# Sediment Properties and Water Movement Through Shallow Unsaturated Alluvium at an Arid Site for Disposal of Low-Level Radioactive Waste Near Beatty, Nye County, Nevada

By Jeffrey M. Fischer

---

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 92-4032

Carson City, Nevada
1992

U.S. DEPARTMENT OF THE INTERIOR

MANUEL LUJAN, JR., *Secretary*

U.S. GEOLOGICAL SURVEY

Dallas L. Peck, *Director*



---

For additional information write to:

District Chief
U.S. Geological Survey
333 W. Nye Lane, Room 203
Carson City, NV 89706

Copies of this report may be purchased from:

U.S. Geological Survey
Books and Open-File Reports Section
Federal Center
Box 25425
Denver, CO 80225

# CONTENTS

## ILLUSTRATIONS

## TABLES

## CONVERSION FACTORS AND VERTICAL DATUM

| Multiply | By | To obtain |
|---|---|---|
| centimeter (cm) | 0.3937 | inch |
| centimeter per day (cm/d) | 0.03281 | foot per day |
| cubic centimeter ($cm^3$) | 0.06102 | cubic inch |
| cubic meter ($m^3$) | 35.31 | cubic foot |
| cubic meter per second ($m^3/s$) | 35.31 | cubic foot per second |
| gram (g) | 0.002205 | pound |
| gram per cubic centimeter per centimeter ($g/cm^3/cm$) | $2.373 \times 10^{-6}$ | pound per cubic foot per foot |
| gram per square centimeter per second ($g/cm^2/s$) | $2.373 \times 10^{-6}$ | pound per square foot per second |
| gram per cubic centimeter ($g/cm^3$) | 62.43 | pound per cubic foot |
| kilometer (km) | 0.6214 | mile |
| liter (L) | 0.2642 | gallon |
| megapascal (MPa) | 10 | bar |
| meter (m) | 3.281 | foot |
| millimeter (mm) | 0.03937 | inch |
| millimeter per hour (mm/hr) | 0.07874 | foot per day |
| square centimeter ($cm^2$) | 0.155 | square inch |
| square centimeter per second ($cm^2/s$) | 0.001076 | square foot per second |
| square meter ($m^2$) | 10.76 | square foot |
| square millimeter per second ($mm^2/s$) | 0.00155 | square inch per second |
| square kilometer ($km^2$) | 0.3861 | square mile |

For temperature, degrees Celsius (°C) can be converted to degrees Fahrenheit (°F) by using the formula °F = [1.8(°C)] + 32.

### SEA LEVEL

In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929, formerly called "Sea-Level Datum of 1929"), which is derived from a general adjustment of the first-order leveling networks of both the United States and Canada.

# Sediment Properties and Water Movement Through Shallow Unsaturated Alluvium at an Arid Site for Disposal of Low-Level Radioactive Waste Near Beatty, Nye County, Nevada

*By Jeffrey M. Fischer*

## ABSTRACT

A commercial disposal facility for low-level radioactive waste has been in operation near Beatty, Nevada, since 1962. The facility is in the arid Amargosa Desert where wastes are buried in trenches excavated into unsaturated alluvial sediments. Thick unsaturated zones in arid environments offer many potential advantages for disposal of radioactive wastes, but little is known about the natural movement of water near such facilities. Thus, a study was begun in 1982 to better define the direction and rates of water movement through the unsaturated zone in undisturbed sediments near the disposal facility. This report discusses the analyses of data collected between 1983 and 1988.

Precipitation during the period October 1984 through February 1988 was 326 millimeters, and measurements of water content of the sediments indicate that the precipitation did not penetrate a depth of more than 1 meter. Surficial deposits at the study site are a sandy silt with a low hydraulic conductivity. This deposit acts as a barrier to infiltration of precipitation. In addition, a clean gravel layer beneath the surficial deposits acts as a capillary barrier and further impedes downward percolation of water.

Thermocouple psychrometers were used to measure water (pressure) potentials and temperatures in the dry alluvial sediments. The psychrometers were installed through 2.5-centimeter-diameter holes drilled a minimum of 3 meters horizontally outward from a 1.5-meter-diameter vertical shaft. Psychrometers were installed between depths of 3 and 13 meters below land surface. Thermocouple psychrometers were also installed in a borehole between depths of 1 and 12 meters below land surface and near-surface psychrometers were installed in shallow excavations.

Water potentials measured during the study period generally ranged from -5.5 megapascals at a depth of 1.6 meters below land surface to -3.2 megapascals at a depth of 13 meters. Seasonal variations in water potentials and temperatures were observed to a depth of about 9 meters. Below a depth of 9 meters, observed water potentials and temperatures did not vary seasonally. Calculations of water flux indicate little water movement through interconnected pores, either as liquid or vapor. Depending on the season, water movement in the upper 9 meters is either upward toward the surface or downward. Between the depths of 9 and 13 meters, water movement was consistently upward. The rate of liquid water movement through interconnected pores was estimated to range from $1\times10^{-2}$ to $1\times10^{-9}$ centimeter per day. This range reflects the large uncertainty in the estimate of unsaturated hydraulic conductivity. The rate of water movement through vapor diffusion generally ranged from $1\times10^{-3}$ to $1\times10^{-6}$ centimeter per day.

The upper 13 meters of sediments at the study site can be divided into three zones on the basis of water potential, water content, temperature, and sodium-chloride content. The upper zone, from land surface to a depth of about 2 meters, is subject to rapid changes in water potentials, water contents, and temperatures in response to precipitation and air-temperature fluctuations. A middle zone, from a depth of about 2 meters to 9 meters, is characterized by unchanging water content and seasonal trends in water potential as vapor moves upward and downward in response to changes in soil temperature. This zone is also characterized by an accumulation of sodium chloride that may be related to the maximum depth of water percolation into the sediments. The lower zone, from a depth of about 9 meters to at least 13 meters, is characterized by relatively unchanging water potentials, water contents, and temperatures. Gradients in the lower zone indicate that water movement, both as liquid and vapor, is upward.

# INTRODUCTION

The general procedure for disposing low-level radioactive waste (LLRW) in the United States has been to bury waste in shallow trenches (Fischer, 1986, p. 1). Typical LLRW is expected to remain hazardous for about 500 years; thus, the waste needs to be isolated from the environment for a long time. A major threat to the integrity of waste isolation is the infiltration of precipitation (Richardson, 1962, p. 211; Fischer, 1986, p. 5; Schulz and others, 1987, p. 2; Schulz and others, 1988, p. 1; Bedinger, 1989, p. 20). Radioactive constituents, dissolved and mobilized by water, can lead to contamination of the environment. Because of the long life of radioactive waste, it is necessary to rely on the hydrogeologic characteristics of a LLRW disposal facility to limit water contact with the waste (Fischer, 1986, p. 4). Arid sites with a thick unsaturated zone are believed to minimize the risk of environmental contamination because the amount of precipitation is low, and little, if any, precipitation is thought to recharge the underlying ground water. Also, sediments in the thick unsaturated zone are expected to slow contaminant movement by adsorbing many of the contaminants (Wollenberg and others, 1983, p. 198).

The U.S. Geological Survey has studied most of the commercial LLRW facilities in the Nation. The purpose of these studies was to evaluate current LLRW facilities and to develop criteria for the selection of future sites. These studies have taken on added importance since Congress passed the Low-Level Radioactive Waste Policy Act (PL 96-573) in 1980. In response to the Act, many states are now considering establishing new LLRW disposal facilities.

The U.S. Geological Survey began studies in 1976 at an arid LLRW disposal facility in the Amargosa Desert, 17 km southeast of Beatty, Nev. (fig. 1). The studies were designed to determine rates and directions of water movement through unsaturated sediments and to assess the potential for contaminant movement. Initial studies at the facility (completed in 1980) concluded that under certain climatic conditions, high rainfall during periods of low evapotranspiration and high sediment water contents, deep percolation of water might occur in areas of bare soil (Nichols, 1987, p. 50). A second study that began in 1982 was designed to define more accurately the rates and direction of water movement through unsaturated deposits beneath an undisturbed area.

The objectives of this latter study were to:

- Determine factors that influence the timing, rates, and directions of water movement through the upper 13 m of undisturbed sediments.
- Calculate rates, quantities, and directions of water movement through the unsaturated zone.
- Further characterize the geology near the LLRW disposal facility.
- Obtain more detailed information on the ground-water flow system beneath the LLRW disposal facility.
- Collect data on current climatic conditions that can be used in conjunction with long-term National Weather Service data to estimate the relations between precipitation, evaporation, and deep percolation.

Because recharge at the site was anticipated to be intermittent and unpredictable--requiring large amounts of precipitation to occur at times when water contents in the sediments are high and evapotranspiration rates are low--the study was planned to last a minimum of 5 years.

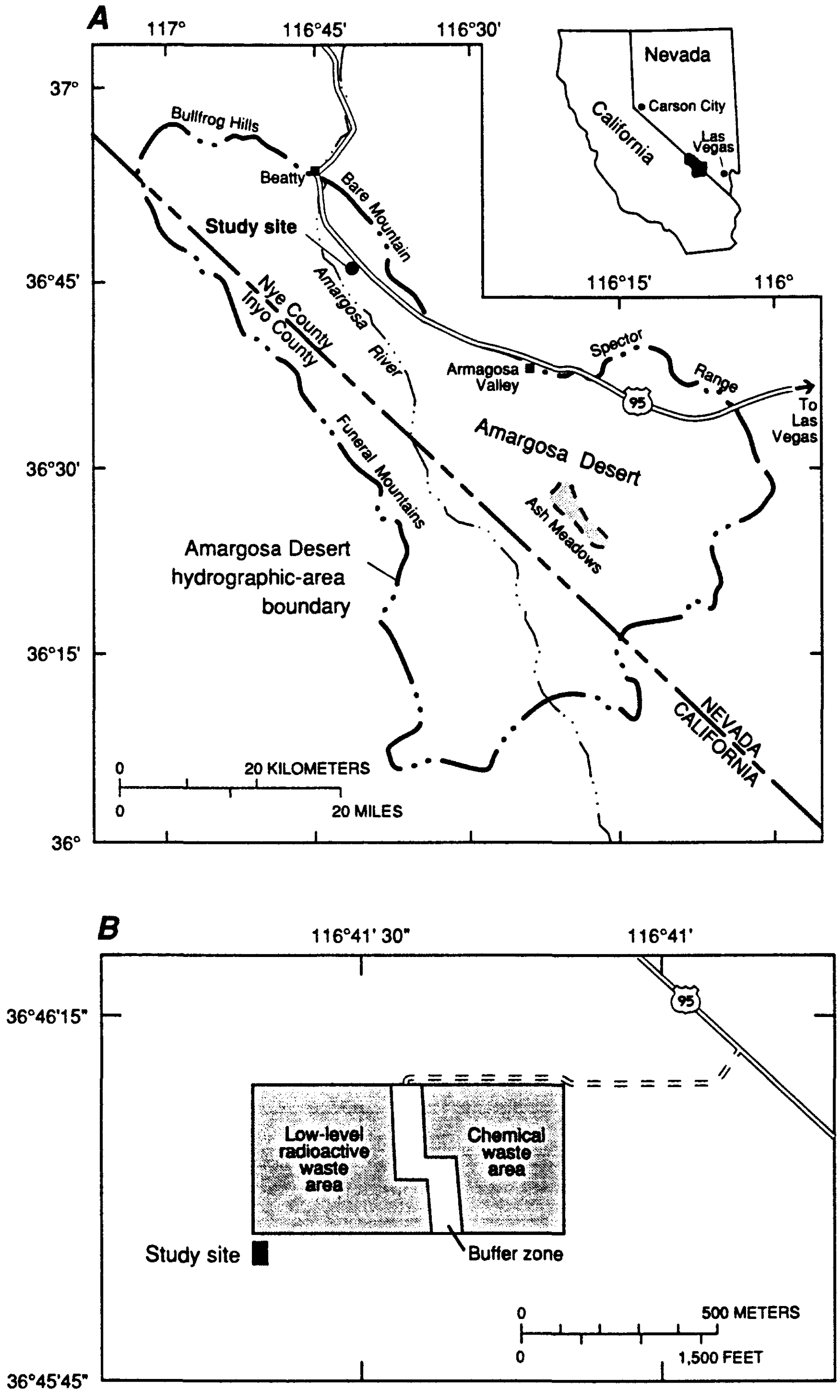



FIGURE 1.--(*A*) Location of study site within the Amargosa Desert hydrographic area, and (***B***) location of study site in relation to waste-desposal areas. Hydrographic area from Harrill and others (1988).

## Purpose and Scope

This report is a summary of a U.S. Geological Survey study at the LLRW disposal facility from 1983 to 1988. Location, geology, and hydrology are briefly discussed in the first part. Methods for calibrating and installing thermocouple psychrometers used to measure water potentials and temperatures in the upper 13 m of unsaturated sediments are presented in the second part. Sediment properties of the upper 14 m of unsaturated sediments are described in the third part of the report and include physical and hydraulic properties, concentration of salts, and thermal diffusivity. Water movement within the upper 13 m of unsaturated sediments is discussed in the last part, which also includes a discussion of water potentials and estimates of liquid-water and water-vapor fluxes.

## Acknowledgments

Support and encouragement by the Branch of Nuclear Waste Hydrology, Low-Level Radioactive Waste Program of the U.S. Geological Survey is gratefully acknowledged. The author cannot possibly acknowledge everyone who assisted in the work and whose assistance at various stages of the project was extremely helpful. Thanks are extended to the U.S. Bureau of Land Management and the Nevada Department of Human Resources, Health Division, for permission to do the study on their land. Personnel at US Ecology, Inc., operators of the nearby waste-disposal facility, are acknowledged for their cooperation and for providing data from their monitoring wells. The author appreciates the help of William D. Nichols and David S. Morgan, both with the U.S. Geological Survey, and Dale Hammermeister, formerly with the U.S. Geological Survey, for their technical assistance in the design of instrumentation to monitor water movement through the unsaturated zone at the study site.

## Setting of Waste-Disposal Facility

The waste-disposal facility is in the Amargosa Desert, 17 km southeast of Beatty and 170 km north of Las Vegas, Nev. (fig. 1). The facility encompasses more than 240,000 $m^2$; the eastern half is used for chemical waste and the western half for low-level radioactive waste. The study site area is outside the waste-disposal facility near the southwest corner. The study site area encompasses 870 $m^2$ and includes a monitoring shaft, an instrument borehole, three neutron-probe access tubes, and a weather station (fig. 2).

Since 1962, approximately 113,000 $m^3$ of LLRW has been buried in shallow trenches at the waste facility. The size of the trenches has changed over time and ranges from 90 to 240 m long, 2 to 15 m deep, and up to 35 m wide. The waste consists mostly of dewatered and solidified forms of nontransuranic byproduct. Waste packaging is not intended to provide containment after burial; the burial itself is considered sufficient for containment. The current burial procedure consists of stacking waste in layers 3 m thick and covering each layer with a half meter of backfill. Trenches are not capped but are backfilled to one-half meter above grade. Backfill consists of an uncompacted mixture of alluvium previously removed from the trenches. Waste is stacked to within a meter of the land surface.

### *Geology*

The Amargosa Desert is a northwest-trending valley in the Basin and Range Province. It is bounded by block-faulted mountains composed of lower Paleozoic rocks and Tertiary volcanic rocks. The valley is formed by normal faulting along the mountain fronts. Moderate to steep sloping alluvial fans have formed at the foot of the mountains, and the central part of the valley slopes gently to the southeast. The waste-disposal facility is located in the gently sloping part of the valley about 1 km west of the toe of an alluvial fan. The valley in this area is 13 km wide.

