramic bulk as illustrated in Figure D12B. There are two basic types of psychrometer: the wet loop (D12A) and the Peltier types (D12C). In the wet loop psychrometers, a water droplet is manually placed on the thermocouple whereas in the Peltier type, water is condensed on the measuring junction by

passing a cooling current through the junction. Each psychrometer measures the vapor pressure of the soil by measuring the temperature depression of the junction caused by evaporation of the water. Peltier psychrometers can measure water potentials down to about -80 bars (Wiebe et al., 1971).

Psychrometers are very sensitive to changes in temperature (100 bar/ $^{\circ}\text{C}$ according to Rawlins and Dalton, 1967) which can usually be controlled to within 0.001°C in laboratory experiments. In the field, however, the temperature cannot be controlled and special precautions must be taken to minimize the effect of changes in temperature (Rawlins and Dalton, 1967; Hsieh and Hungate, 1970; Chow and DeVries, 1973). Rawlins and Dalton (1967) minimized the effect by measuring the current in the psychrometer before and after the cooling. The duration of the cooling current must be kept to a minimum if the temperature is changing rapidly. Hsieh and Hungate (1970) developed a temperature-compensated psychrometer which has two series thermocouples connected with opposite polarities. The compensated psychrometer had 1/4 of the fluctuation that the uncompensated psychrometer had when the temperature varied between 20.1 and 20.9°C in 5 minute cycles (Hsieh and Hungate, 1970). Chow and DeVries (1973) developed a three-terminal double-loop psychrometer. They reported that this psychrometer had a thermal stability 40 times better than the uncompensated psychrometer for ambient temperature fluctuations greater than $0.2^{\circ}\text{C}/\text{min}$. Water potentials recorded in their experiments agreed within ± 0.4 bar of the results obtained on the same material tested in a porous-plate extractor.

Kay and Low (1970) developed a thermistor psychrometer containing two thermistors, one located above the sample and one in a reference chamber. The device worked well in calibrated salt solutions and a silt loam but in tests on a clay paste, the thermistor psychrometer resulted in a potential up to 23.5% higher than measured with a tensiometer at the same moisture content. The authors gave no explanation as to which was correct nor why the measurements differed.

A salinity sensor (Figure D13A) measures osmotic potential indirectly. The device contains a thin ceramic wafer which is pressed against the side of a borehole with a spring. After the ceramic wafer reaches diffusional equilibrium with the soil, the conductivity between two electrodes in the wafer indicates the salinity of the pore water (Rawlins, 1971).

Osmotic and total potential have been measured simultaneously with the combined thermocouple psychrometer and salinity sensor of Ingvalson et al. (1970) shown in Figure D13C. The true matric potential can be determined from the difference between the two readings. In calibrating the sensor, the calibrating solution should have the same sodium: calcium ratio as the soil to be measured. It was found that if the matric potential dropped below -2 bars, the ceramic cup desaturated, which changed the calibration of the salinity sensor. In the calibration of the sensor, measurements were repeatable within ± 0.5 bar over a 0 to -10 bar range.

Gypsum blocks have been used to measure indirectly the in-situ potential (Yong and Warkentin, 1975). As the soil moisture changes, so do the moisture content and the electrical resistance of the block. Hysteresis of the blocks and soil is a large source of error; therefore, the blocks should not be used in zones where rapid wetting and drying occurs. Coating the electrodes with nylon improves the sensitivity of the blocks. If the salt concentration in the pore water exceeds 0.2%, the measured resistance is affected (Yong and Warkentin, 1975). The average life span of a gypsum block is 3 to 5 years (Soil Moisture Equipment Co. Catalog).

Phene et al. (1971) developed a device to measure matric potential from the rate of temperature dissipation from a porous block (D14A). The difference in temperature at the center of a block, before and after a heating pulse is applied to the center of a block, is calibrated with matric potential. The blocks can be made of gypsum, which dissolves with time, or of ceramics, which can be made to obtain a certain sensitivity. Phene et al. (1971) measured matric potential to -15 bars, with a precision of ± 0.1 bar.

Methods For Measuring Moisture Content

The moisture content of soil can be measured by either gravimetric or nuclear methods. Gravimetric methods require oven drying the soil and are, therefore, destructive by nature. Nuclear methods are fast and non-destructive but not as accurate as gravimetric methods and can be affected by variations in the chemical composition of the soil.

Gravimetric determination involves weighing a wet sample of soil, drying it in an oven, and reweighing the dried soil. The water content equals the difference between the two weights (weight of water) divided by the dry weight of the soil (weight of solids). Soil science experimenters refer to the water content by volume, θ , which can be derived from the gravimetric

water content by the following relationship (Yong and Warkentin, 1975):

$$\theta = \frac{\gamma_d}{\gamma_w} w$$

where w = gravimetric water content
 γ_d = bulk dry density of the soil
 γ_w = density of water

Several indirect methods have been developed to circumvent the destructive nature of gravimetric tests. In the laboratory, many researchers have used gamma-ray attenuation devices to measure water content (for example, Watson, 1966; Topp, 1971; Gilham et al., 1976). Gamma rays from a source such as Cesium 137 emitted at one side of a soil column, pass through the soil and are measured by a counter on the other side of the column. If the geometry of the system and the absorptive properties of the soil and container are known, the water content can be measured as a function of the intensity of the detected gamma rays. The gamma-ray-attenuation technique can be used to measure water contents in the field by excavating an annular space and forming a cylindrical pedestal of soil, which then can be tested as shown in Figure D15A as if it were a laboratory soil column (Watson et al., 1975).