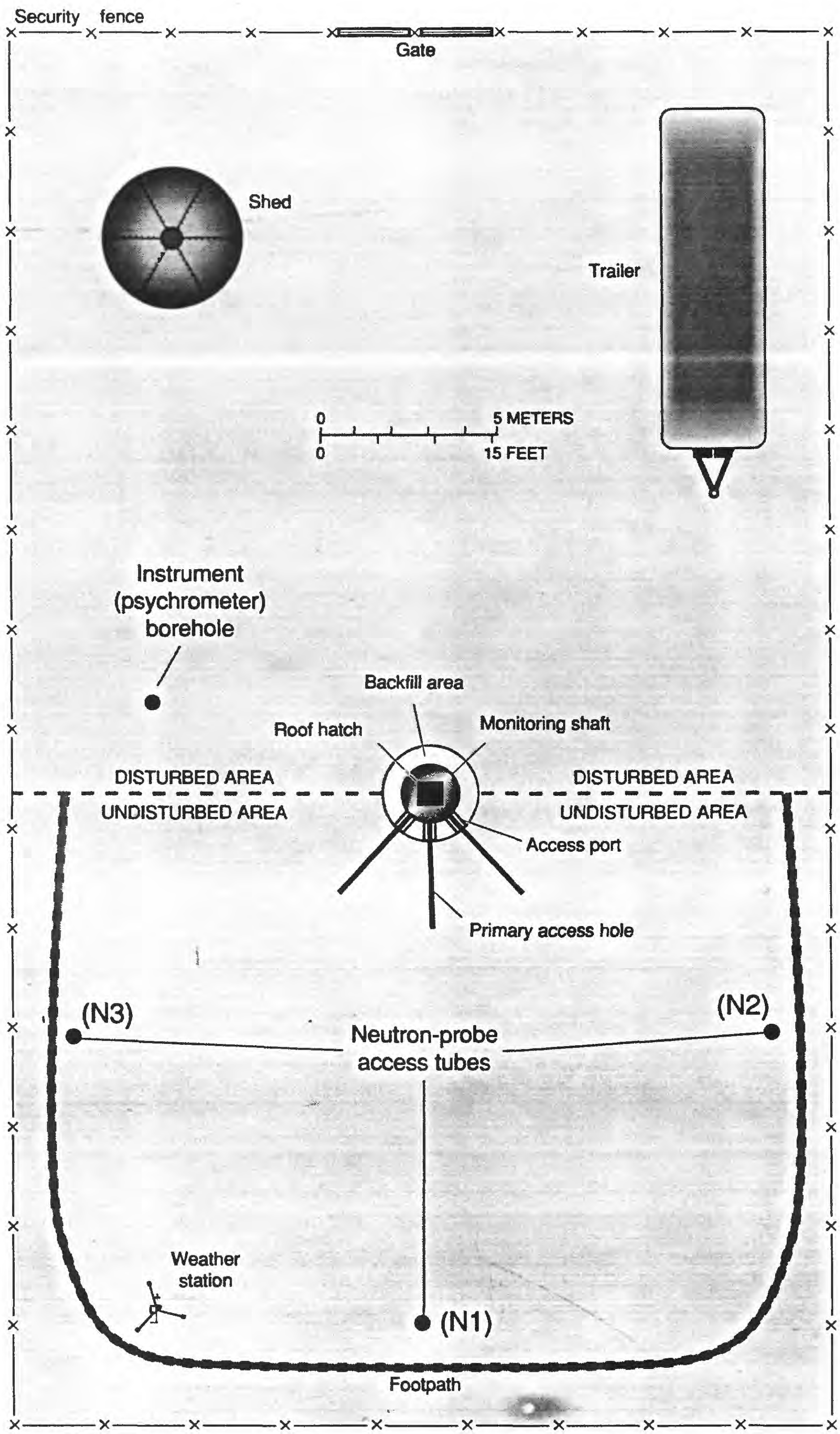



FIGURE 2.--Location of monitoring shaft, instrument (psychrometer) borehole, weather station, and neutron-probe access tubes. Location of study site is shown in figure 1.

The waste-disposal facility is underlain by more than 175 m of unconsolidated alluvial-fan, fluvial, and marsh deposits (Clebsch, 1968, p. 78-80). Depth to ground water is generally 85 to 115 m below land surface. To further define the geology of the unsaturated zone, monitoring wells 301, 302, and 303 (fig. 3) were logged using neutron-moisture, natural-gamma, and gamma-gamma geophysical instruments in 1983. Logs were not calibrated for casing size or water content. Thus, the logs could not be used to calculate density, porosity, or water content. However, the logs provide useful information about the types of units present, and their continuity, thickness, and variations in density. Clay lenses within the sediments were identified on the basis of low-density readings and high water contents. Conversely, gravel layers were identified on the basis of low-density readings, low water contents, and low natural-gamma readings. Additional information about the geology was obtained during the drilling of more monitoring wells at the waste-disposal facility between 1985 and 1988.

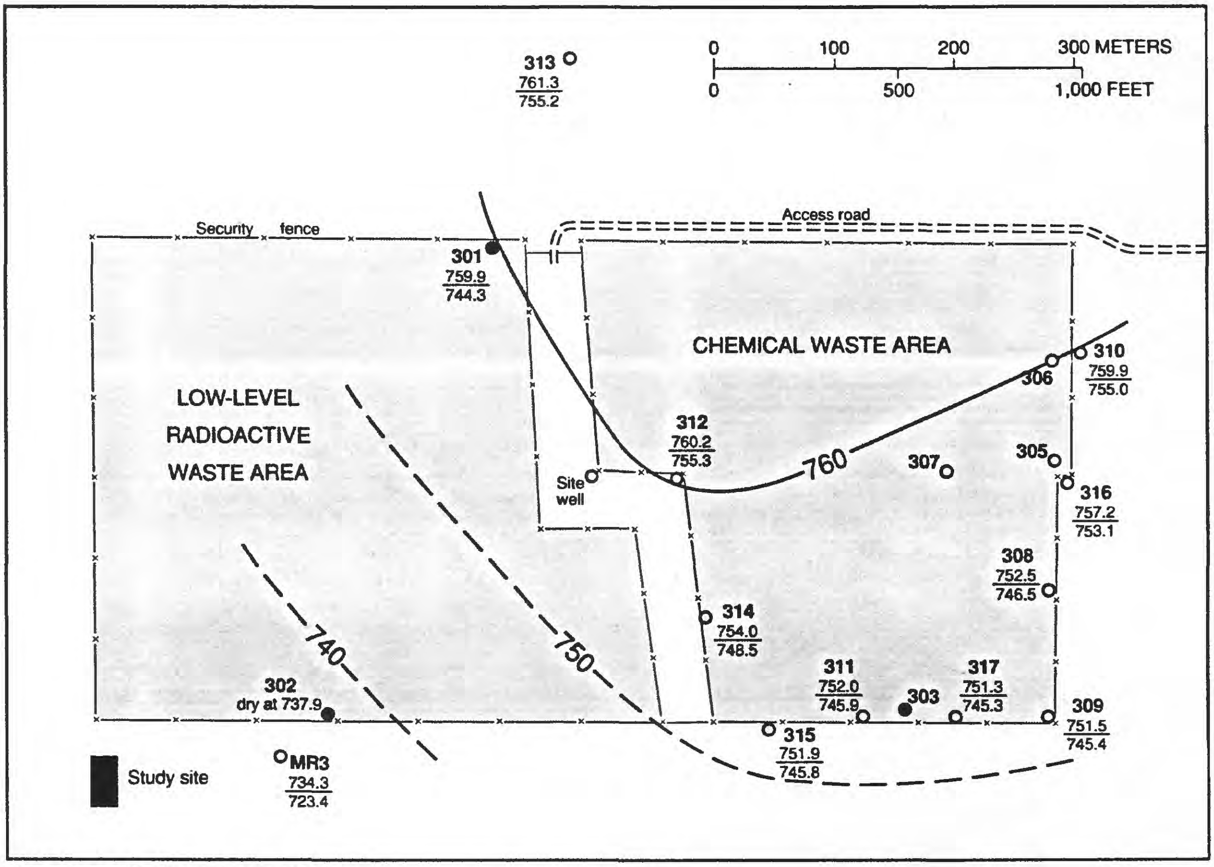
Surficial deposits of sandy silt are about a half-meter thick and are areally extensive within the Amargosa Desert. These deposits are commonly covered with a veneer of imbricated gravels (desert pavement). Nichols (1987, p. 8-9) previously described, on the basis of one well log, the subsurface geology at the LLRW disposal facility as consisting of unconsolidated poorly sorted fluvial cobbles, gravels, sands, and silts from a meter below land surface to a depth of 30 m, and fanglomerate and debris flow deposits from a depth of 30 to 80 m below land surface. Additional lithological and geophysical logs indicate that fluvial and debris-flow deposits in the area of the LLRW disposal facility differ in thickness and areal extent (fig. 4). For example, thin clay lenses were observed in the log of well 303 near depths of 10, 20, and 25 m that were not observed in the logs of other wells. Likewise, a clay lens observed in the log of well 302 at a depth of about 37 m was not observed in the logs of other wells. The most consistent unit was a clay layer, probably a marsh deposit, detected at depths ranging from 70 to 90 m. The unit has a thickness of about 20 m and was observed in all the well logs. On the basis of altitude of the top of the clay, it was determined to dip to the south-southeast at approximately 0.07 m/m. The direction of dip is similar to that shown by Nichols (1987, fig. 5). Alternating layers of clean and muddy sands predominate below the clay layer to a depth of about 125 m below land surface. Below a depth of about 125 m, gravelly sands extend to a depth of at least 200 m (Mark Group, 1989, p. 7).

## *Climate*

The Amargosa Desert is one of the driest areas in the Nation. Mean annual precipitation ranges from 114 mm at Beatty to 74 mm at Amargosa Valley--formerly Lathrop Wells (Nichols, 1987, p. 15). Seasonal and spatial variation can be considerable. Average monthly estimates of potential evapotranspiration at Beatty range from about 40 mm in December to 330 mm in July (Nichols, 1987, p. 26).

Vegetation at the study site is sparse with the dominant plant being the creosote bush (*Larrea tridentata*). Its roots are generally shallow, usually extending less than 1.7 m below land surface (Barbour and others, 1977, p. 61). Concentration of creosote-bush roots near land surface indicates the plants are opportunistic--active only when soil moisture is available. Most roots observed during the digging and drilling at the study site were within the upper 2 m of soil. Some roots, however, were observed to a depth of 7 m below land surface, but it is uncertain whether these roots were active or fossilized.

A Penman weather station was installed in August 1984 at the study site to monitor solar radiation, windspeed, relative humidity, air temperature, soil-heat flux, and precipitation. During the study period, data were collected each minute and averaged hourly and daily with a Campbell Scientific, Inc., model CR21 datalogger. The data were transmitted daily via telephone communications to a computer where the transmitted data were checked and processed. For this report, the two parameters of interest were precipitation, measured with a tipping bucket rain gage, and air temperature, measured with a thermistor. Precipitation data indicated how much water was available to infiltrate (assuming no overland flow) and temperature data indicated when evaporative demand was high.




Base from US Ecology, Inc., 1990.

EXPLANATION

——750— — GROUND-WATER LEVEL CONTOUR — Shows altitude of water levels as determined from monitoring wells. Contour interval is 10 meters. Dashed where approximately located. Datum is sea level

313 O 761.3 / 755.2 GROUND-WATER MONITORING WELL — Three-digit numbers and "MR3" are well identifications. Number above line is water-surface altitude, in meters, for December 1988. Number below line is well-bottom altitude, in meters. Datum is sea level. Water level at well MR3 measured by U.S. Geological Survey; all other wells measured by US Ecology, Inc.

FIGURE 3.--Waste-desposal areas showing location of monitoring wells and approximate altitude of ground-water levels in December 1988. Solid circles indicate wells for which geophysical logs are available.

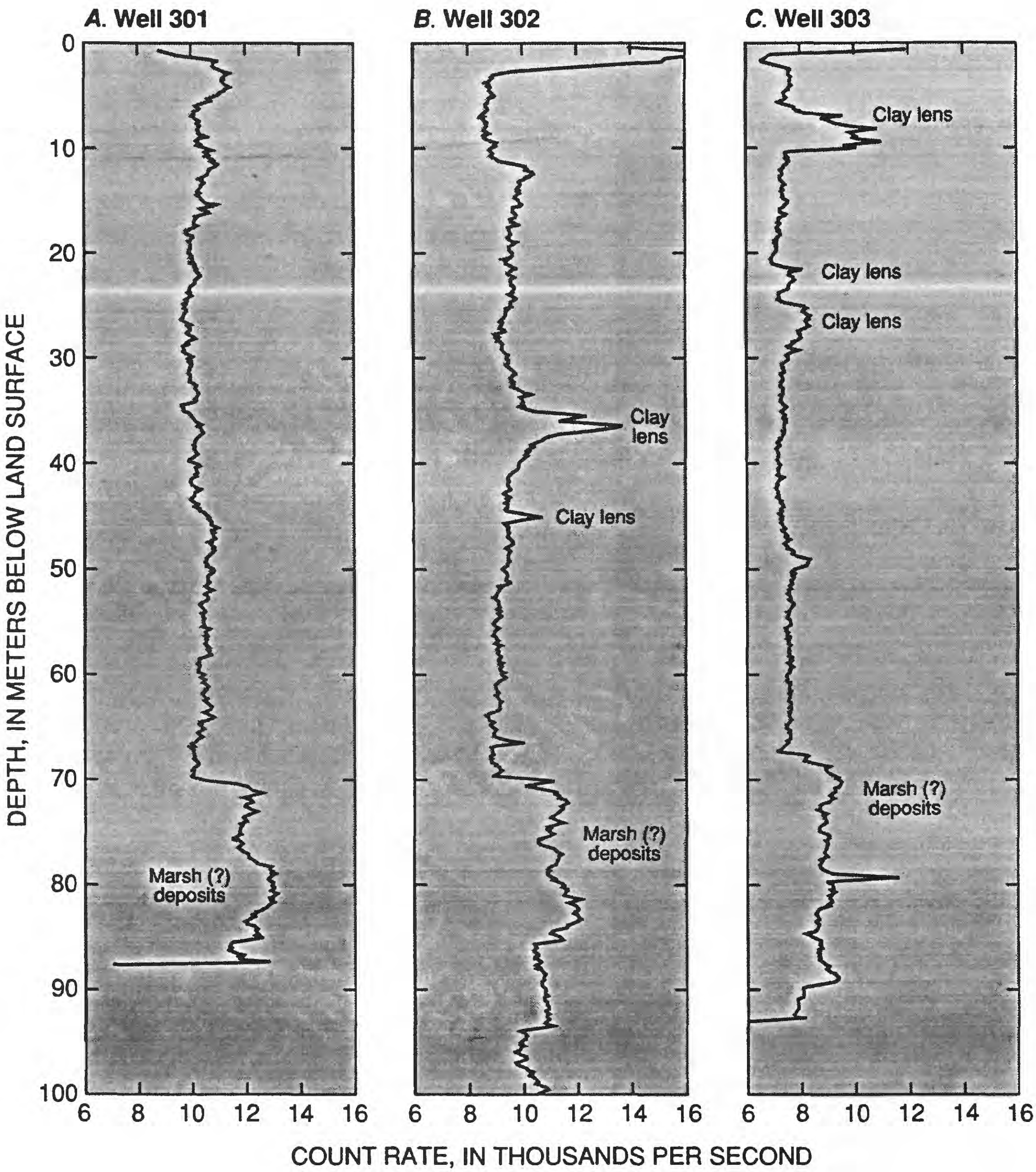



FIGURE 4.--Distribution of clay lenses and marsh (?) deposits as interpreted from gamma-gamma logs and geologic information from three monitoring wells. Interpretations of clay lenses and marsh (?) deposits are also based on natural-gamma and neutron-moisture logs, which are not shown. Location of wells is shown in figure 3.

Precipitation for 1985, 1986, and 1987 was 35.0, 74.3, and 114.4 mm, respectively (table 1). Although the years 1985 and 1986 were drier than the average, they are well within the range of variations described by Nichols (1987, p. 16). Total precipitation during the period from October 1984 through February 1988 was 326 mm. Most of the precipitation measured during the study occurred between the months of November and March, usually in the form of rain. Typically, storms last for less than a day and produce little precipitation. Precipitation did not exceed 38 mm for any 24-hour period and total precipitation did not exceed 43 mm for any month during the period of measurement (table 1). Total precipitation for a given storm never exceeded 40 mm and only exceeded 10 mm in six storms (fig. 5). The pattern of precipitation during the period from October 1984 to February 1988 fits within the range of historical data (Nichols, 1987, p. 18).

The mean yearly air temperature, from January 1985 through December 1987, was approximately 19°C. Mean daily temperatures were typically between 25 and 34°C during the summer (June through August), whereas mean daily temperatures were typically between 2 and 15°C during the winter (December through February). The study site is typical for a desert with hot summer days and cool nights. Maximum temperatures during summer days are usually between 40 and 45°C, whereas summer nights are about 20 to 25°C cooler (fig. 6). Maximum temperature during winter months is usually between 10 and 20°C, whereas minimum temperatures are usually between 5 and -5°C. Temperatures measured at the site from 1985 through 1987 are comparable to those measured over the past 40 years at Beatty and Amargosa Valley--formerly Lathrop Wells (Nichols, 1987, p. 22).

TABLE 1.--*Total monthly precipitation and number of days per month having precipitation at study site from October 1984 through February 1988*

[Values are in millimeters, rounded to nearest 0.01 millimeter; minimum measurable precipitation was 0.254 millimeter. Symbol: --, month not included in study period discussed herein.]

| | Total precipitation and, in parentheses, number of days having precipitation | | | | |
|---|---|---|---|---|---|
| Month | 1984 | 1985 | 1986 | 1987 | 1988 |
| January | -- | 2.29 (3) | 14.97 (4) | 11.92 (3) | 13.45 (4) |
| February | -- | .00 (0) | 8.37 (3) | 6.33 (4) | 9.13 (2) |
| March | -- | .00 (0) | 9.37 (6) | 32.74 (7) | -- |
| April | -- | .00 (0) | 6.34 (2) | 4.82 (2) | -- |
| May | -- | 5.33 (1) | 1.77 (1) | 5.06 (6) | -- |
| June | -- | 5.07 (2) | .00 (0) | .76 (1) | -- |
| July | -- | 1.52 (2) | 8.62 (3) | 1.00 (3) | -- |
| August | -- | .00 (0) | 3.04 (3) | 7.87 (1) | -- |
| September | -- | 4.06 (1) | .50 (1) | .00 (0) | -- |
| October | 0.00 (0) | .00 (0) | 3.30 (2) | 9.64 (2) | -- |
| November | 42.67 (3) | 11.66 (5) | 8.88 (2) | 21.08 (3) | -- |
| December | 37.33 (9) | 5.08 (1) | 9.14 (3) | 13.19 (3) | -- |
| Total by year | -- | 35.01(15) | 74.30(30) | 114.41(35) | -- |

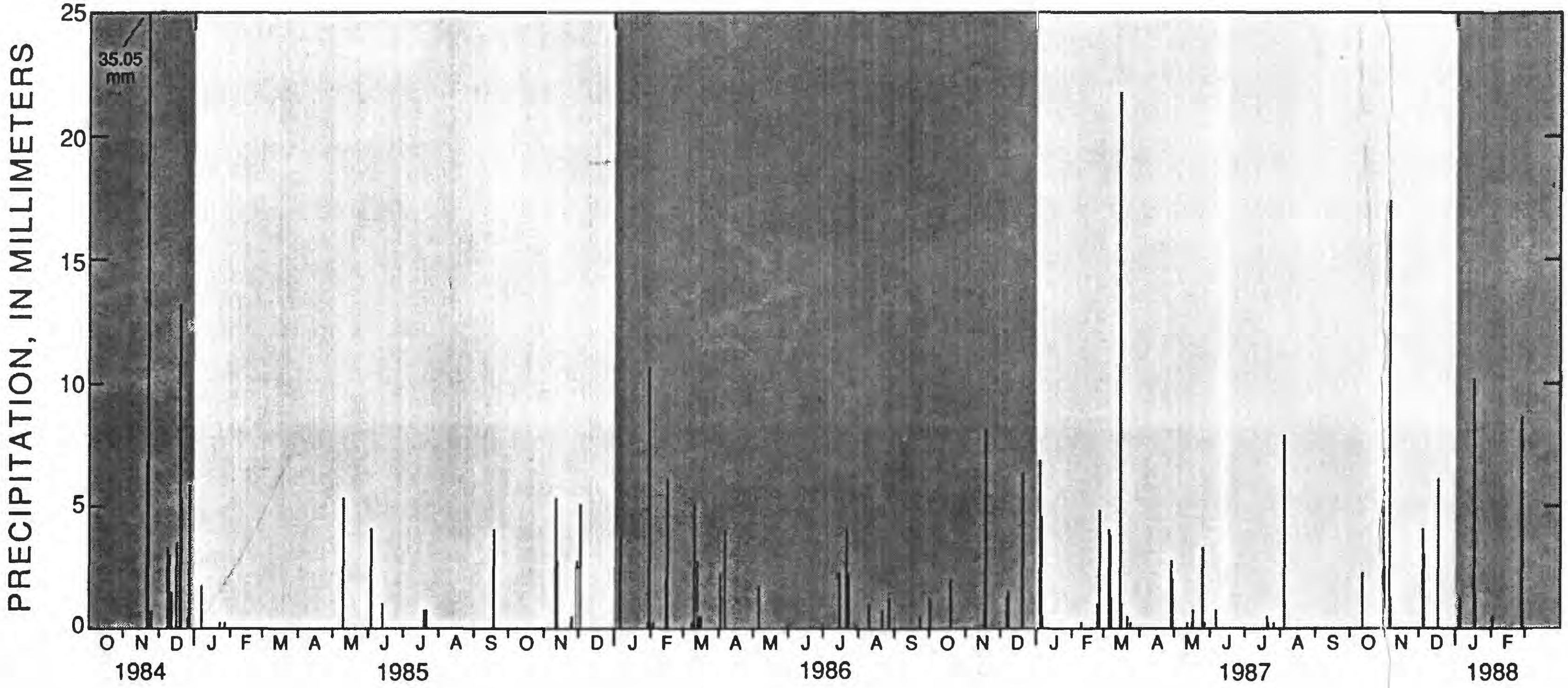



FIGURE 5.--Daily totals of precipitation measured at study site from October 1984 through February 1988.

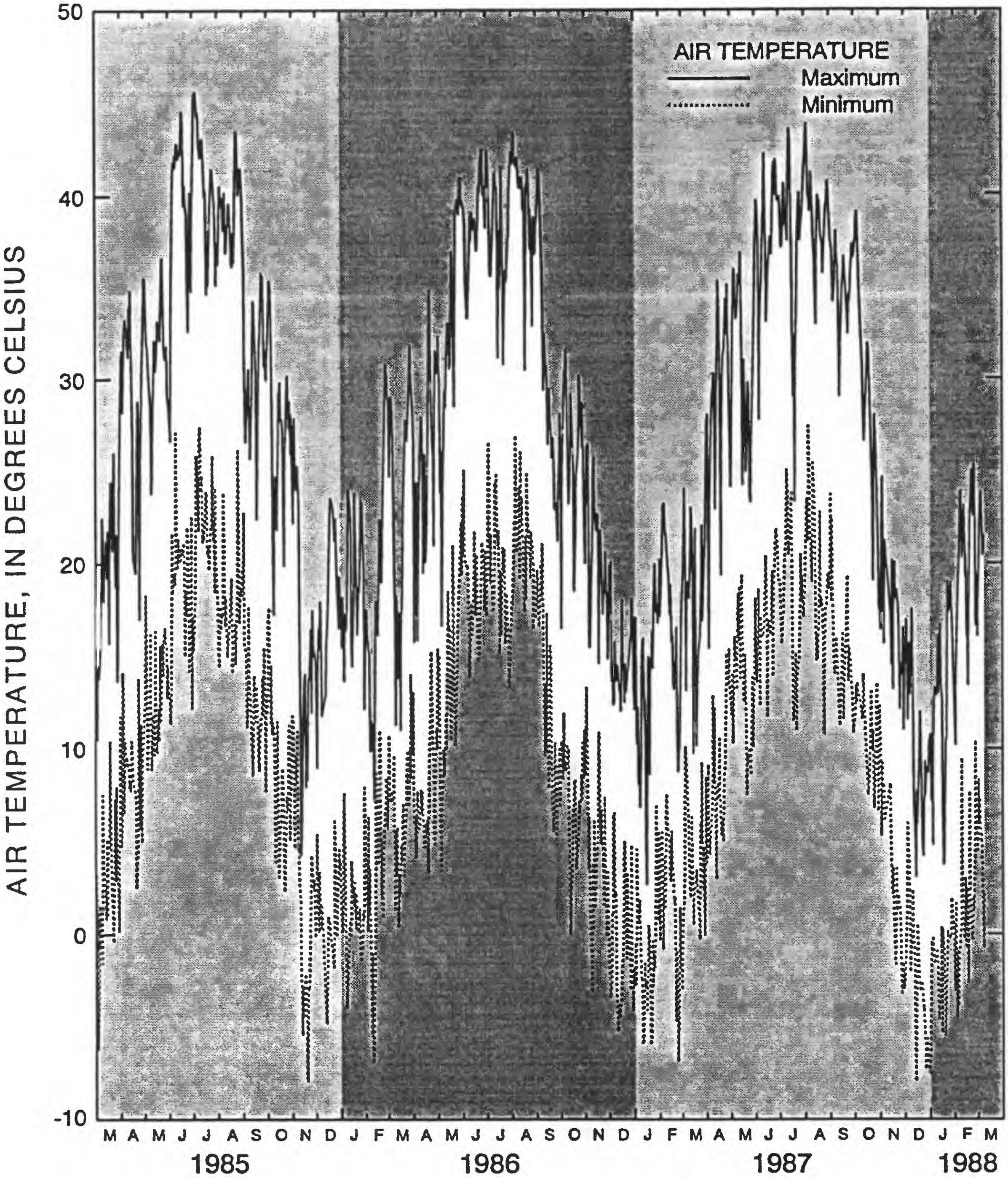



FIGURE 6.--Daily maximum and minimum air temperatures at study site from March 1985 through March 1988.

## *Hydrology*

The relatively flat surface of the central Amargosa Desert slopes gently to the southeast. It is dissected by numerous small dry washes. The main drainage, the Amargosa River, is incised as much as 4 m into the valley floor and approaches to within 3 km of the LLRW disposal facility. The river is dry most of the time near the latitude of the study site (Nichols, 1987, p. 11). Most channels in the flat central part of the Amargosa Valley drain to the southeast. However, the alluvial fans northeast of the facility drain to the southwest off the mountain fronts. Flow off the fans is captured by southeast-trending channels about 0.5 km east of the LLRW disposal facility. In most years, these channels have no flow. However, once in August 1983, flash flooding was observed at the LLRW disposal facility and at the study site.

Depth to water in monitoring wells in the vicinity of the LLRW disposal facility ranges from about 85 to 115 m below land surface. All wells are perforated or screened in unconsolidated basin-fill deposits. A carbonate-rock aquifer may underlie the unconsolidated deposits near the LLRW disposal facility (Winograd and Thordarson, 1975). The general direction of ground-water flow across the LLRW disposal facility, on the basis of water-level measurements in 13 wells during December 1988, is toward the southwest (fig. 3). The apparent water-level gradient is approximately 0.06 m/m. In contrast, the general direction of ground-water flow across the chemical-waste disposal facility is toward the south with an apparent water-level gradient of approximately 0.04 m/m. The reasons for differences between apparent ground-water flow directions and gradients across the two facilities are unknown.

The general slope of the potentiometric surface at the LLRW disposal facility is consistent with the general potentiometric surface in basin-fill deposits, which generally slopes parallel to the surface gradient of the Amargosa River (Kilroy, 1991, p. 9). The general slope of the potentiometric surface at the LLRW disposal facility suggests that ground-water flow is from the north.

## Concepts of Unsaturated Flow

The unsaturated zone is defined as the area of the subsurface where water in the pores is at less than atmospheric pressure. Usually, both liquids and gases occupy the pore space within the sediments. In this report, the movement of liquid water through interconnected pores is referred to as liquid-water flow, whereas the movement of water as a gaseous vapor is referred to as water-vapor flow. The amount of water in the unsaturated zone varies and can be described in several ways. Gravimetric water content $(\theta_g)$ is the mass of water per mass of sediment (gram per gram) and volumetric water content $(\theta_v)$ is the volume of water per volume of sediment (cubic centimeter per cubic centimeter). Volumetric water content is always less than the porosity except at saturation where it equals the porosity.

Flow of liquid water in the unsaturated zone can be described by a modified form of Darcy's Law that accounts for the fact that hydraulic conductivity is not constant (Hillel, 1982, p. 113-114). For one-dimensional vertical flow of liquid water in the unsaturated zone, Darcy's Law can be written as:

$$q_w = -K(\psi)\frac{d\phi}{dz} \qquad (1)$$

where $q_w$ = specific discharge, in centimeters per day;
$\phi$ = hydraulic head, in centimeters;
$z$ = elevation above an arbitrary datum, in centimeters; and
$K(\psi)$ = unsaturated hydraulic conductivity, in centimeters per day.

Unsaturated hydraulic conductivity is not constant but is a hysteretic function of the pressure head $(\psi)$, water content $(\theta)$, and sediment wetting history. It is written as $K(\psi)$ or $K(\theta)$ to show this relation. Usually, hysteresis is ignored by assuming a monotonic relation of $K(\theta)$ to $\psi$ for wetting or drying sediments. The hydraulic head in eq. 1 is the sum of two components: an elevation head ($z$) and a pressure head $(\psi)$. In the unsaturated zone, the pressure head has a negative value and is referred to in this report as the water potential $(\psi_w)$. Water potential in the unsaturated zone is comprised of a matric potential $(\psi_m)$ and the solute potential $(\psi_s)$ (Hanks and Ashcroft, 1980, p. 21-27). Matric potential is related to adsorptive forces in the sediment and solute potential is a function of solute concentrations and temperature. A number of units are used to express water potential. The unit used in this report is megapascals, with 1 MPa equal to 9.87 atmospheres. For dry sediments, water potential greatly exceeds elevation head. For this reason, elevation head is ignored and eq. 1 can be rewritten as:

$$q_w = -K(\theta)\frac{d\psi_w}{dz} \qquad (2)$$

where $K(\theta)$ is assumed to be monotonically related to $\psi_w$.

## USE OF THERMOCOUPLE PSYCHROMETERS TO DETERMINE WATER POTENTIALS

Screen-covered thermocouple psychrometers (TCP) were used to measure water potentials at the study site. TCP's do not measure water potentials directly, but rather relate relative humidity in sediments to water potential via the Kelvin equation (Weibe and others, 1971). The instrument operates by applying an electrical current to a thermocouple junction to cool it. Water from the surrounding sediment condenses on the cooled junction until the cooling current is stopped. After cessation of the cooling current, water on the junction begins to evaporate and an electric current is created via the Peltier effect. The strength of the current is related to temperature and water potential of the sediment. Studies have shown most TCP's have a linear response to water potential between -0.5 and -6.5 MPa for a given temperature (Brown and Collins, 1980). Below this range the electrical current readings drop off because Peltier cooling can no longer condense water on the wet junction (Weibe and others, 1971, p. 22). This means that a TCP voltage reading may correspond to two different water potentials, as shown in figure 7A. The correct water potential for a given TCP can be determined by looking at the width and height of the response curve (fig. 7B). Lower water potentials have a short response time, whereas higher water potentials have a long response time.

### Calibration

Thermocouple psychrometers used at the study site were calibrated in the laboratory over sodium-chloride solutions of varying concentrations. The calibration procedure is outlined by Brown and Collins (1980) and Brown and Bartos (1982). All TCP's were constructed with 18 m of lead wire and then calibrated in the laboratory. Calibration of the TCP's with the 18 m of wire to be used in the field was necessary because adding the wire following calibration could change the calibration results. For calibration and field use, a cooling current of 8 milliamps was applied to each TCP for 30 s. Voltage readings were taken 1 s after the cessation of the cooling current. This procedure was repeated at temperatures of 15, 22, and 29°C, and over salt solutions with water potentials ranging from -0.4 to -7.8 MPa (fig. 8). The calibration temperatures and water potentials were within the range of those expected in the field.

Thermocouple psychrometers that did not produce a linear calibration curve between -0.4 and -7.8 MPa were not used in the field. Repetition of the calibration procedure for 10 of more than 100 calibrated TCP's showed the calibration procedure was accurate to within about 5 percent for a given water potential (±0.2 at -4.0 MPa). Voltage readings in the field were converted to water potentials using individual TCP calibration curves. Initially, the shape of the TCP response curve was examined to ensure that the readings were within the calibration range. A four-point interpolation was performed using the nearest calibration points that represented the temperature and voltage readings above and below those measured in the field.

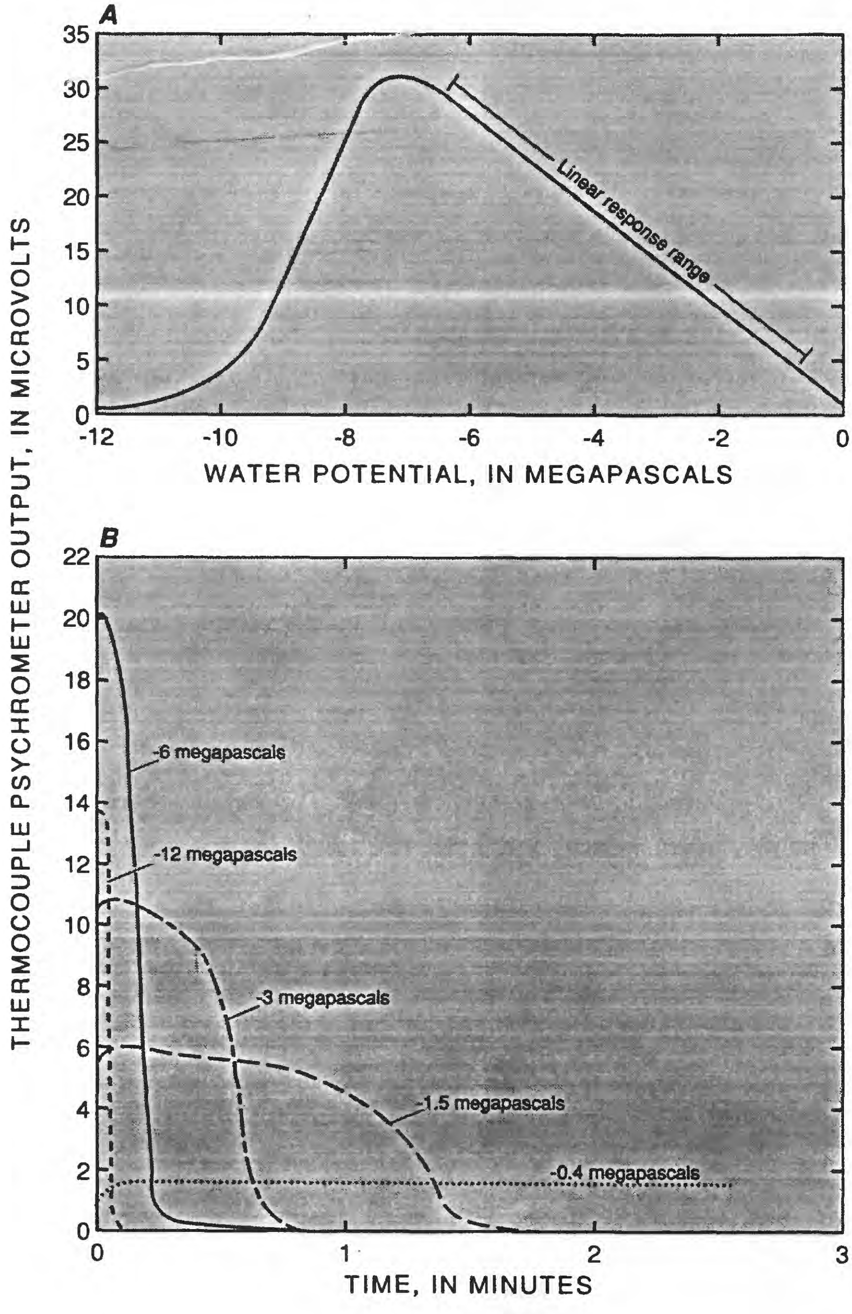



FIGURE 7.--Typical electric response of thermocouple psychrometers using a cooling time of 15 seconds: (*A*) response at various potentials (modified from Wiebe and others, 1971, figure 10), and (*B*) response curves for selected water potentials over time (modified from Wiebe and others, 1971, figure 9).