A similar method, derived specifically for field use, is the neutron probe (Roulier et al., 1972; Royer and Vachaud, 1975). The neutron probe contains a source of fast neutrons such as radium or beryllium, and a counter to monitor the rate at which neutrons return to the probe. The rate of return of the neutrons depends on the number of hydrogen atoms present in the soil which is proportional to the water content of the soil. The neutron probe requires calibration for each soil type encountered and is most useful in monitoring changes in the water content of a particular soil, rather than in determining the spacial distribution of in-situ water content at a particular site. Further, the presence of boron, potassium or iron in the soil can affect the accuracy of the probe (Yong and Warkentin, 1975). Consequently, the neutron probe should be used in conjunction with a more accurate technique, such as gravimetric determinations, when determining the initial moisture content versus depth distribution in a soil profile.

Traditionally, only one of the aforementioned parameters, either potential or moisture content, has been measured in the field. Thus, laboratory tests on remolded or undisturbed samples have been required to obtain the θ vs ψ relationship of the soil.

Royer and Vachaud (1973) have shown that hysteresis of the θ vs Ψ relationship can be significant and that in field situations when the soil profile is subjected to both wetting and drying such hysteresis must be accounted for. Furthermore, laboratory samples inherently are disturbed, which means that the soil structure has been changed significantly. Therefore, the Ψ vs θ curve obtained in the lab may not be applicable to the field situation and both water content and potential should be measured simultaneously in the field.

Methods of Measurement of Hydraulic Conductivity and Diffusivity

Laboratory Methods

Table 4a of Appendix A contains references to many methods available for the determination of hydraulic conductivity, K, or diffusivity, D, of undisturbed or reconstituted samples in the laboratory. A few of the most common methods used today are described below. Most laboratory tests to determine K or D involve either steady state experiments, transient experiments, or indirect methods (which utilize other soil properties such as grain-size distribution to estimate K or D).

Steady state tests involve applying a potential gradient to a soil column and measuring the resultant flow of water. In one such test the soil sample is placed in a long column saturated at one end and with a controlled water loss at the other (Figure D16AB). Based on the geometry of the column and the rate of water loss, an average conductivity can be calculated. Another steady-state method involves clamping a sample between two porous plates which are maintained at different potentials and thus inducing a flow of moisture in the sample. By measuring the rate of moisture flow through the soil and the potential in the soil, the hydraulic conductivity can be calculated.

Diffusivity can also be determined from steady-state tests if water contents are measured or by using the Ψ vs θ curve for the soil. Table 4a lists several steady-state experiments in which water contents were determined either gravimetrically or by gamma ray attenuation techniques and D was calculated. In these tests potentials were generally measured with tensiometers and so are restricted to potentials above -0.8 bar.

Transient flow experiments can be performed in one of several manners, such as illustrated in Figure D17A. For example, one can subject a sample in a pressure cell or pressure membrane device to a change in pressure and measure the induced outflow of water (Yong and Warkentin, 1975). Transient flow experiments also include tests of soil columns in which potentials and water contents are measured as the soil takes on water (sorption) or as the soil drains (desorption). Waston (1966) developed an "instantaneous profile method" to determine $K(\theta)$ from instantaneous potential and moisture content measurements along columns of soil. Weeks and Richards (1967) developed empirical equations for determining K and D from measurements of potentials and flow rates during transient flow. They reported that the equations gave reasonable values of average hydraulic conductivity for a given column subjected to potentials down to -1 bar. As with the steady state experiments, the potentials considered in the transient flow experiments usually depended on measurements with tensiometers and, therefore, were restricted to potentials above -0.8 bar. Mehuys et al. (1975) measured potentials down to about -60 bars with thermocouple psychrometers and computed $K(\theta)$ and $K(\psi)$ using the method of Weeks and Richards (1967). They found that the method of Weeks and Richards may be valid for a much wider range of potentials than originally reported and suggested this range may include potentials as low as -50 bars.

Alemi et al. (1976) determined the hydraulic conductivity of soil cores by using a centrifuge but found that compaction of the sample during the test may affect the results.

Diffusivity can be calculated from the water-content profile after infiltration of water into a horizontal column of soil by using the flow equations presented in Section 3 (Yong and Warkentin, 1975; Bruce and Klute, 1956; and Jackson, 1964). Hillel and Gardner (1970) measured diffusivity by placing a less permeable crust on top of a column of soil and measuring the infiltration with time. Different potentials in the sample were obtained by varying the hydraulic resistance of the crust.

Indirect methods of computing $K(\theta)$ from θ - ψ curves and pore-size measurements have been developed by Millington and Quirk (1960), Marshall (1958), and Childs and Collis-George (1950). These methods have been applied to several soils with varying success. (In general, the best agreement was found for coarse-grained soils.) In all cases, a "matching" factor was required in order to make the calculated curves agree with experimental results. Typically, K_0 (the saturated hydraulic conductivity) was used to match the curves.

Field Methods

Table 4b of Appendix A contains a summary of various methods for the field measurement of hydraulic conductivity or diffusivity, some of which are discussed below. Most field tests require application of water to the soil surface and monitoring of the resultant change in water content with depth as redistribution takes place. Thus, these methods are essentially "instantaneous profile" methods similar to the lab tests of Watson (1966).