## Installation

A 13.7-m deep monitoring shaft was designed to allow access to an undisturbed vertical profile within the unsaturated zone for the purpose of measuring water potentials beneath an undisturbed, naturally vegetated area. Installation of the TCP's through access holes drilled horizontally outward from the shaft avoided the disturbance of a vertical sediment column above the instruments. This method was chosen over a vertical drill hole in which instruments are placed within backfilled materials because materials above the instruments could greatly alter natural moisture conditions and hydraulic properties even if carefully placed (Morgan and Fischer, 1984, p. 621).

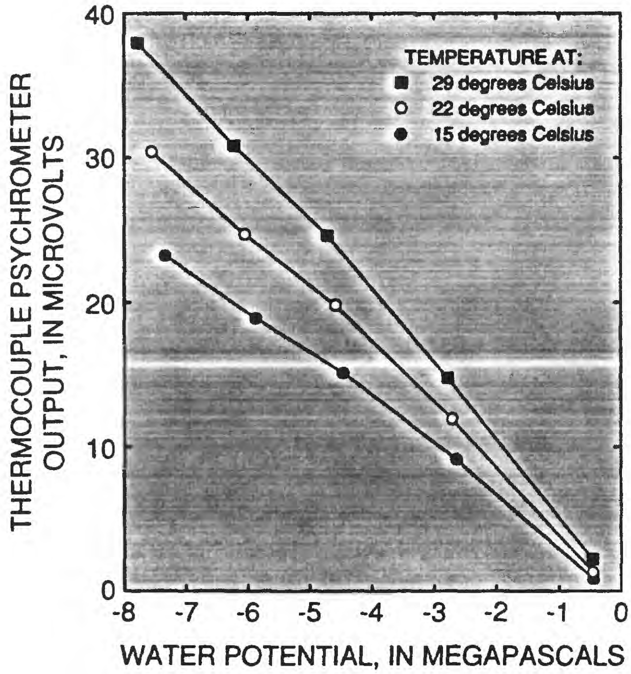



FIGURE 8.--Typical calibration curves for a thermocouple psychrometer calibrated over different sodium-chloride (salt) solutions with equivalent water potentials ranging from -0.4 to -7.8 megapascals and at temperatures of 15, 22, and 29 degrees Celsius.

The vertical monitoring shaft was installed just outside the southwest corner of the LLRW disposal facility in 1983 (figs. 1 and 2). The shaft was not located within the LLRW disposal facility because of safety concerns. The shaft is 1.52 m in diameter and extends 13.7 m below land surface. The prefabricated steel shaft was emplaced in a hole measuring 2.44 m in diameter that was excavated by a crane drill using bucket and flight augers. The annulus outside the shaft was backfilled with dry sand and the surface of the excavation was sealed with concrete. One quadrant of the monitoring shaft contains 33 lateral ports--3 ports at 11 levels between depths of 3 and 13 m (fig. 9). TCP's installed in the monitoring shaft are designated with L (left), M (middle), and R (right), followed by a number from 1 to 11 that indicates level in the shaft. These ports allow access to the undisturbed sediments outside the monitoring shaft. The access ports face an area with natural vegetative cover and undisturbed surface conditions (fig. 2).

TCP's were installed in the shaft through secondary access holes between depths of 3 and 13 m. Secondary access holes (fig. 10) were drilled laterally out from the monitoring shaft access ports with a blast-hole drill. These holes were drilled through the dry and unstable stony alluvium to a minimum distance of 3 m from the shaft to minimize the effect of temperature fluctuations in the shaft on TCP readings. A similar monitoring shaft at the Nevada Test site had induced lateral temperature gradients and seasonal fluctuations as much as 1.8 m from the shaft wall (E.P. Weeks, U.S. Geological Survey, oral commun., 1983). Removal of the cuttings presented a problem because the addition of a liquid in the dry sediments might interfere with TCP readings for years. For this reason cuttings were removed with air, which dried the sediments but allowed for reequilibration in several months. When cuttings were too large to be blown out of the secondary access holes, they were mechanically removed with a long rod. The secondary access holes were cased with 2.5-cm-diameter polyvinylchloride (PVC) pipe sealed in place with polyurethane foam (fig. 10). The foam provided a thermal and vapor barrier.

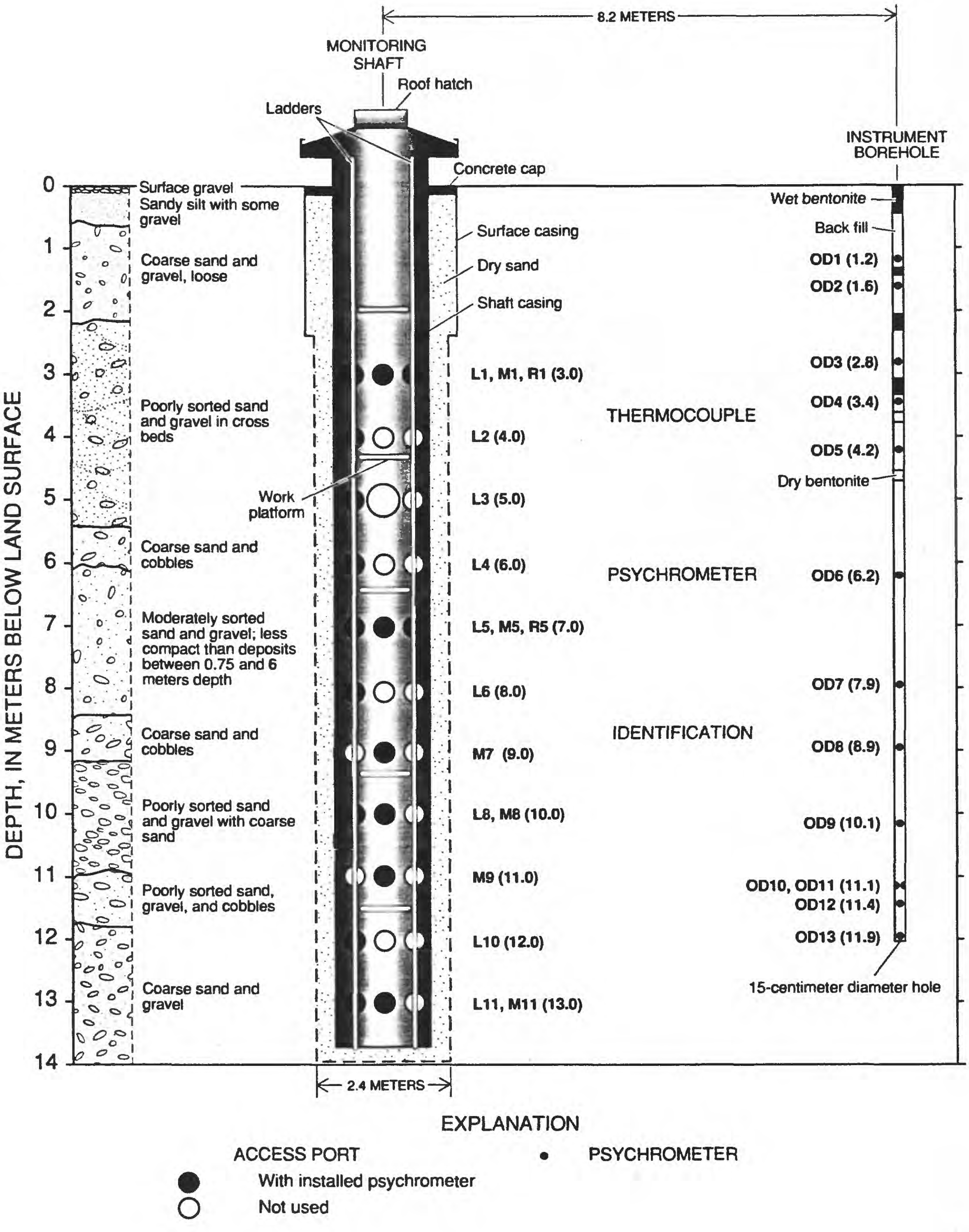



FIGURE 9.--Monitoring shaft and instrument borehole showing locations of thermocouple psychrometers. Location of monitoring shaft and instrument borehole shown in figure 2. View of monitoring shaft looking south. Distance between monitoring shaft and borehole not to scale. Thermocouple psychrometers installed in the monitoring shaft are designated with L (left), M (middle), and R (right), followed by a number from 1 to 11 that indicates level in shaft. Thermocouple psychrometers installed in the instrument borehole are designated OD1 through OD13. Depth of each thermocouple psychrometer, in meters below land surface, is given in parentheses.

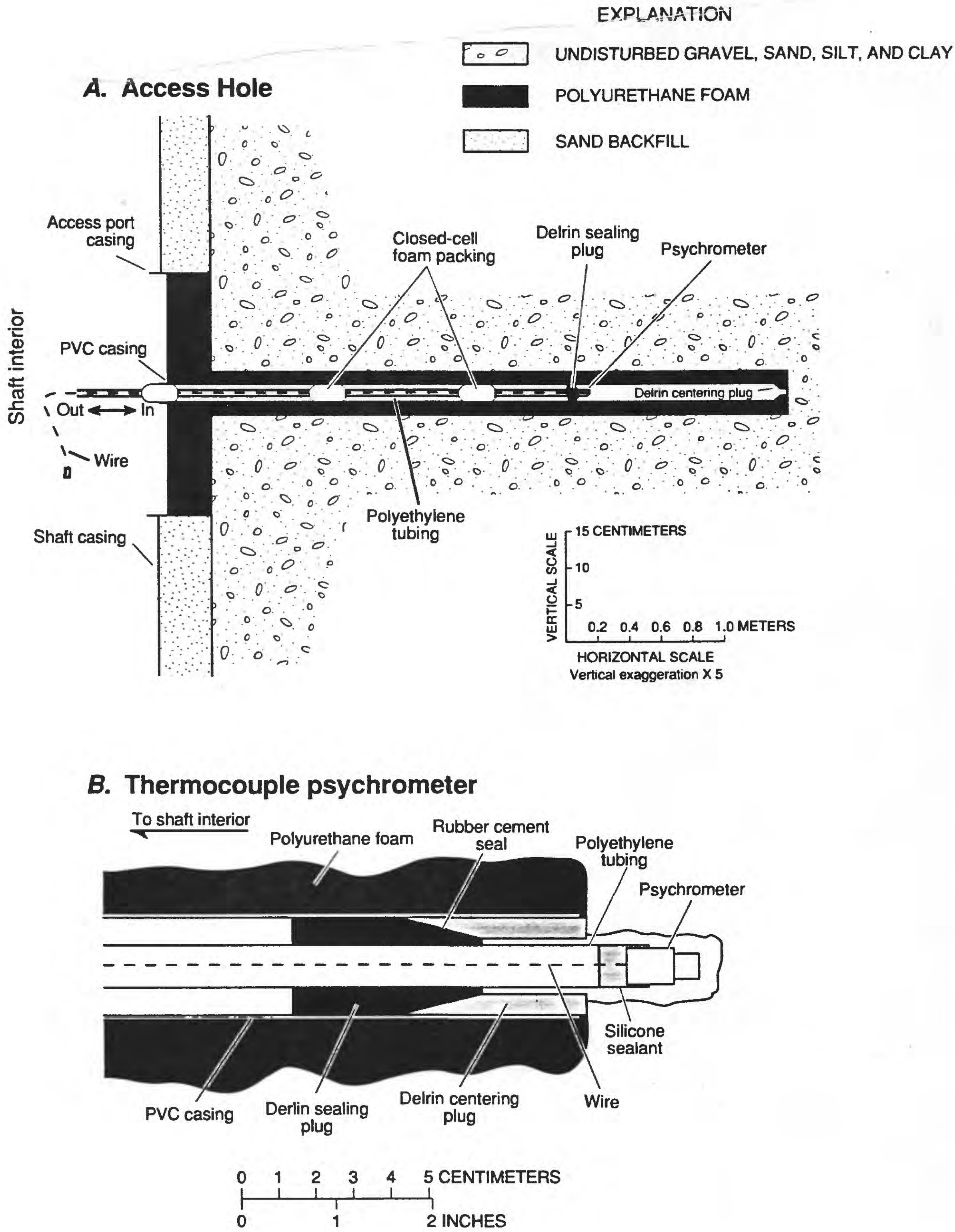



FIGURE 10.--Diagrams of: (*A*) completed access hole, and (*B*) installed thermocouple psychrometer at end of access hole. (Modified from Morgan and Fischer, 1984, figures 3 and 4.) PVC, polyvinylchloride.

Use of screen-covered TCP's for long-term monitoring of water potentials under field conditions has been criticized because of possible soil and microorganism contamination through the screen covering and because the calibration of TCP's can shift over time (Brown and Johnston, 1976, p. 995). For this reason a unique method for installing TCP's was developed and reported by Morgan and Fischer (1984, p. 624-627). The installation method allowed for TCP removal, recalibration, and replacement. The PVC pipe, installed laterally into the sediments as previously described, was left open at the end to allow access to the sediments. TCP's were mounted on the end of 9.5-mm-outside-diameter tubing composed of semirigid polyethylene. The TCP wires were threaded through the tubing until only the top of the TCP was exposed at the end. Silicon sealant injected into both ends of the polyethylene tubing was used to hold the TCP's in place and also used to seal the tubing (fig. 10). The polyethylene tubing was used to push the TCP's into, and pull them out of, the 2.5-cm-diameter PVC pipe (fig. 10). Closed foam packing wrapped around the outside of the polyethylene tubing approximately every 90 cm was used to isolate the TCP within the 2.5-cm-diameter PVC pipe, and was used as a thermal barrier between the TCP and the shaft. The end of the semirigid polyethylene tubing with the TCP was fitted with a coneshaped sealing plug made of Delrin, an industrial grade plastic. This plug was designed to fit precisely into a centering plug (also made of Delrin) at the end of the PVC pipe to form a tight seal (fig. 10). Initially, a rubber o-ring was used to seal between the two PVC plugs. Thus, TCP's were isolated from the shaft by both thermal barriers (the foam packing and plugs) and vapor barriers (the silicone sealant and rubber o-ring).

Effectiveness of the rubber o-ring was tested in the laboratory prior to field installation. The seal was first tested by applying pressure to a section of PVC pipe in which a TCP had been installed. No leaks were detected during the tests. A second test consisted of installing two TCP's through 50-cm lengths of PVC pipe that penetrated into a sealed box containing dry sand. The readings from these two TCP's were compared to two other TCP's that had been buried in the sand. After a 4-week equilibration period, all four TCP's agreed to within 0.3 MPa of each other. Following an additional 2 weeks of continued monitoring, two sodium-chloride solutions, one having an equivalent water potential of 0 MPa and the other having an equivalent water potential of -10 MPa, were placed inside each of the two PVC pipes to test for possible leaks in the TCP installation through the PVC pipe. Even after 2 months, all readings from the TCP's remained within 0.3 MPa of each other. Thus, the TCP's in the pipes were not affected by the salt solutions, an indication the seal was effective.

TCP's were installed in the monitoring shaft in late 1984 and early 1985. Initial installation was limited to a depth of 6 m so the installation procedure could be further tested. Duplicate TCP's were installed at a depth of 3 m. On the basis of initial TCP readings, several possible problems were identified. These included air transfer along the annular space outside the PVC pipe caused by poor cleaning of the drill hole prior to polyurethane foam emplacement and failure of the o-ring after repeated replacement of the TCP's. In subsequent installations, pebbles were completely removed from the hole prior to foam emplacement and rubber cement was used instead of the o-ring to seal the two plugs together. The TCP's in the upper 6 m of the monitoring shaft were replaced in 1986 and installation of TCP's to a depth of 13 m in the monitoring shaft was completed in 1987. Laboratory tests on the replaced TCP's showed that, prior to cleaning, some had changed calibration by as much as -0.5 MPa. After cleaning, half the TCP's returned to their previous calibration curves, which indicated that periodic cleaning was necessary. However, three TCP calibration curves were permanently altered, probably because electrical connections had been resoldered in the field. Subsequently, broken wires were not resoldered in the field. Instead the whole TCP was replaced and the soldered TCP was recalibrated.

Several TCP's were installed in a vertical borehole in July 1986 for comparison with TCP's installed in the shaft. The borehole was drilled to a depth of 12 m using the ODEX air-hammer drilling method. The borehole is about 8 m west-northwest of the monitoring shaft (fig. 2) in an area void of vegetation. TCP's were placed in the cased borehole at approximately 1-m intervals between depths of 1 and 12 m (fig. 9). TCP's installed in the instrument borehole are designated OD1 through OD13, with OD1 being the shallowest and OD13 being the deepest. The hole was backfilled with cuttings as the casing was withdrawn. Except for TCP's above the depth of 6 m, no seals were placed between TCP's for fear the addition of liquid would interfere with water potentials in the borehole. TCP's installed from depths of 1 to 4 m had small amounts of wetted bentonite placed between them to prevent preferential movement of wetting fronts down the borehole. Data on water potentials from TCP's installed in both the borehole and shaft were read and stored at least every 4 hours using a Campbell Scientific CR-7 datalogger. These data were routinely retrieved and stored for analysis on a minicomputer via telecommunications.

## SEDIMENT PROPERTIES

The upper 14 m of alluvium was studied in detail within the study site because sediments in this area could be reached easily and because the 14-m depth approximately coincides with the maximum depth of the trenches. In addition, most of the daily and yearly variations in water potential, moisture content, and temperature occur within this depth interval. When possible, data from the deeper sediments or sediments outside the study site were also collected.

### Physical Properties

The stratigraphy of the upper 14 m of alluvium at the study site is shown in figure 11. Lithologic descriptions in figure 11 are based on samples collected during drilling of neutron-probe access tubes, instrument borehole, monitoring shaft, and ground-water monitoring wells, and from geophysical logs. The surface deposits consist of desert pavement (imbricated gravels) overlying a 0.5- to 1-m thick deposit of sandy silt. Beneath the sandy silt, and extending to a depth of 2.2 m, is an incompetent, moderately sorted, coarse sandy gravel that is relatively free of fine-grained matrix. The sandy silt and gravel layers appear to extend over the entire LLRW disposal area. Between the depths of 2.2 and 5.2 m, the sediments consist of steeply crossbedded cobbly and gravelly coarse sands. The crossbeds grade from a poorly sorted sand at the top to gravels at the base. A poorly sorted cobble, sand, and gravel layer underlies the crossbeds at a depth between 5.2 and about 6.0 m. Below the cobbles, between the depths of 6.0 and 8.5 m, is a moderately sorted, medium sand with some gravel. A cobble and gravel layer with little fine-grained matrix underlies the sand and gravel at an approximate depth between 8.5 and 9.0 m. This layer grades into a sandy, cobbly, poorly sorted gravel that extends to a depth of about 11.8 m. Between the depths of 11.8 and 14 m, the sediments are mostly coarse sands with gravel. Except for the upper 2 m, the stratigraphy is specific to the study site. Deposits within the LLRW facility, although similar, show variations in texture typically associated with fluvial deposition.

Three neutron-probe access tubes (fig. 2), consisting of 13-cm diameter steel pipe, were installed during 1984 using the ODEX air-hammer drilling method (Hammermeister and others, 1985). The ODEX drilling method was used because it was quick, it did not add liquids to the formation that could interfere with instrument readings, and because it provided a stable cased hole that could be easily sampled. Two of the access tubes (N1 and N3 in fig. 2) were installed to a depth of 14 m and the third (N2 in fig. 2) to a depth of 31 m. In 1986, an instrument borehole was also drilled using the ODEX drilling method to a depth of 14 m. The stony alluvium was difficult to sample; however, solid tube drive cores, 61 cm long and 10 cm in diameter, were driven approximately every 1.5 m while drilling the 14-m deep access tubes and the instrument borehole. Approximately 17 unconsolidated and 8 semi-consolidated core samples were obtained during this study. It was not possible to determine how representative these samples were of sediments at a given depth owing to the limited number of core samples. Furthermore, the samples are skewed toward the finer grained sediments because the coarser grained gravels and cobbles were too large to fit inside the core barrel.

Textural properties of the core samples, such as grain-size distribution, porosity, and bulk density, were determined in the laboratory. These properties are needed for the determination of moisture-retention characteristics of the sediments. In wet sediments with high water potentials, moisture retention depends primarily on the capillary effect and the pore-size distribution (Hillel, 1982, p. 76). As water potentials decrease, moisture retention is due increasingly to adsorption, and is, thus, dependant on texture and specific surface area of the sediments.

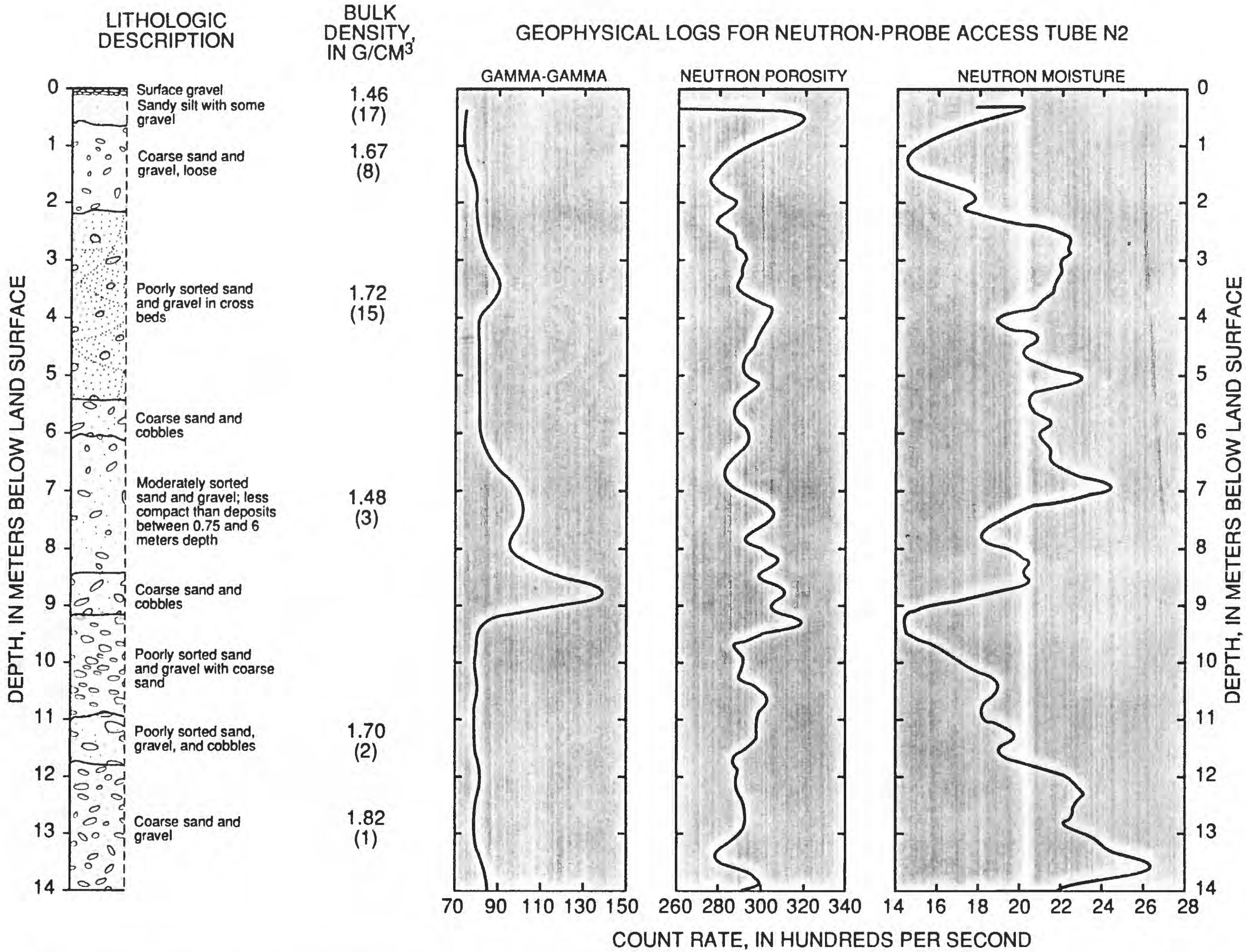



FIGURE 11.--Generalized stratigraphy and selected geophysical logs run during December 1984 in neutron-probe access tube N2 at study site. Average bulk density is listed for each lithologic unit for which data are available. Averages include values from table 2 and from William D. Nichols (U.S. Gerological Ssurvey, written commun., 1978). Number in parentheses below the average is number of bulk-density determinations. Abbreviation: G/CM3, grams per cubic centimeter.

Grain-size distribution was determined by first weighing a dry-bulk sample and then sieving the sediments through a 64-mm screen to remove gravels. After the gravel fraction was weighed, a 100-g split of the remaining sample was wet sieved to separate the sand fraction from the silt and clay fractions. The sand was dried, mechanically sieved, and weighed to determine the sand-size distribution. Particle-size distributions for the silt and clay fractions were done using pipette analysis (Folk, 1974).

In general, the sediments collected below a depth of 1 m, from neutron-probe access tubes N1 and N3 and the instrument borehole, were poorly sorted sandy gravels or gravelly sands. Samples ranged from 30 to 58 percent sand and 17 to 47 percent gravel (table 2). The silt and clay content is greatest in the upper 0.75 m of sediments and averaged 64 percent for 10 samples. The silt and clay content of nine samples collected between the depths of 0.75 and 7.0 m below land surface averaged about 8 percent, whereas the silt and clay content of six samples between the depth of 7.9 and 13.1 m averaged only 3 percent.

TABLE 2.–*Laboratory results of particle-size distribution, mean-grain and bulk densities, and porosity of sediment samples collected to a depth of about 13 meters below land surface*

[Samples to depth of 2 meters were collected during shallow excavations. Samples below depth of 2 meters were collected from cores taken during drilling of holes for neutron-probe access tubes and instrument borehole. Abbreviations: g/cm$^3$, grams per cubic centimeter; mm, millimeters; --, not determined]

| | Particle size, percentage of-- | | | | | Density (g/cm$^3$) | | | |
|---|---|---|---|---|---|---|---|---|---|
| | | | | | | Mean-grain | | | |
| Depth below land surface (meters) | Gravel (>64 mm) | Gravel (2-64 mm) | Sand (0.063-2 mm) | Silt (0.004-0.063 mm) | Clay (<0.004 mm) | Size fraction (>2mm) | Size fraction (<2mm) | Bulk | Porosity (percent) |
| 0.00-0.75[a] | 0 | 8 | 28 | 53 | 11 | 2.34 | 2.39 | 1.46 | 30.3 |
| 0.75-2.2[b] | 10 | 46 | 35 | 6 | 2 | 2.36 | 2.39 | 1.67 | 34.1 |
| 2.44 | 8 | 35 | 50 | 5 | 2 | 2.34 | 2.35 | 1.76 | 27.0 |
| 3.97 | 18 | 17 | 54 | 8 | 3 | 2.31 | 2.39 | 1.60 | 33.6 |
| 5.19 | 30 | 32 | 32 | 4 | 1 | 2.44 | 2.46 | -- | -- |
| 5.49 | 19 | 40 | 33 | 4 | 4 | 2.35 | 2.40 | -- | -- |
| 7.02 | 7 | 36 | 49 | 4 | 4 | 2.32 | 2.34 | 1.40 | 42.9 |
| 7.93 | 15 | 47 | 35 | 2 | 0 | 2.41 | 2.34 | 1.58 | 34.4 |
| 8.54 | 10 | 33 | 53 | 3 | 1 | 2.48 | -- | 1.47 | 39.1 |
| 10.07 | 38 | 38 | 22 | 2 | 0 | 2.53 | 2.31 | -- | -- |
| 11.29 | 21 | 45 | 30 | 2 | 1 | 2.52 | 2.46 | 1.61 | 33.2 |
| 11.59 | 15 | 46 | 35 | 3 | 1 | 2.45 | 2.46 | 1.79 | 25.7 |
| 13.12 | 8 | 29 | 58 | 3 | 2 | 2.34 | 2.47 | 1.82 | 25.5 |

[a] Values are averages of 10 samples, except bulk density, which is an average of 17 samples.
[b] Values are averages of 4 samples, except bulk density, which is an average of 8 samples.

Core samples collected during this study and a previous study (unpublished data provided by W.D. Nichols, U.S. Geological Survey, Carson City, Nev.) were used to determine bulk density $(\rho_b)$. Porosities ($n$) of the cores were calculated from bulk-density estimates using the equation:

$$n=1-\rho_b/\rho_s \qquad (3)$$

where $\rho_s$ is the density of the solid sediment grains, or grain density. Grain densities were determined using a water-displacement technique. Gamma-gamma density logs and neutron-porosity logs (fig. 11) were run on the neutron-probe access tubes within the study area to extend the range of the point data on bulk density and porosities.

Geophysical logs and laboratory data indicate that, except for surficial sediments and the sand and gravel and coarse sand and cobble layers between the depths of about 6 and 9 m, bulk density of the sediments was generally uniform to a depth of 14 m (fig. 11). Bulk density of the sand and gravel layers between the depths of 6 m and about 8.5 m averaged 1.48 g/cm$^3$, whereas the gravel layer between the depths of 0.75 and 2.2 m averaged 1.67 g/cm$^3$ and the sandy silt above a depth of about 0.75 m averaged 1.46 g/cm$^3$ (fig. 11). The average bulk density of the remaining five samples was 1.72 g/cm$^3$. Mean grain densities averaged 2.38 g/cm$^3$ and were fairly uniform for all grain sizes and depths (table 2). Thus, changes noted in the gamma-gamma logs, between the depths of about 6 and 9 m, indicate that the sand and gravel and sand and cobble layers in the interval have higher porosities than the surrounding sediments. On the basis of the laboratory data, the porosity of the surficial deposits of sandy silt is 30 percent (table 2), the porosity of the gravel between the depths of about 0.75 and 2.2 m is 34 percent, and the porosity of the sand and gravel and sand and cobbles between the depths of about 6 and 9 m is 40 percent. Porosity of the sediments for the other intervals is about 28 percent.

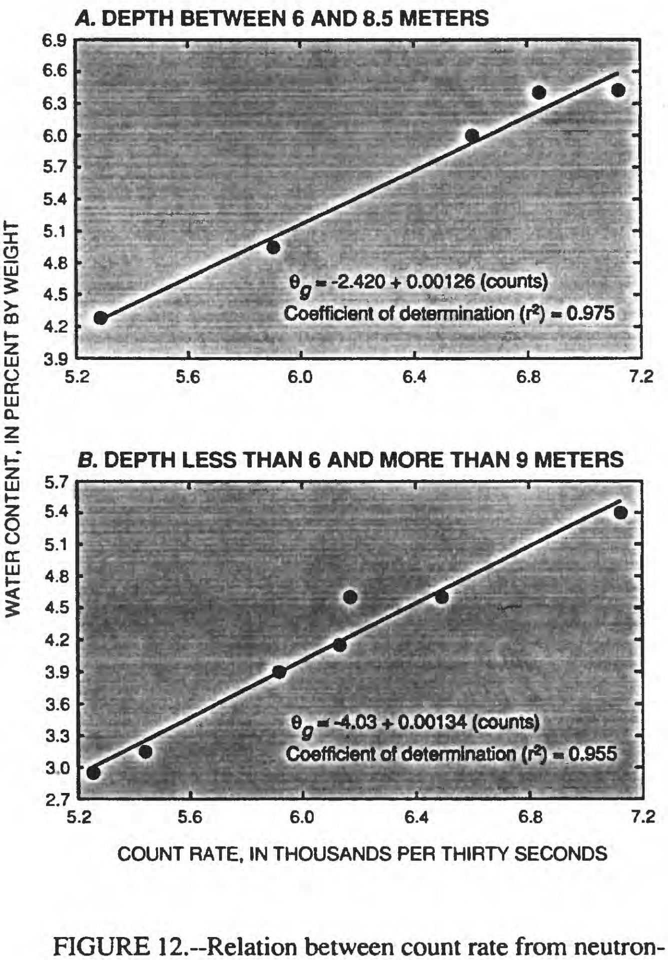



FIGURE 12.--Relation between count rate from neutron-moisture probe and gravimeterically determined water content ($\theta_g$) of samples collected (*A*) between depths of 6 and 8.5 meters below land surface and (*B*) at depths of less than 6 meters and more than 9 meters below land surface.

## Water Contents

Measurements of sediment water contents were made for three reasons: to determine changes in water storage within the unsaturated zone, to measure depth of wetting front advances, and to use in estimating unsaturated hydraulic conductivity. Measurements of water contents were usually made on a monthly basis; however, more frequent measurements were made during infiltration events. Except for the upper half meter of deposits, measurements of water contents were made with a neutron-moisture probe (Campbell Pacific Nuclear Corporation model 503). Measurements of water content of the top 50 cm of sediments were determined periodically by collecting samples and using a standard laboratory procedure of weighing the samples, oven drying them at 110°C for 24 hours, and reweighing them.

The neutron-moisture probe was calibrated with 12 samples collected while drilling the neutron-probe access tubes. Samples were split into at least two subsamples and weighed in the field. Gravimetric water content for each subsample was determined later in the laboratory following oven drying using a procedure reported by Gardner (1965). An average gravimetric water content was then determined for each sample. Immediately following access-tube installation, nine neutron-moisture probe readings were made at each depth where a core sample had been taken. (Three readings were taken 15 cm above the depth of the core sample, three readings at the depth of the core sample, and three readings 15 cm below the depth of the core sample.) The nine neutron-moisture probe readings were averaged and plotted against the average gravimetric water content of each sample. On the basis of this data, the less dense sediments (as determined from bulk densities) between a depth of about 6 and 9 m (fig. 11) have a different calibration curve than sediments above and below that depth interval. The two regression curves, shown in figure 12, have coefficient of determination (r-squared) values of 97.5 and 95.5 percent, respectively. For volumetric water content used in calculating water storage, the gravimetric water content was converted to volumetric water content using the following equation:

$$\theta_v = (\rho_b / \rho_w) * \theta_g \qquad (4)$$

where $\rho_w$ is water density.