Rose et al. (1964) irrigated a soil surface, measured in-situ water contents with neutron moderation equipment, and then determined the corresponding potentials from laboratory θ - Ψ curves. Roulier et al. (1972), on the other hand, measured moisture potential with tensiometers and moisture content with a neutron probe in the field. They also performed similar tests on laboratory samples 30-cm and 6-cm long of undisturbed cores of the same material. Good agreement was obtained between in-situ tests and the lab tests on 30-cm long cores but not the 6-cm long cores. According to Baker et al. (1974), the instantaneous profile method may incur problems in the field if the soil strata are not horizontal or if the hydraulic conductivity of the layers in the profile varies.

Baker (1977) performed tests on a cylindrical pedestal of in-situ soil which was formed by excavating an annular hole as illustrated in Figure D15A. A crust of gypsum and sand was placed on top of the pedestal to maintain a constant suction in the soil. Baker (1977) then measured the water entering the soil from an infiltrometer and computed K. By varying the composition of the surface crust different potentials were established and the corresponding flows were measured to establish K vs Ψ .

Laboratory Measurement of Thermal Properties of Soil

Several researchers have performed experiments to measure moisture and heat flow under thermal gradients. Cary (1965) measured liquid and vapor flow in a soil sample placed between two plates maintained at different temperatures. Cassel et al. (1969) measured soil water movement in a column of soil under a longitudinal thermal gradient. They found that radial thermal losses occurred in the experiments and that a two-dimensional model should be used to analyze the heat and water transfer in a cylindrical column. Moench and Evans (1970) measured the thermal conductivity and diffusivity in two soils by embedding a cylindrical thermal probe in a sample of each material. They

reported an accuracy of 4% in the measurement of thermal conductivity, based on conductivity determinations on calibrated solutions.

Jury (1973) performed experiments to measure transport coefficients for liquid and vapor flow under thermal gradients in columns of soil. He measured thermal conductivity and moisture diffusivity on samples with a range of water contents corresponding to 20% to 90% pore saturation. The measured thermal conductivity agreed within 7% with values calculated using the method of DeVries (1963).

Application To the Modeling of a Vadose Zone Repository

Solution of the general flow equations requires determination of the following parameters:

1. for Ψ -based equations - K and Ψ
2. for Θ -based equations - D and Θ

Applying flow equations to the mathematical simulation of the flow of moisture in the vadose zone requires knowledge of the above parameters as they exist in-situ. Measurement of these parameters requires field exploratory borings and either in-situ or laboratory tests on the soil in question. However, presently available exploration techniques and testing methods place severe restrictions on the application of current unsaturated-flow theory to the prediction of the viability of vadose zone repositories.

Most research has concentrated on agricultural soils at water contents near or above field capacity and indicates that the simplicity and accuracy of measurement techniques decrease as water content decreases. Field measurements have been restricted to the upper few meters of the soil profile and most laboratory measurements have been restricted to reconstituted soil samples. Tensiometers cannot measure potentials below -0.8 bar and psychrometers exhibit extreme sensitivity to temperature fluctuations. Reconstituted soil samples do not accurately represent in-situ soil structure and undisturbed samples are very difficult to obtain. A typical arid site vadose zone repository (VZR) would be constructed in non-agricultural soils with very low moisture contents deep below the ground surface, soil water potentials may be hundreds of bars below zero, and surface temperatures may fluctuate several tens of degrees. Consequently, one cannot expect to accurately predict the performance of a VZR based on parameters measured with currently available measuring devices.

The most reasonable approach to modeling the flow of moisture in the vadose zone would probably include a θ -based analysis with field measurement of water content vs depth and laboratory determination of D vs θ for "undisturbed" soil samples. This approach would minimize extrapolations since D can be measured at much lower water contents than ψ and would minimize error in obtaining field data since water content determinations are less complicated and less temperature-sensitive than measurements of soil water potential. However, such an approach would be limited in that "undisturbed" samples are very difficult to obtain from depths common to waste depositories (i.e., 150 to 200 m) and laboratory tests on small samples may not represent the overall diffusivity of the field soil (i.e., scale effects).

Insufficient data preclude any conclusions as to the reliability of the measurement of thermal properties of soil, i.e., $\partial T / \partial z$ and D_T (see Section 3.3.2). Some laboratory work has been done, but not on soils at very low water contents, and few, if any, field measurements of T and D_T at depths greater than a few meters have been reported in the literature.

A Fig.4.18. Soil-water tensiometer for field use. (Soilmoisture Equipment Co.)

Yong and Warkentin (1975)

A

B

C Fig.4.16. Apparatus for measuring soil-water potential vs. water content. B. Unit for use to - 1 bar. C. Unit for use to - 25 bar.

Yong and Warkentin (1975)

Fig. 1. Wet-loop psychrometer (from Richards and Ogata, 1958) showing insulating cover (1), liquid bath (2), masonite cover (3), test tube (4), brass cylinder (5), copper disk (8), copper tube (7), plastic handle (6), lead wires (11), and stainless steel sample container with end cap (18). A droplet of water is held in the silver ring supported by chromel and constantan wires in the cavity formed in the soil sample (17).

Rawlins (1971)

Fig. 2. Cross section of the thermocouple psychrometer used by Rawlins and Dalton (1967) for *in situ* measurement of soil water potential showing Teflon insert (1), copper heat sink (2), chromel-constantan thermocouple (3), copper lead wires (4), acrylic tubing (5), epoxy resin encapsulating heat sinks (6), and ceramic bulb (7). Water is condensed on the thermocouple junction by Peltier, cooling immediately preceding each measurement. Rawlins (1971)

Fig. 3. Temperature-compensated Peltier psychrometer assemblies. (A) This complete assembly can be used either as a conventional psychrometer or as a thermally compensated one, depending on the connections of the four external lead wires. (B) A simplified temperature-compensated psychrometer with only two external lead wires. (C) Top view diagram of (B) psychrometer. (After Hatch and Hungate, (11)). Rawlins (1971)

A. Fig. 1- Front and sectional views showing: (1) ceramic plate, (2) front electrode, (3) back electrode, (4) nylon tubing, (5) epoxy block, (6) thermistor, (7) outer Lucite case, (8) spring for holding the sensitive element against soil, (9) release pin, and (10) pull wire.

Salinity Sensor, Richards (1966)

B. Fig. 8. Cross-sectional view of the osmotic transducer proposed by Peck and Rabbidge (1969) showing (1) semipermeable membrane, (2) O-ring seal, (3) brass body, (4) 15-bar bubbling pressure ceramic, (5) clamping ring, (6) strain gauge pressure transducer, (7) O-ring seal, (8) vent, (9) O-ring seal, (10) filling plug [After Peck and Rabbidge, (21).] Rawlins (1971)

C. Fig. 1- Cross section of the combined thermocouple psychrometer and salinity sensor used for measuring water and osmotic potential of soil water.

Ingvalson et al. (1970)

A FIG. 7. Cross-sectional sketch of the matrix potential sensor proposed by Phene et al. (22).
Rawlins (1971)

A Fig. 1—Schematic diagram of the crust test apparatus. TB = manometer board, S = scale, B = burette stand, M = manometer device, Cy = metal cylinder with sharpened lower edge, C = crust, PC = plastic cover, Sc = wing nut, W = water inlet, A = air bleeder, RG = rubber gasket, Ca = cap, P = porous cup, θ = zero mercury level, H = height of mercury rise above zero level, G = distance between porous cup level and zero mercury level.

Baker (1977)

A Fig. 1—Schematic diagram of apparatus.

Elrick and Bowman (1964)

B Fig. 1—Diagram of pressure cell apparatus.

Haridasan and Jensen (1972)

A Fig. 1. A schematic drawing of the experimental apparatus. (a) 25-cm column showing the constant head burettes, air lines, tensiometers, and one pressure transducer. (b) 10-cm column without the attachments. The numbers on the right of each column give the positions of the tensiometers T_1 , T_2 , T_3 , T_4 , T_5 and the air ports AP. The pressure transducer is indicated by PT and the soil by S. The gamma ray beam traversed the soil in a direction perpendicular to the diagram.

Topp (1971)

APPENDIX E

EFFECTS OF THERMAL GRADIENTS

APPENDIX E

Since the early experiments of Bouyoucos (1915), many soil physicists have investigated the flow of water due to thermal gradients in soils. Gurr et al. (1952) performed several tests in which salt was used as a tracer to monitor the flow of moisture due to thermal gradients in a closed system (i.e., no drainage from soil). Their results indicated (1) a net flow of moisture in the direction of decreasing temperature, and (2) an increase in the salt concentration at the warmer end of the soil specimen. These results imply there was a flow of moisture as vapor down the thermal gradient, which may have been accompanied by a lesser liquid flow in the direction of increasing temperature, due to induced moisture gradients. The data of Gurr et al. (1952), reproduced on page E-4, illustrate this phenomenon well.

Hutcheon (1958), E-5, determined the effect of initial volumetric water content, θ , on the resultant temperature-induced moisture gradients for three soils. Decreasing the initial water content increases the resultant temperature gradient.

Most equations which have been proposed to model unsaturated flow due to thermal gradients are of the diffusion type. (See Section 3.3.2 of text.) Although these equations have not been verified extensively, some experiments have been performed to show the application of the various theories to real soil (mostly relatively moist soil). In this appendix, the relationship of diffusivity to water content is examined and some experimental measurements of moisture are discussed.

Diffusivity

It was shown previously (Section 3.3.1 of the text) that diffusion theory can be used to mathematically describe isothermal unsaturated flow. Jackson (1963), E6A, obtained experimental data which indicate that diffusivity increases with increasing ambient temperature. However, Jackson (1964), E6BCD, conducted experiments to test the validity of diffusion theory for soils with low water contents subjected to thermal gradients and found an anomaly in the D vs θ relationship at low water contents; there was a water content in the dry range

at which diffusion reached a secondary maximum.

Jackson (1964) explored this phenomenon further and found that there were no simple relationships among D , θ , and soil texture or porosity. Rather, he indicated there must be other influencing factors such as structure, type of clay mineral, and the amount and type of ions on the clay complex. For the most part, the diffusivities he measured were in general agreement with predictions based on the method of Philip (1955).

Jury (1973), E7, reported data from experiments in which he measured the diffusion coefficient of a medium sand. He found a relation between D and θ that was similar to that of Jackson, 1964, with the exception that he found the anomaly in the D vs θ curve at higher θ .