Changes in water contents, determined from neutron-moisture probe measurements, were observed to a depth of about a meter following a few large precipitation events (>25 mm) between December 1984 and March 1988. Measurements were made every 25 cm between the depths of 0.25 and 10 m and every 50 cm below a depth of 10 m. Gravimetric water contents at a depth of 1 m ranged from about 2 to 4 percent. More commonly, changes in water contents were observed only within the upper 80 cm of the sediments. Gravimetric water contents within 10 cm of land surface ranged from near zero in the summer to 21 percent (0.21 g/g) following precipitation events and were determined by periodically collecting samples and analyzing water content in the laboratory.

Water contents below a depth of 1 m showed little variation over time. Gravimetric water contents generally ranged from 4 to 8 percent; with lower water contents in clean sand or gravel layers and higher water contents in layers containing silt or clay. For example, low gravimetric water contents (below 3 percent) were observed in neutron-probe access tube N2, close to the gravel layers at depths of 1.5 and 9 m (fig. 13). Higher gravimetric water contents (about 8 percent or more) were observed at depths of 7, 14, and 25 m. These higher water contents are associated with sediments that contain greater amounts of silt and clay. Small variations in lithology between neutron-probe access tubes are reflected in slightly different water-content profiles in each tube (fig. 13). The lack of change in water contents over time below a depth of 1 m indicates little change in rates of water movement, at least within the sensitivity range of the neutron-moisture probe measurements used to estimate water contents.

## Hydraulic Properties

Saturated hydraulic conductivities of core samples were estimated using two methods. Deeper, more difficult to obtain core samples (depth greater than 75 cm) were analyzed using air-permeability tests to determine intrinsic permeability. These values were converted to saturated hydraulic conductivities using methods described by the American Petroleum Institute (1952). Air permeabilities were determined on the deeper core samples because the use of water would have eliminated the core samples from additional testing. Where samples were easily obtained, usually at the surface, a constant-head permeameter was used to determine the saturated hydraulic conductivity. Duplicate surface samples were used for later water retention tests. For both methods, hydraulic conductivity was measured in triplicate for each sample and the results averaged.

Saturated hydraulic conductivities of the sediments ranged from 1 cm/d to 48 cm/d (table 3). These values are less than the values expected for a clean sand or gravel, but are not unreasonable because most sediments are poorly sorted with at least 10 percent fine sand and silt. In addition, measurements were generally made on the finer grained alluvial sediments because the sampling method did not provide for whole samples of the cobble and gravel layers. Hydraulic conductivity values for the current study are generally an order of magnitude less than values reported by Nichols (1987, p. 47). Some of the differences between estimates may be due to different types of samples. Estimates reported by Nichols were determined using reconstituted samples that had been sieved to remove the gravels. Sediments composed of gravels within a fine-grained matrix generally have a lower hydraulic conductivity than the matrix itself because gravels reduce the cross-sectional area available for flow (Bouwer and Rice, 1984).

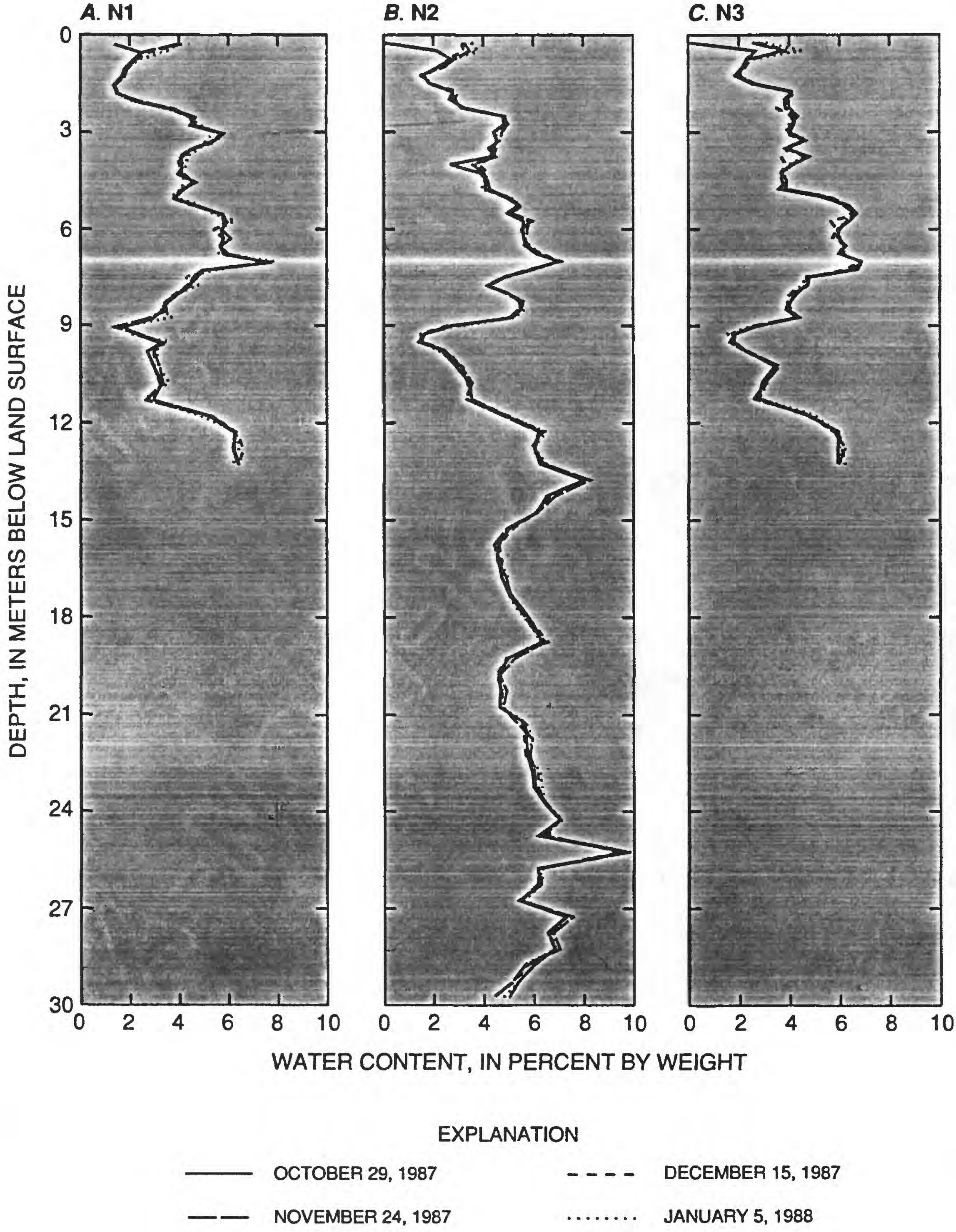



FIGURE 13.--Changes in water content on four dates between October 29, 1987, and January 5, 1988, from three neutron-probe access tubes. Location of neutron-probe access tubes. is shown in figure 2.

TABLE 3.--*Laboratory measurement of saturated hydraulic conductivity and calculated unsaturated hydraulic conductivity for selected samples*

[Abbreviations: cm/d, centimeters per day; $cm^3/cm^3$, cubic centimeters per cubic centimeter; m, meters; MPa, megapascals]

| | Hydraulic conductivity[1] | | | Values used to calculate unsaturated hydraulic conductivity[2] | | | | |
|---|---|---|---|---|---|---|---|---|
| Sample depth (m) | $K_s$ (cm/d) | $K_\ell$ (cm/d) | $K_f$ (cm/d) | $\theta_s$ ($cm^3/cm^3$) | $\theta_{v\ell}$ ($cm^3/cm^3$) | $\theta_{vf}$ ($cm^3/cm^3$) | $\psi_{wf}$ (MPa) | *b* |
| 0.3 | 2.3 | $2 \times 10^{-6}$ | $9 \times 10^{-20}$ | 0.472 | 0.164 | 0.016 | <-6.8 | 5.15 |
| .6 | 14 | $2 \times 10^{-6}$ | $8 \times 10^{-11}$ | .402 | .057 | .016 | <-6.8 | 2.55 |
| 1.5 | 37 | $6 \times 10^{-5}$ | $5 \times 10^{-9}$ | .352 | .084 | .030 | -4.0 | 3.11 |
| 2.6 | 46 | $2 \times 10^{-5}$ | $3 \times 10^{-8}$ | .382 | .132 | .083 | -4.5 | 5.48 |
| 4.1 | 1.1 | $7 \times 10^{-8}$ | $6 \times 10^{-11}$ | .393 | .115 | .068 | -4.6 | 5.23 |
| 5.8 | 48 | $3 \times 10^{-4}$ | $1 \times 10^{-5}$ | .353 | .137 | .104 | -4.0 | 4.78 |
| 7.3 | 20 | $4 \times 10^{-7}$ | $3 \times 10^{-7}$ | .381 | .107 | .106 | -4.4 | 5.48 |

[1] Symbols: $K_s$, saturated hydraulic conductivity; $K\ell$, unsaturated hydraulic conductivity at water potential ($\psi_w$) of -1.5 MPa from moisture-retention data; $K_f$, unsaturated hydraulic conductivity at field water potential ($\psi_{wf}$) and field water content.

[2] Saturated volumetric water contents ($\theta_s$) and volumetric water contents ($\theta_{v\ell}$) at water potential of -1.5 MPa determined from laboratory data. Field volumetric water contents ($\theta_{vf}$) are estimated from neutron-moisture probe measurements in access tube N2, June 11, 1987. Field water potentials ($\psi_{wf}$) are averages from thermocouple psychrometers in monitoring shaft and instrument borehole.

Unsaturated hydraulic conductivities were determined following the method described by Campbell (1974) and extended for use on stony desert sediments by Mehuys and others (1975). The unsaturated hydraulic conductivity at a given water potential, $K(\psi_w)$ in centimeters per day, for a desorbing sediment can be written as:

$$K(\psi_w) = K_s \left(\frac{\theta_v}{\theta_s}\right)^{2b+3} \tag{5}$$

where $K_s$ = saturated hydraulic conductivity, in centimeters per day;
$\theta_v$ = volumetric water content at a given water potential, dimensionless;
$\theta_s$ = saturated volumetric water content (porosity), dimensionless; and
$2b+3$ = an empirical constant related to texture.

The saturated hydraulic conductivity and saturated water content (porosity) were determined from laboratory analyses of core samples. Volumetric water content was determined either from laboratory moisture retention experiments, for water potentials from 0 to 1.5 MPa, or from neutron moisture-probe measurements at the field site. In the latter case, gravimetric water content was converted to volumetric water content using eq. 4 and laboratory estimates of bulk density.

The empirical constant ($2b+3$) was determined from laboratory tests for soil-moisture retention. The pressure plate extractor method (Klute, 1986, p. 635-662) was used on core samples to relate changes in volumetric water content to water potential. Incremental pressures of 0.1, 0.3, 0.5, 1.0, and 1.5 MPa were applied and maintained on 3-cm high sections of core samples for 24 to 48 hours, depending on when flow of water from the extractors ceased. Additional pressures of 0.01, 0.03, and 0.07 MPa were applied and maintained on samples of the surficial deposits. The core samples were weighed prior to applying additional pressure. A moist silica slurry was used to ensure good hydraulic contact between the core sample and the pressure plate. Volumetric water content was determined at each pressure increment using a bulk density determined at the end of the experiment. Results of moisture retention measurements indicate that the sediments lose almost 70 percent of their moisture at -0.1 MPa. Thereafter, moisture release is controlled by the fine-grained silt and clay fraction.

The parameter $b$ was determined as the slope of a least-square regression of the log $\left(\frac{\theta_v}{\theta_s}\right)$ against log $(\psi_w)$ and is based on the following relation:

$$\frac{\psi_w}{\psi_\varepsilon}=\left(\frac{\theta_v}{\theta_s}\right)^{-b} \qquad (6)$$

where $(\psi_\varepsilon)$ is the air-entry potential of the sample. The air-entry potential was assumed to be the intercept in the least-square regression. Coefficient of determination (r-squared) values for individual samples ranged from 73 to 91 percent.

Unsaturated hydraulic conductivities estimated from laboratory and field measurements range from $3x10^{-4}$ to $9x10^{-20}$ cm/d. At -1.5 MPa estimates of unsaturated hydraulic conductivity from moisture-retention data of seven samples range from $10^{-4}$ to $10^{-8}$ cm/d (table 3), five to eight orders of magnitude less than the estimates of saturated hydraulic conductivity. These values are slightly greater than the three estimates reported at -1.5 MPa (range $10^{-6}$ and $10^{-12}$ cm/d) from a previous study near the LLRW disposal facility (Nichols, 1987, p. 47). Water loss in the sediments was minimal at pressures below -1.5 MPa. Thus, the unsaturated hydraulic conductivity of the samples determined by extrapolating the moisture-retention data to -4.5 MPa were within one to four orders of magnitude of the values estimated at -1.5 MPa (table 3). The only exception was a sample of the fine-grained sandy silt at a depth of 0.3 m below land surface. The unsaturated hydraulic conductivity of this sample decreased from $10^{-6}$ to $10^{-20}$ cm/d between water potentials of -1.5 and -4.5 MPa. This estimated decrease was caused by the ability of the sample to retain more water at lower water potentials relative to the other samples.

Distribution of unsaturated hydraulic conductivities for sediments at the site is incomplete because few core samples were obtained from the cobbly layers or below a depth of 8 m. However, some inferences on constraints to vertical flow can be made. First, in such dry sediments, the clean gravel layers between the depths of about 0.75 and 2.2 m and 8.5 and 9 m will act as capillary barriers and impede vertical flow. And second, the surficial deposits of sandy silt limit infiltration during storms because of their low saturated hydraulic conductivity. Storms in excess of 10 mm/hr probably produce overland flow, as has been observed on two occasions at the study site. The most likely scenario for deep percolation at the study site would be during an extended period of light rain under high antecedent moisture conditions. The effective vertical unsaturated hydraulic conductivity under these conditions is difficult to determine.

## Concentration of Salts

Sodium chloride (NaCl) can be used as an indicator of ground-water recharge rates in semi-arid to arid regions (Allison and Hughes, 1983). In addition, the accumulation of salts in pore water can have a significant effect on water potentials. Concentrations of salts in the sediments from the study site were measured by adding a known amount of water to a known weight of oven-dried sediment. Relative concentrations of salts were then determined by measuring the specific conductance of the solution and correcting for the weight of sample and water added.

Analysis of the alluvium for salts using specific conductance showed a zone of elevated concentrations between the depths of 2 and 8 m below land surface (fig. 14). Maximum concentrations were measured between the depths of 2 and 4 m. A similar trend was observed for chloride (Fouty, 1989) from analyses of the same samples, although chloride concentrations did not decrease as much as specific conductance between the depths of 6 and 8 m. Chemical analysis of extracts from four samples indicate that 70 to 85 percent of the cations were accounted for by sodium, and 55 to 75 percent of the anions were accounted for by chloride. The next most abundant ions were calcium and sulfate, accounting for 5 to 22 percent of the cations and 9 to 35 percent of the anions, respectively. Other ions, accounting for less than 10 percent of the total, were potassium, magnesium, nitrate, nitrite, and fluoride.

The concentration of NaCl between the depths of 2 and 8 m may indicate where water ceased to move through the pores as a liquid. Other possible explanations for such a distribution of NaCl in the sediments include changing rates of infiltration or salt input (Stone, 1984). However, if water had moved deeper as a liquid it would have carried NaCl with it and produced a uniform distribution with depth (Allison and Hughes, 1983). The low concentrations of NaCl in the upper meter of sediments implies that some precipitation has penetrated deeper than 1 m. Assuming vertical flow, the concentration of NaCl between the depths of 2 and 8 m indicates that percolating water has reached that depth and either evapotranspired or continued to move as vapor, which carries no salts. The rooting depth of the creosote bush is about 2 m, which is only slightly less than that of the maximum concentration of NaCl, suggesting that plant transpiration may be important in removing water from the sediments. In addition, the cobble layer at a depth of about 8 m may act as capillary barrier to downward movement of water. Low concentrations of NaCl below a depth of 9 m either reflect the concentrations of ancient percolating waters or are background levels from the depositional environment (Fouty, 1989, ch. 3, p. 13).