Moisture Flux

As mentioned previously, there are very few data from experiments in which moisture fluxes due to thermal gradients were measured in soils at very low water contents. Cary and Taylor (E8, E9), Cary (E10B), Rollins *et al.* (E11ABC) and Cary (E11D) have provided some useful data, although all of their experiments were on soils at volumetric water contents greater than 10%. Nevertheless, their data illustrate several characteristics of thermally induced moisture flow:

1. As the temperature gradient increases, the moisture flux increases (E8, E9, E10B, E11BC).
2. As the average temperature increases, the influence of temperature gradient increases causing greater fluxes due to an increased vapor flow component (E8, E9, E10B).
3. Cary's data (E9A to F) indicate a temperature gradient of 0.2°C/cm (with a cool side temperature of 15°C) will cause a moisture flux from 1 to 3 cm/yr depending on the potential gradient or water content during the experiment.

Application To the Modeling of a Vadose Zone Repository

At the time of burial, the high temperature of HLW and spent fuel will tend to oppose any tendency for moisture flow toward the depository. Hence, if no moisture could contact the waste, there would be no mechanism for transport of radio-nuclides away from the depository.

The prediction of thermally induced moisture flow is still in a nascent state. There are several tables reproduced herein which summarize the various prediction techniques. Philip and DeVries (E12) compared several predictions with measured values and found that diffusion theory tends to underestimate flow by as much as 18 times. They subsequently proposed a different theory which is discussed in the text (Section 3.3.2). Cassel et al. (E13) compared the prediction methods of Philip and DeVries, Cary and Taylor, and pure diffusion theory (Ficks Law). Their results, summarized in table form, indicate the methods of Philip and DeVries agreed well with experiments on Columbia sandy loam.

A TABLE 2

Some properties of Uribae loam and washed Holder fine sand

		LOAM SOIL	WASHED FINE SAND
Particle size distribution	2-0.2 mm	% 1.9	20.0
	0.2-0.02 mm	% 50.2	80.0
	0.02-0.002 mm	% 31.3	0.0
	<0.002 mm	% 16.6	0.0
Total soluble salts		% 0.077	—
Chloride content per 100 gm. soil	me.	0.14	—
Moisture equivalent	%	20.0	2.5
Permanent wilting percentage		6.3	—
Plastic limit	%	15.8	—

Gurr et al., 1952

FIG. 2. DISTRIBUTION OF WATER AND CHLORIDES IN COLUMNS OF LOAM SOIL, OF VARYING INITIAL WATER CONTENT, SUBJECTED TO A TEMPERATURE GRADIENT FOR 5 DAYS
Gurr et al., 1952

B FIG. 3. DISTRIBUTION OF WATER AND CHLORIDES IN A COLUMN OF LOAM SOIL, OF 11.3 PER CENT INITIAL WATER CONTENT, SUBJECTED TO A TEMPERATURE GRADIENT FOR 14 DAYS. Gurr et al., 1952

D FIG. 4. DISTRIBUTION OF WATER AND CHLORIDES IN TWO COLUMNS OF WASHED FINE SAND, TO WHICH NaCl WAS INITIALLY ADDED AT THE COLD END AFTER BEING SUBJECTED TO A TEMPERATURE GRADIENT FOR 10 DAYS.

Gurr et al., 1952

A Figure 3. Comparison of moisture distribution at different initial moisture contents within soil columns subjected to a temperature gradient from 10 to 25 C.
Hutcheon, 1958

A Figure 2—Soil-water diffusivity vs. volumetric water content for three soil materials at three temperatures. Jackson, 1963

B Fig. 3—Diffusion coefficients for Adelsola I. at three sets of boundary and initial conditions. Each symbol indicates the average of two measurements. Jackson, 1964

C Fig. 4—Diffusion coefficients for Pine soil as a function of gravimetric water content. Jackson, 1964

D Fig. 5—Diffusion coefficients for Pachappa I. as a function of gravimetric water content. Jackson, 1964

A Figure 5.11 Moisture Diffusivity vs Water Content
 July, 1973

Figure 2—Water flux through Millville loam soil with an approximate moisture content of 15.9%. Other approximate characteristics of this sample were a bulk density of 1.27, a suction of 51 centibars, and an air-filled pore space of 31%. Experimentally measured flow rates are shown as points and the smooth curves with cool side temperature as a parameter are predicted by the equation $J_v = 5.4435 \times 10^3 \Delta \log T + 6.6627 \times 10^3 \Delta 1/T$.

Cary and Taylor, 1962a

Figure 4—Water flux through Millville loam soil with an approximate moisture content of 16.7%. Other approximate characteristics of this sample were a bulk density of 1.12, a suction of 37 centibars, and an air-filled pore space of 39%. Experimentally measured flow rates are shown as points and the smooth curves with cool side temperature as parameters are predicted by the equation $J_v = 4.2130 \times 10^3 \Delta \log T + 5.0564 \times 10^3 \Delta 1/T$.

Cary and Taylor, 1962a

Table 1—Summary of phenomenological coefficients when ΔT was measured across a distance of 7.5 cm.

Sample	Wt. % water	Sat. th. centibars	Bulk density	% of air- filled pore	L_{12} $\times 10^3$	L_{21} $\times 10^3$	L_{22} $\times 10^3$	L_{23} $\times 10^3$	L_{32} $\times 10^3$
Air	20	-	1.01	100	6.701	1.265	1.925	1.901	6.5
Millville loam soil	15.9	54	1.27	31	8.512	1.165	2.712	2.409	5.9
Millville loam soil	16.7	37	1.12	39	4.750	0.801	1.947	1.868	1.9
Millville loam soil	16.3	47	1.10	40	3.329	1.101	2.351	2.360	6.7

Cary and Taylor, 1962a

Figure 3—Heat flux through the same sample considered in figure 2. Experimentally measured flow rates are shown as points and the curves with cool side temperature as parameters are predicted by adding J_v to the equation $J_q = 4.6910 \Delta \log T + 580.00 \Delta 1/T$.

Cary and Taylor, 1962a

Figure 5—Heat flux through the same sample considered in figure 4. Experimentally measured flow rates are shown as points and the three smooth curves with cool side temperature as parameters are those predicted by adding J_v to the equation $J_q = 3.3287 \Delta \log T + 393.06 \Delta 1/T$.

Cary and Taylor, 1962a

Figure 1—Water flux through Millville loam soil. Approximate physical characteristics were a bulk density of 1.39 and a suction of 19 centibars. Experimentally measured flow rates are shown as points, and the smooth curves with cool side temperature as a parameter are predicted by the equation

$$J_w = \left(\frac{T^2 (1.568)}{1.524} + 1.174 \cdot 10^5 \Delta \log T \right) \cdot \left(\frac{T^2 (1.5127)}{1.524 \cdot 10^4} + 1.175 \cdot 10^5 \Delta \frac{1}{T} \right)$$

Figure 2—Heat flux through the same system shown in figure 1. Experimentally measured flow rates are shown as points, and the smooth curves with cool side temperature as a parameter are predicted by the equation

$$J_h = (0.96027 + 3.3632) \Delta \log T + (62.594 + 402.10) \Delta \frac{1}{T}$$

Cary and Taylor, 1962b

Figure 3—Water flux through Millville loam soil. Approximate physical characteristics were a bulk density of 1.39 and a suction of 11 centibars. Experimentally measured flow rates are shown as points and the smooth curves with cool side temperature as parameters are predicted by the equation

$$J_w = \left(\frac{T^2 (2.9112)}{1.524} + 1.246 \cdot 10^5 \Delta \log T + \left(\frac{T^2 (1.561)}{1.524 \cdot 10^4} + 1.175 \cdot 10^5 \Delta \frac{1}{T} \right) \right) \cdot \left(\frac{T^2 (1.5127)}{1.524 \cdot 10^4} + 1.175 \cdot 10^5 \Delta \frac{1}{T} \right)$$

Figure 4—Heat flux through the same system shown in figure 3. Experimentally measured flow rates are shown as points and the smooth curves with cool side temperature as a parameter are predicted by

$$J_h = (1.0262 + 2.4552) \Delta \log T + (66.893 + 305.22) \Delta \frac{1}{T}$$

Cary and Taylor, 1962b

Figure 5—Water flux through Millville loam soil with a moisture content of about 27%. Other approximate characteristics were a bulk density of 1.39, a moisture suction of 3 centibars and an air-filled pore space of 10%. Experimentally measured flow rates are shown as points and the smooth curves with cool side temperature as parameters are predicted by the equation

$$J_w = \left(\frac{T^2 (1.8459 \cdot 10^4)}{1.524 \cdot 10^4} \Delta \log T + \frac{T^2 (1.3217 \cdot 10^4)}{1.524 \cdot 10^4} \Delta \frac{1}{T} \right) \cdot \left(\frac{T^2 (1.5127)}{1.524 \cdot 10^4} + 1.175 \cdot 10^5 \Delta \frac{1}{T} \right)$$

Cary and Taylor, 1962b

Figure 6—Heat flux through the same sample considered in figure 5. Experimentally measured flow rates are shown as points and the smooth curve is predicted by the equation

$$J_h = 1.2239 \Delta \log T + 84.260 \Delta \frac{1}{T}$$

The experimental points are labeled with the parameters of cool side temperature.

Cary and Taylor, 1962b

A TABLE 2

Predicted and observed values of moisture flow at various soil moisture suctions where J_v is the vapor transfer, J_t is transfer due to the pressure gradient, and J_{lt} is liquid-phase transfer due only to the thermal gradient

Suction	K	T	VT	Δ	J_v	J_t	J_{lt}	J_m	
								Net	Observed
cm. Hg	$\frac{d\theta}{d\psi}$ °C/cm. Hg.	°C.	°C./cm.	cm. Hg./cm.	mm. day				
5	1.30×10^{-3}	7.0	0.71	-0.2	0.13	-0.09	0.85	0.89	0.86
5	1.63	18.7	0.63	-1.8	0.22	-1.08	0.93	0.07	0
5	1.29	32.4	0.42	-2.1	0.32	-1.76	0.78	-0.66	-0.81
10	1.19	6.3	0.64	2.4	0.15	0.31	0.15	0.91	0.92
10	1.20	17.8	0.54	-2.6	0.25	-0.47	0.18	0.26	0.14
10	1.43	31.6	0.82	0.9	0.92	0.23	0.92	2.07	2.08
18	0.93	8.6	1.12	-6.1	0.31	-0.15	0.60	0.47	0.63
18	1.14	20.0	0.98	0.3	0.49	0.03	0.66	1.17	1.11
18	1.04	33.2	0.56	-4.5	0.62	-0.65	0.18	0	0.32

J. W. Cary, 1965

B FIG. 1. Net heat flux and net thermal moisture flux shown as a function of temperature gradient. Data are for suctions of 5, 10, and 18 cm. Hg, at average temperatures of 8, 18, and 33°C.