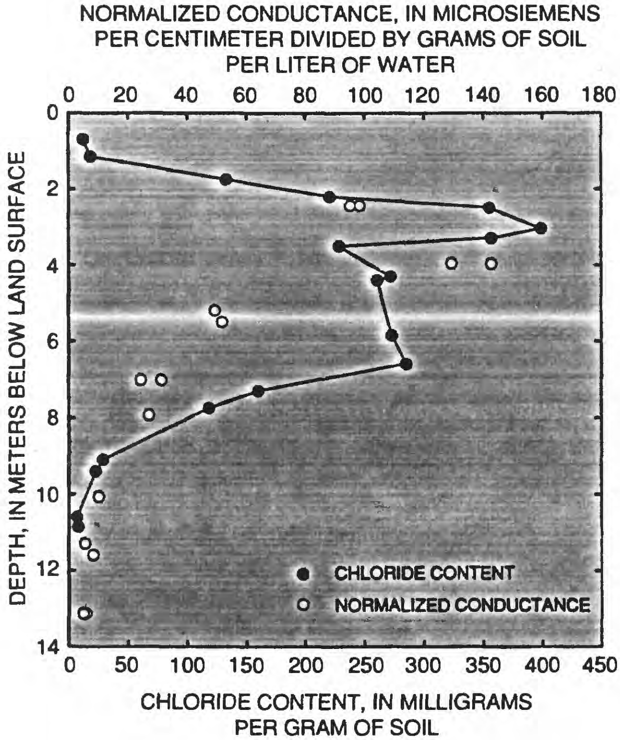



FIGURE 14.--Distribution of normalized conductance values and chloride concentrations from samples collected at instrument borehole between depths of 0.5 and 13 meters. Conductance was normalized by dividing measured conductance by weight of soil used, in grams, and volume of water added, in liters.

The effect of salts on water potentials in the unsaturated zone was determined by estimating osmotic potentials in the subsurface. Osmotic potentials were estimated from NaCl concentration, water content, and temperature of the sediments. Other salts were not considered because NaCl usually accounts for more than 85 percent of the total salts. Maximum NaCl concentrations of pore water, in excess of 6 g/L, were found at a depth of 2.3 m, and concentrations in excess of 1 g/L were present between the depths of 1.5 and 7.4 m. For calculating osmotic potentials, it was assumed that NaCl concentrations were totally dissolved in the pore water. Calculated osmotic potentials were -0.2 MPa at a depth of 1.8 m, decreasing to a minimum of -0.5 MPa at a depth of 2.3 m, and gradually increasing to -0.15 MPa at a depth of 7.5 m. From the land surface to a depth of 1.5 m, and below 8 m, the calculated osmotic potentials were considered insignificant, as values ranged from -0.1 to 0 MPa.

## Thermal Diffusivity

Thermal diffusivity ($D_T$) is an indication of the ability of the sediments to transmit heat and is used when estimating water movement in response to thermal gradients. Thermal diffusivity is the ratio of thermal conductivity to volumetric specific heat, which changes with water content and composition of the sediments (Campbell, 1977, p. 16). Higher values of thermal diffusivity for a sediment indicate a greater ability to transmit heat. In general, coarse-grained sediments have higher values of thermal diffusivity than fine-grained sediments for a given volumetric water content.

Temperatures measured from TCP's were used in the calculation of thermal diffusivity. Seasonal temperature fluctuations were observed at most of the thermocouple psychrometers and followed a pattern similar to the seasonal trend in air temperature. Mean daily air temperature (daily averages of values recorded hourly) at the site ranged from a low of about -2°C in January and December 1987 to a high of about 34°C in July and August 1987 (fig. 15A). (See figure 2 for location of the weather station in relation to the monitoring shaft and instrument borehole.) Mean daily temperature at a depth of 1.2 m below land surface (OD1) ranged from a low of about 13°C in January 1988 to a high of about 28°C in September 1987, whereas the mean daily temperature at a depth of 6 m (L4) ranged from a low of about 20°C in May and June 1987 to a high of about 22°C in December 1987 (fig. 15B). Mean daily subsurface temperatures are averages of values recorded every 4 hours. Below a depth of 9 m, temperatures were nearly constant (fig. 15C). Maximum and minimum temperatures at depth were increasingly lagged in time from the maximum and minimum air temperatures. Similar trends in subsurface temperature were observed at a site in eastern Washington (Isaacson and others, 1974). By comparing the time lag of the maximum temperatures at different depths, the temperature front was estimated to move downward at a rate of about 5 cm/d.

Thermal diffusivity of the sediments was calculated on the basis of measured surface and subsurface temperatures and observed seasonal changes in temperature with depth using an equation presented by Johnson and Lowery (1985, p. 1548). The time rate of change in subsurface temperature is:

$$\partial T/\partial t = D_T(\partial^2 T/\partial z^2) \qquad (7)$$

where $T$ = temperature, in degrees Celsius;
$t$ = time, in seconds; and
$z$ = depth, in meters.

This equation can be simplified assuming the sediments are homogeneous and infinitely deep, and that the temperature at the surface is (Campbell, 1977, p. 15):

$$T(z=0,t) = \overline{T} + A(o)\,[\sin(\omega t)] \qquad (8)$$

where $\overline{T}$ = average surface temperature, in degrees Celsius;
$A(o)$ = amplitude of the surface temperature fluctuation [1/2(maximum temperature - minimum temperature)], in degrees Celsius;
$\omega$ = angular frequency of the oscillation ($2\pi/P$), in per second; and
$P$ = period of temperature cycle, in seconds.

The sine is in radians. If $t = 0$, the start of the calculation period, surface temperature is equal to $\overline{T}$ and rising. From these assumptions, temperatures at any depth and time can be calculated from Campbell (1977, p. 16):

$$T(z,t) = \overline{T} + A(0)\,e^{-(z/z_D)}\sin(\omega t - z/z_D) \qquad (9)$$

where $z_D$ is the damping depth. The damping depth is related to thermal diffusivity by:

$$z_D = (2D_T/\omega)^{1/2} \qquad (10)$$

Solving for $D_T$ yields:

$$D_T = \frac{(z_D)^2\omega}{2} \qquad (11)$$

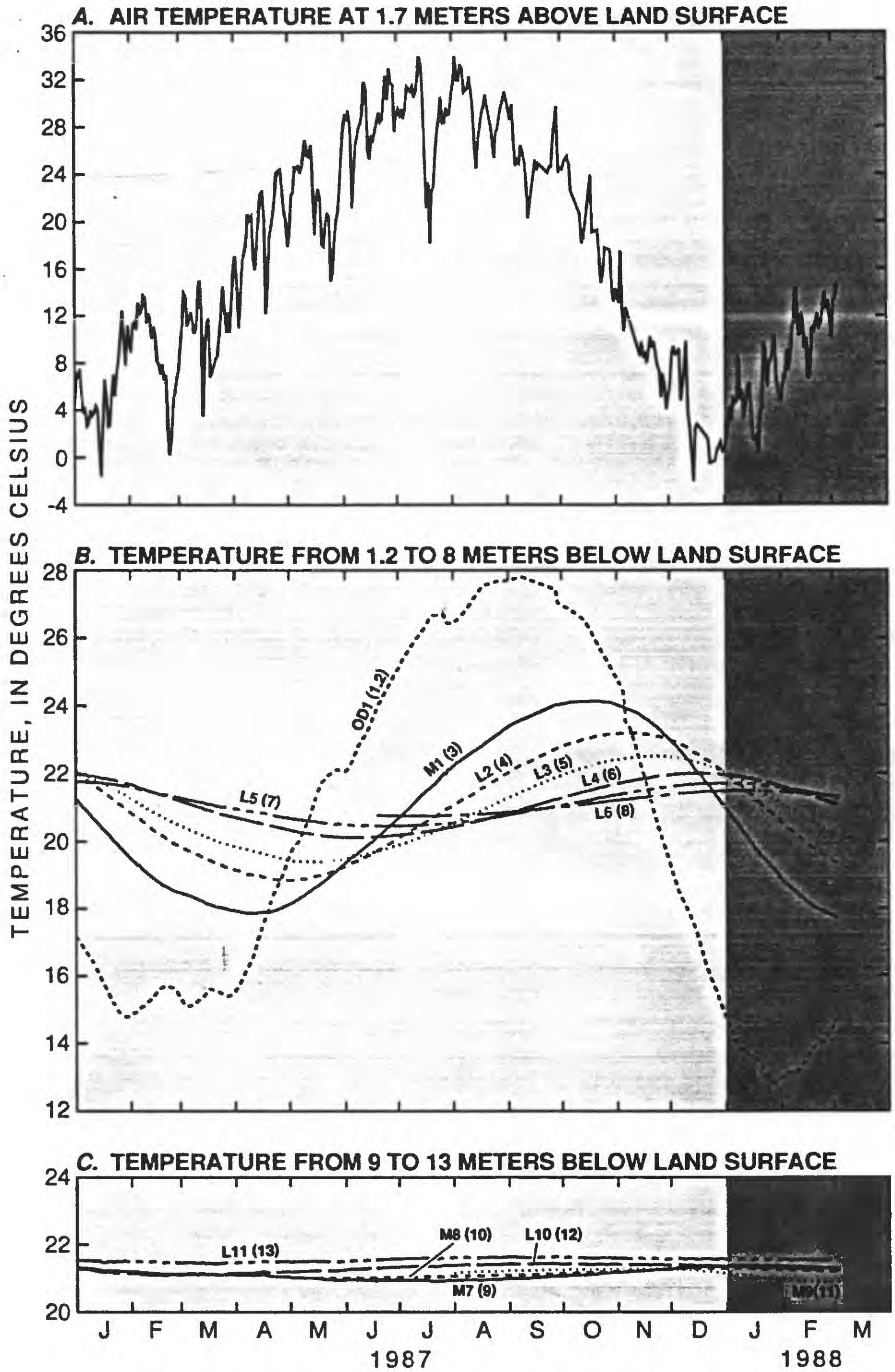



FIGURE 15.--Temperature trends observed at study site from January 1987 to March 1988. (*A*) Air temperature at 1.7 meters above land surface. (*B*) Temperature at depths from 1.2 to 8 meters below land surface. (*C*) Temperature at depths from 9 to 13 meters below land surface. Number in parentheses indicates depth of thermocouple psychrometers in meters below land surface. Location of thermocouple psychrometers is shown in figures 2 and 9.

Johnson and Lowery (1985, p. 1549) noted that the damping depth is related to the amplitude of the temperature fluctuations at depth as follows:

$$1/z_D = -d[\ln A(z)]/dz \qquad (12)$$

where $A(z)$ is the observed temperature fluctuation at depth $z$. A linear regression of the natural log of $A(z)$ onto $z$, where the slope from the regression is equal to $1/z_D$, was used to compute $z_D$.

A similar analysis was done using temperature data obtained from TCP's installed at the shaft during 1986 and 1987 and from a near-surface TCP (depth of 0.2 m below land surface) monitored during 1987. Data for 1987 are shown in table 4. Standard deviations for temperatures measured to depths of 9 m were used to correlate annual changes in amplitude of temperature fluctuations with depth. The amplitude of temperature fluctuations at a given depth is approximately equal to three times the standard deviation, assuming temperature values are normally distributed about a mean (Spiegel, 1961, p. 123). The relation between the natural log of the annual temperature fluctuation and depth is shown in figure 16. Linear regression of the natural log of three times the standard deviation of annual temperature fluctuations to depth resulted in the following relation:

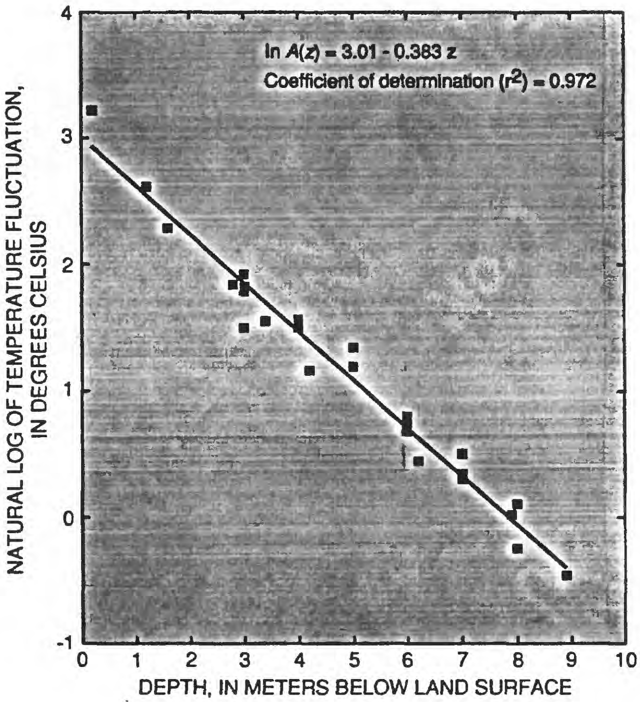



FIGURE 16.--Correlation of natural log of annual temperature fluctuation, $A(z)$, to depth below land surface, $z$. Annual temperature fluctuation is estimated as three times standard deviation. Temperature fluctuations from depths of 0.2 to 9 meters are used in correlation. Values of standard deviation of temperature between depths of 1.6 and 9 meters for 1987 are listed in table 4. Additional values for 1986 from thermocouple psychrometers in monitoring shaft between depths of 3 and 9 meters are included in correlation. Observed temperature fluctuation at a depth of 0.2 meters is from a near-surface thermocouple psychrometer monitored during 1987.

$$\ln A(z) = 3.01 - 0.383z \qquad (13)$$

The coefficient of determination (r-squared) for the regression is 0.97, indicating that the regression reasonably approximates the observed changes in annual temperature fluctuations with depth. The damping depth was estimated from the slope in eq. 13 as 1/0.383 or about 2.6 m. Substituting this value and a value for $\omega$ of $2 \times 10^{-7}$/s (the angular frequency for 1 yr) into eq. 11 yields a thermal diffusivity of $6.8 \times 10^{-7}$ m$^2$/s or 0.68 mm$^2$/s. This value may represent an average thermal diffusivity at the study site as the sediments are not homogeneous and uniform with depth. However, substitution of the calculated value of $z_D$ into eq. 9 yielded results that predicted maxima and minima within ±10 days at depths of 3, 5, and 7 m. Calculated temperatures agreed within 3°C at a depth of 3 m and within 1°C at depths of 5 and 7 m. Thermal diffusivity of individual units in the sediments may be different from the estimated value but nonetheless, the value indicates that the sediments at the study site are good thermal conductors.

TABLE 4.--*Mean and standard deviation of water potential and temperature for 1987 from thermocouple psychrometer measurements in monitoring shaft and instrument borehole*

[Means and standard deviations are based on daily averaged values of measurements taken every 4 hours]

| Thermocouple psychrometer identification[1] | Depth below land surface (meters) | Water potential (megapascals) Mean | Water potential (megapascals) Standard deviation | Temperature (degrees Celsius) Mean | Temperature (degrees Celsius) Standard deviation |
|---|---|---|---|---|---|
| | | Monitoring Shaft | | | |
| L1 [2] | 3.0 | -3.7 | -0.73 | 20.9 | 2.10 |
| M1 [2] | 3.0 | -4.2 | -.46 | 20.8 | 2.07 |
| L2 | 4.0 | -4.6 | -.20 | 21.1 | 1.52 |
| L3 | 5.0 | -4.3 | -.66 | 21.0 | 1.10 |
| L4 | 6.0 | -4.2 | -.94 | 21.1 | .70 |
| L5 | 7.0 | -4.3 | -.17 | 21.1 | .48 |
| M5 | 7.0 | -4.5 | -.12 | 21.2 | .46 |
| L6 [3] | 8.0 | -3.9 | -1.09 | 21.1 | .26 |
| M7 | 9.0 | -4.5 | -.09 | 21.1 | .14 |
| M8 [4] | 10.0 | -4.6 | -.18 | 21.1 | .11 |
| M9 [4] | 11.0 | -4.0 | -.18 | 21.2 | .08 |
| L10 | 12.0 | -3.7 | -.02 | 21.3 | .12 |
| L11 | 13.0 | -3.2 | -.07 | 21.5 | .06 |
| M11 | 13.0 | -3.1 | -.18 | 21.5 | .05 |
| | | Instrument Borehole | | | |
| OD1 | 1.2 | -4.0 | -1.39 | 20.4 | 4.54 |
| OD2 | 1.6 | -4.0 | -1.28 | 20.7 | 3.28 |
| OD3 | 2.8 | -4.5 | -.13 | 21.1 | 2.11 |
| OD4 | 3.4 | -4.5 | -.10 | 21.3 | 1.58 |
| OD5 | 4.2 | -4.6 | -.16 | 21.4 | 1.07 |
| OD6 | 6.2 | -4.0 | -.17 | 21.5 | .52 |
| OD7 | 7.9 | -4.4 | -.61 | 21.6 | .34 |
| OD8 | 8.9 | -4.5 | -.19 | 21.6 | .21 |
| OD10 | 11.1 | -4.0 | -.66 | 21.7 | .14 |
| OD11 | 11.1 | -4.0 | -.12 | 21.7 | .21 |
| OD12 | 11.4 | -3.5 | -.94 | 21.6 | .14 |
| OD13 | 11.9 | -3.5 | -.49 | 21.8 | .32 |

[1] Thermocouple psychrometers are identified in figure 9.

[2] Mean and standard deviation of water potentials are based on measurements from January 1, 1986, to December 31, 1986.

[3] Mean and standard deviation of water potentials are based on measurements from January 1, 1987, to June 30, 1987.

[4] Mean and standard deviation of temperature and water potential are based on measurements from June 1, 1987, to December 31, 1987.

# WATER MOVEMENT

Because of the extremely low water potentials at the site, water moves through the shallow unsaturated alluvium as both liquid and vapor. Direction of water movement in the unsaturated sediments varies depending on seasonal trends in water potentials and temperatures. In the following sections, water potentials in the upper 13 m of unsaturated alluvium are discussed and estimates of liquid-water and water-vapor fluxes are presented.