J. W. Cary, 1965

A TABLE 1
SOIL CHARACTERISTICS

Soil Series	Classification	Attribute Limits	Centrifuge Moisture Equivalent	Capillary Rates
Rollins loam (551)	Lo	A 400 27.3 20.6 2.7	11.34	11.00
Mechanical Analysis				
Particle Size	$\%$	Carbon Exchange Capacity	pH	Oxygen Matter (%)
4 to 6 μ	0.3	11.2 M.E./100g	8.35	0.21
6 to 10 μ	0.2			
10 to 20 μ	3.1			
20 to 50 μ	82.6			
50 μ to	13.4			
100 μ	11.4			

Rollins et al., 1954

C Figure 11. Relationship between flow rate and thermal gradient.

Rollins et al., 1954

B TABLE 2
RELATIONSHIP BETWEEN FLOW RATE AND THERMAL GRADIENT

Total No.	Dry Density (gms/cm³)	ρ_w H ₂ O	% Air Voids	Temp. Diff.	Rate (gms/hr cm²) of flow
1	1.32	21.8	21.5	40° - 2°C	4.40×10^{-3}
				30° - 2°C	2.86×10^{-3}
				20° - 1°C	1.42×10^{-3}
2	1.33	22.1	20.5	40° - 2°C	4.46×10^{-3}
				30° - 2°C	2.66×10^{-3}
				20° - 1°C	1.43×10^{-3}
3	1.32	22.4	20.2	40° - 0°C	0.81×10^{-3}
				30° - 0°C	1.36×10^{-3}
				20° - 1°C	2.51×10^{-3}
4	1.34	22.5	19.6	40° - 1°C	1.32×10^{-3}
				30° - 0°C	0.47×10^{-3}
				20° - 1°C	4.46×10^{-3}
5	1.26	17.07	30.7	40° - 2°C	5.29×10^{-3}
				30° - 2°C	3.33×10^{-3}
				20° - 1°C	1.74×10^{-3}
6	1.26	17.3	30.4	40° - 0°C	0.71×10^{-3}
				30° - 0°C	0.71×10^{-3}
				20° - 1°C	3.32×10^{-3}
7	1.22	16.55	33.9	40° - 2°C	5.30×10^{-3}
				30° - 2°C	3.32×10^{-3}
				20° - 1°C	1.70×10^{-3}
8	1.23	16.55	33.0	40° - 0°C	0.71×10^{-3}
				30° - 0°C	0.71×10^{-3}
				20° - 1°C	3.47×10^{-3}
9	1.23	16.55	33.0	40° - 2°C	5.54×10^{-3}
				30° - 2°C	3.47×10^{-3}
				20° - 1°C	1.64×10^{-3}

Rollins et al., 1954

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

Type of porous material	Moisture content on suction in boxes	Thermal flow into H ₂ O dish (no °C/cm)	cm H ₂ O head equiv. to 1°C gradient	Mean temperature	Type of solution flowing	Phase of flow	Source of data
Sintered glass, 100A mean pore diameter	saturated	0	0	25	distilled water	liquid	1
Dry Pore 600 cellophane	saturated	0.1*		10	deph. distilled water	liquid	12
Cellophane acetate 27A mean pore diameter	saturated		7.0	20	distilled water	liquid	7
Melvinco filter 100A mean pore diameter	saturated		0.10	20	distilled water	liquid	7
Wetmont (epoxide paste 10 ³)	saturated	0.1†	1.0†	25	0.1N NaCl	liquid	2
Yolo loam and $\rho_w = 1.1$	0.09	2.0	5.5	15	0.01N CaSO ₄	liquid + vapor	1
Houston Black Clay	0.12	0.7		36	distilled water	liquid + vapor	1
Houston Black Clay	1.1	0.8		17	distilled water	liquid + vapor	1
Middleville loam $\rho_w = 1.2$	saturated		0.07	25	water	liquid	20
Middleville silt loam	2.2		1.1×10^4	25	water	vapor + liquid	19
Middleville silt loam	126		8.3×10^4	25	water	vapor	19
Columbia loam $\rho_w = 1.2$	0.07	1.8	3	19	distilled water	20% vap. 80% liq.	2
Columbia loam $\rho_w = 1.2$	0.21	0.9	13	8	distilled water	33% vap. 67% liq.	2
Columbia loam $\rho_w = 1.2$	0.21	2.0	11	31	distilled water	55% vap. 45% liq.	2
Columbia loam $\rho_w = 1.2$	0.13	1.6	250	25	distilled water	vapor + liquid	2
Stable air	0.4	0.4	8×10^4	27	distilled water	vapor	1

* Flow sectional area estimated as 1 cm² and membrane thickness at 0.1 cm.

† Results changed with aging.

‡ Previously unpublished data obtained by the author in experiments similar to those reported in (2).

J. W. Cary, 1966

A TABLE 1 - Experimental data on water-vapor transfer in soils under temperature gradients

Authors	Flow movement identified as in vapor phase	Porosity $\theta + a$	Volumetric moisture content θ	Volumetric air content a	Ratio of observed transfer to value predicted by simple theory $\frac{q}{q_0}$	Remarks ^a
Landwehr [1932]	Liquid flow assumed proportional to Cl movement (NaCl used as tracer)	0.472 ^b	0.074	0.398 ^b	3.6	Uncertainty in this value about 10%
Tyler and Loeferle [1954]	Gaps in column presumed to restrict movement to vapor phase	0.555 ^b	0.206 ^b	0.349	18.	Recomputed by us (from Fig. 5 after one day).
		0.657 ^b	0.138 ^b	0.499	9.8	Computed by us (from Fig. 4 after 2 days).
Tyler and others [1954]	Liquid return flow (cold to hot) via external capillary presumed to equal vapor flow (hot to cold) in column	0.690	0.179	0.511	4.6	Computed by us (from Fig. 6).
		0.541	0.202	0.339	4.6	" " (trial no. 7).
		0.521	0.260	0.261	5.7	" " (from Fig. 7).
		0.503	0.288	0.215	6.0	Computed by Rolling and others (trial no. 1).
		0.457	0.308	0.149	6.8	Computed by us (from Fig. 8).
		0.407	0.329	0.078	9.8	" " (from Fig. 9).

^a These values computed on assumption that S.G. of soil material = 2.65.

^b Figure numbers in this column refer to figures appearing in the corresponding paper cited in column 1.

Philip and DeVries, 1957

A
Tabl —Comparison of observed net water fluxes through 1-cm sections to those predicted by Fick's law, the Philip-de Vries Theory, and the Taylor-Cary Theory for the 0 to 76-hour interval of run B1

Column Interval	$Q_{obs}(t) \cdot 10^3$	$Q_{Fick} \cdot 10^3$	$\frac{Q_{Fick}}{Q_{obs}(t)}$	$Q_{p.d.v.} \cdot 10^3$	$\frac{Q_{p.d.v.}}{Q_{obs}(t)}$	$Q_{T.C.} \cdot 10^3$	$\frac{Q_{T.C.}}{Q_{obs}(t)}$
cm	cm/day	cm/day		cm/day		cm/day	
1	1.80	2.66	1.48	10.9	6.06	0.34	0.311
2	2.33	1.39	0.61	6.83	2.98	0.14	0.052
3	4.64	0.874	0.188	1.88	1.05	0.14	0.030
4	5.75	1.52	0.264	4.46	0.445	0.30	0.052
5	6.67	1.11	0.166	5.45	0.477	0.12	0.018
6	7.44	1.22	0.161	5.86	0.444	0.45	0.060
7	8.76	1.51	0.187	6.72	0.405	0.27	0.031
8	9.44	1.66	0.197	7.80	0.524	0.55	0.058
9	9.72	1.80	0.206	8.47	0.573	0.57	0.065
10	9.88	1.71	0.196	8.47	0.532	0.57	0.061
11	9.89	1.59	0.167	7.56	0.441	0.57	0.061
12	9.92	1.63	0.181	6.44	0.333	0.42	0.047
13	9.92	1.58	0.175	10.4	1.11	0.70	0.084
14	9.92	1.49	0.120	9.34	1.03	0.47	0.047
15	9.96	1.52	0.170	9.35	1.04	0.68	0.076
16	9.89	1.79	0.201	10.1	1.14	0.55	0.060
17	9.81	1.98	0.230	11.8	1.18	0.85	0.102
18	9.09	2.21	0.266	12.8	1.70	0.42	0.047
19	9.46	2.18	0.305	16.0	2.17	0.14	0.012

B
Table 3—Comparison of observed net water fluxes through 1-cm sections to those predicted by Fick's law, the Philip-de Vries Theory, and the Taylor-Cary Theory for the 76 to 186-hour interval of run B1

Column Interval	$Q_{obs}(t) \cdot 10^3$	$Q_{Fick} \cdot 10^3$	$\frac{Q_{Fick}}{Q_{obs}(t)}$	$Q_{p.d.v.} \cdot 10^3$	$\frac{Q_{p.d.v.}}{Q_{obs}(t)}$	$Q_{T.C.} \cdot 10^3$	$\frac{Q_{T.C.}}{Q_{obs}(t)}$
cm	cm/day	cm/day		cm/day		cm/day	
1	1.87	2.78	1.61	11.8	6.59	0.43	0.26
2	3.01	1.43	0.472	6.92	2.24	0.046	0.029
3	4.24	0.856	0.205	4.71	1.11	0.095	0.022
4	5.36	1.52	0.291	4.73	0.882	0.19	0.035
5	6.51	1.31	0.174	5.77	0.314	0.16	0.022
6	7.14	1.22	0.171	5.73	0.402	0.41	0.077
7	7.95	1.51	0.193	6.76	0.454	0.23	0.029
8	8.64	1.68	0.192	7.71	0.492	0.59	0.068
9	8.19	2.08	0.267	8.75	0.532	0.40	0.044
10	9.61	1.79	0.186	8.31	0.468	0.47	0.049
11	9.75	1.50	0.144	7.51	0.70	0.53	0.054
12	9.68	1.63	0.169	8.27	0.456	0.45	0.047
13	8.36	1.58	0.169	10.2	1.19	0.66	0.071
14	6.72	1.40	0.206	5.23	1.06	0.88	0.078
15	7.74	1.52	0.195	9.00	1.16	0.28	0.030
16	6.42	1.79	0.279	9.96	1.53	0.32	0.059
17	4.81	1.81	0.197	11.3	2.33	0.66	0.14
18	3.03	2.17	0.716	13.3	4.39	0.35	0.12
19	1.31	2.14	1.65	12.5	9.54	0.18	0.14

Cassel et al., 1969