## Water Potentials

Interpretation of the water-potential data was difficult without being sure which observed variations were natural and which were due to instrument error or installation problems. By comparing data from the monitoring shaft with data from the adjacent instrument borehole, some TCP error was eliminated and confidence in the data was increased. For example, if similar trends were observed in TCP's in both installations at similar depths then the trends were probably caused by natural processes. Conversely, if data from one TCP at a particular depth differed greatly from the data from other TCP's at the same depth, the difference probably was caused by a faulty TCP or improper installation.

### *Reliability of Measurements*

Water-potential data from duplicate TCP's installed at the same depth within the monitoring shaft were compared to assess the reproducibility of the data. Water-potential measurements for each TCP were taken every 4 hours and averaged daily. Duplicate TCP's were installed at depths of 3, 7, and 13 m within the shaft. The duplicate TCP data at a depth of 7 m were comparable (fig. 17). Water potentials at a depth of 13 m remained nearly constant at about -3.3 MPa (fig. 18), whereas water potentials at a depth of 7 m varied seasonally from -4.0 to -4.7 MPa (fig. 17). Water potentials from a depth of 3 m also showed seasonal variation; however, the readings on the three TCP's were rarely in agreement. Values ranged from -3.5 to -5.5 MPa, with individual TCP's varying more than 2.0 MPa within a month. It is possible that the TCP's installed at the depth of 3 m in the monitoring shaft have installation problems because these were the first access holes drilled in the monitoring shaft. A different type of glue was used to seal the end plugs in place at this depth. This glue may not provide an adequate seal between the TCP and the shaft environment, thus affecting the measurements of the TCP. Comparison of temperatures measured within the shaft to temperatures measured at the TCP's in the soil indicate that temperature fluctuations in the shaft were not affecting the TCP's. This conclusion was generally supported by similar temperature readings at comparable depths from TCP's installed in the borehole. The possibility that vapor may be transferred between the shaft and the sediments at 3 m cannot, however, be ruled out.

Comparison of water potentials from TCP's buried at similar depths in the borehole were generally in agreement (figs. 17 and 18). For example, water potentials from TCP's between the depths of 1.2 and 1.6 m were in general agreement, as were those from TCP's at depths of 2.8, 3.4, and 4.2 m. Water potentials from TCP's between the depths of 11.1 and 11.4 m, also had similar values and trends. TCP's buried above a depth of 8 m showed seasonal changes in water potential ranging from -3.5 to -5.5 MPa. However, during two short periods in mid-November 1987 and late January 1988, extreme but short-lived changes in water potentials were observed in TCP's at depths below 7 m in the instrument borehole (fig. 18). These changes followed storms of 17 mm and 10 mm on November 5, 1987, and January 17, 1988, respectively. The cause of these rapid changes in water potential is unknown. A possibility is that wetting of surface deposits increased air pressure in the subsurface and produced preferential air flow down the backfill material within the instrument borehole. The possibility of preferential air flow is supported by temperature increases of 1 to 2°C (fig. 19). This may explain, in part, the relatively high standard deviations of water potentials in the instrument borehole observed during 1987 (table 4). Reasons why the TCP's responded to these storms and not previous storms are unknown. Similarly, reasons why TCP's at shallower depths did not respond to the same storms are also not known. Perhaps the wet and dry bentonite used as a seal between TCP's to depths of about 5 m (fig. 9) isolated these TCP's from the preferential air flow. Analysis of the data is further complicated because some TCP's in the borehole may not be functioning properly even though they produce values. TCP's OD9 and OD11, installed at depths of 10.1 and 11.1 m, seem to have stopped functioning correctly about September 1987.

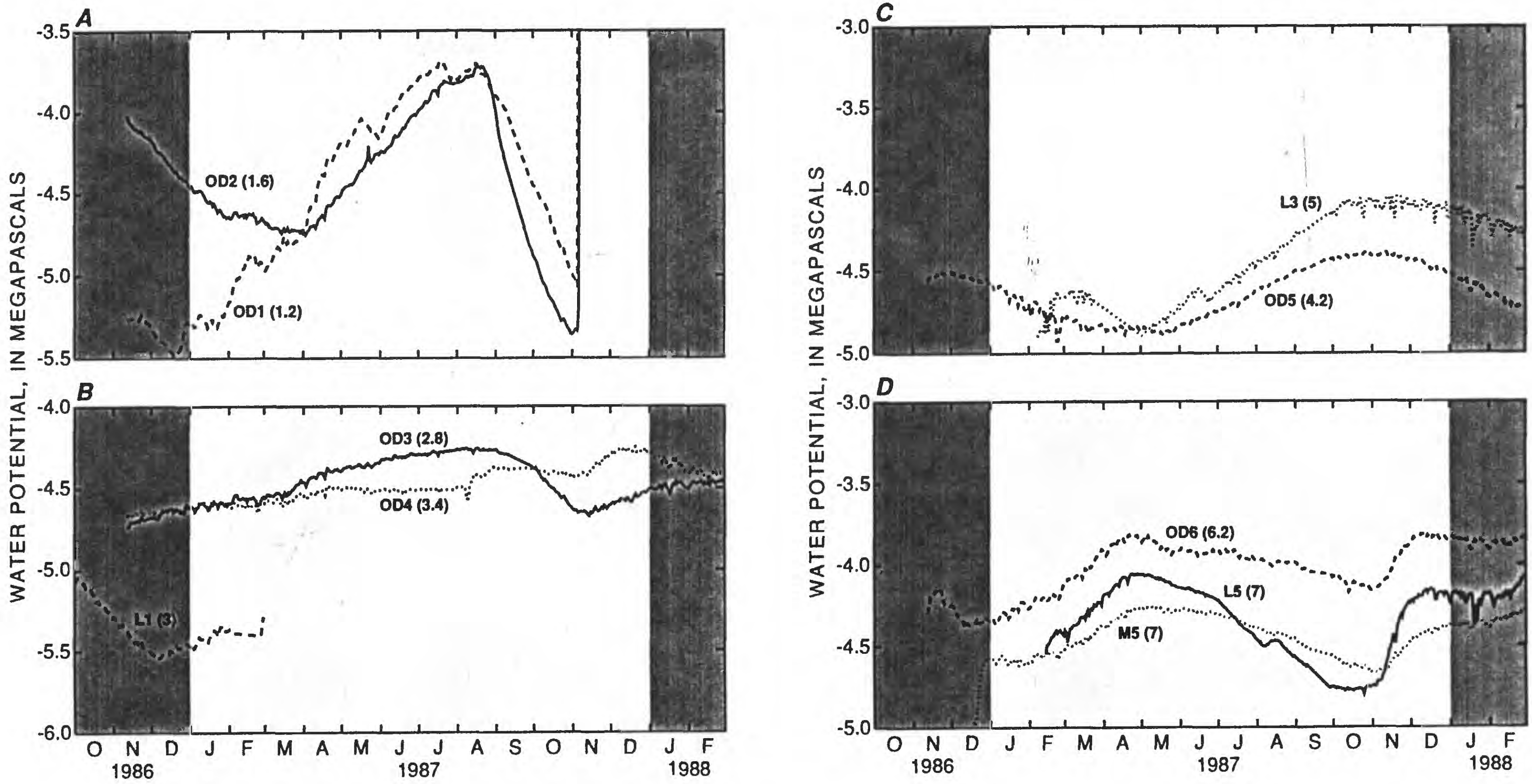



FIGURE 17.--Water potentials determined from thermocouple psychrometers installed in monitoring shaft and instrument borehole between depths of 1 and 7 meters during October 1986 through February 1988. (*A*) Thermocouple psychrometers OD1 and OD2; (*B*) thermocouple psychrometers OD3, OD4, and L1; (*C*) thermocouple psychrometers OD5 and L3; and (*D*) thermocouple psychrometers OD6, M5, and L5. Location of thermocouple psychrometers is shown in figures 2 and 9. Number in parentheses indicates depth of thermocouple psychrometer, in meters below land surface.

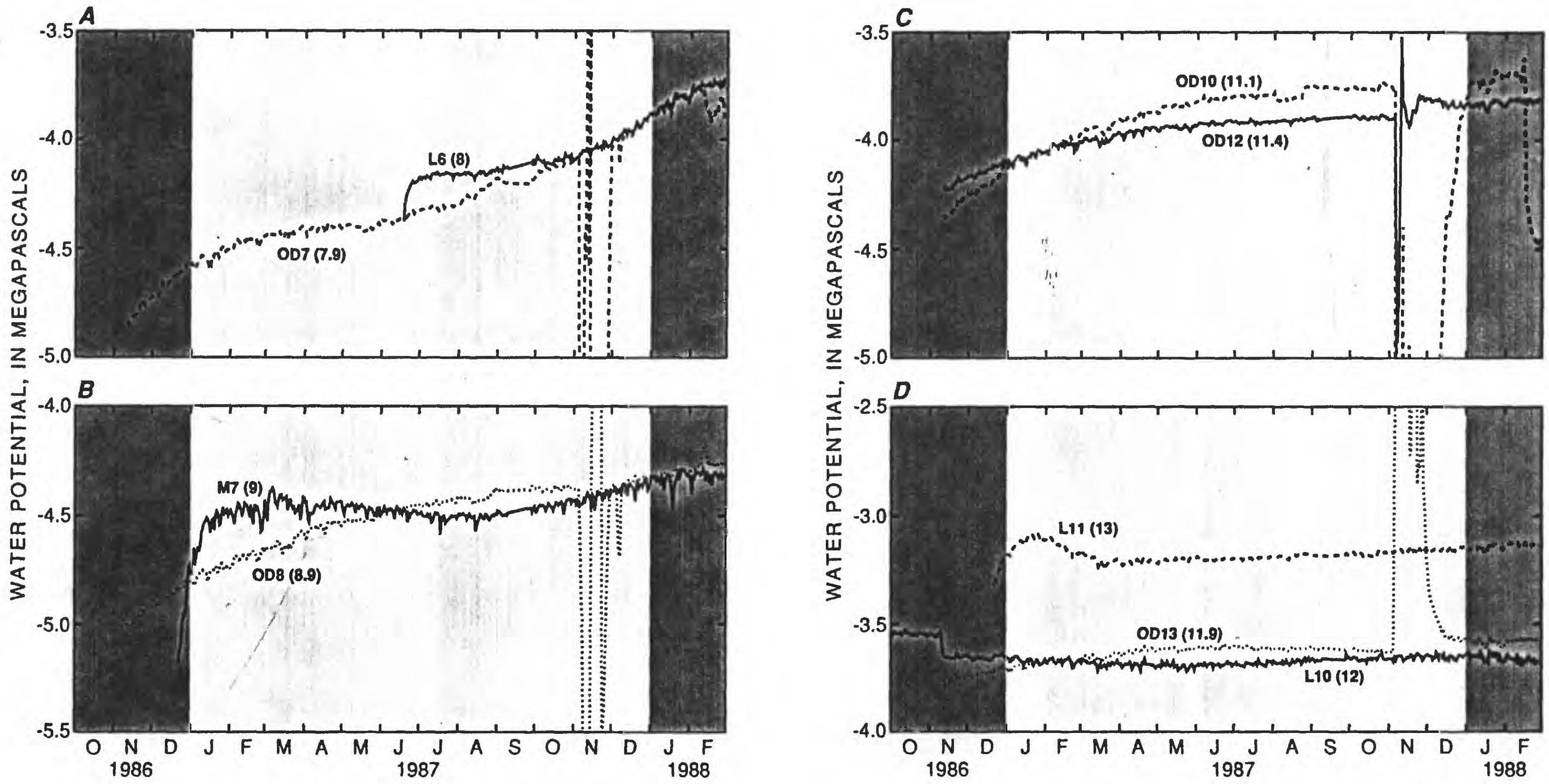



FIGURE 18.--Water potentials determined from thermocouple psychrometers installed in monitoring shaft and instrument borehole between depths of 8 and 13 meters during October 1986 through February 1988. (*A*) Thermocouple psychrometers OD7 and L6; (*B*) thermocouple psychrometers OD8, and M7; (*C*) thermocouple psychrometers OD10 and OD12; and (*D*) thermocouple psychrometers OD13, L10, and L11. Location of thermocouple psychrometers is shown in figures 2 and 9. Number in parentheses indicates depth of thermocouple psychrometer, in meters below land surface.

Comparison of water potentials from TCP's buried at similar depths in both the borehole and the shaft were generally in agreement. Water potentials for a given depth below 4 m determined from TCP's in both the shaft and borehole, except as previously noted, agreed to within 0.3 MPa of each other (figs. 17 and 18). For example, water potentials at depths of 4.2 m and 6.2 m in the borehole were comparable to water potentials at 5 m and 7 m in the shaft (fig. 17). Likewise, water potentials at depth of 8, 9, and 12 m in both installations were comparable (fig. 18). Water potentials measured above 4 m in both installations showed considerable variation over time (figs. 17 and 20); however, TCP readings from a depth of 3 m in the borehole (fig. 17) were not as variable (-4.2 to -4.7 MPa) as those from a depth of 3 m in the monitoring shaft (fig. 20) during the period October 1985 to March 1987 (-4.2 to -5.6 MPa). This difference may indicate that the TCP's installed at the depth of 3 m in the shaft are inadequately sealed or water added to the layers of bentonite in the borehole is being slowly released into the dry cuttings next to the TCP's. The other difference between the two installation procedures is that short-term water potential changes and concurrent changes in temperature observed in the borehole were not observed in the monitoring shaft. Although the method of backfilling the borehole may allow for preferential air movement, it may be possible, but is unlikely, that pressures and air temperatures at depth actually did increase, and that something about the TCP installation in the shaft attenuated these changes.

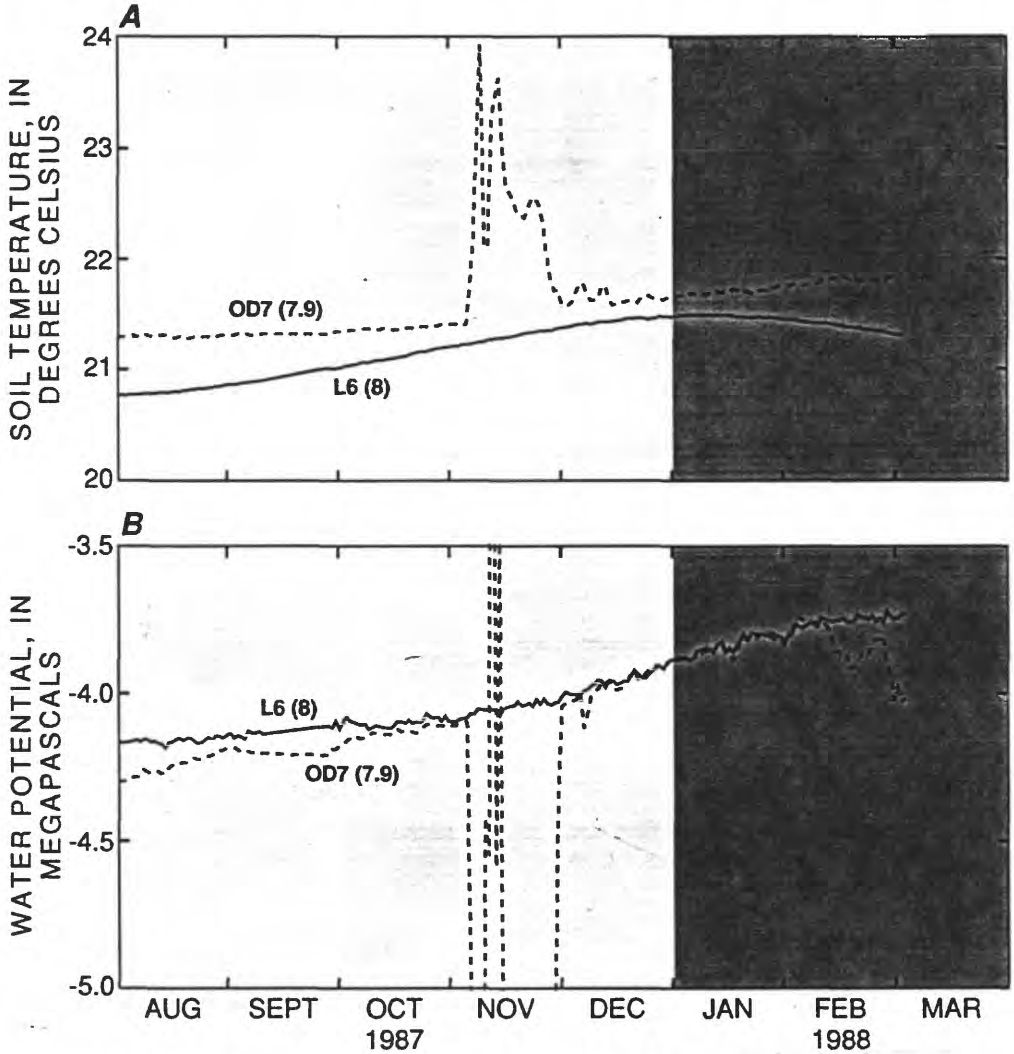



FIGURE 19.--Comparison of (*A*) temperatures and (*B*) water potentials measured from August 1987 to March 1988 for thermocouple psychrometers installed at a depth of about 8 meters below land surface in monitoring shaft and instrument borehole. Location of thermocouple psychrometers is shown in figures 2 and 9. Number in parentheses indicates depth of thermocouple psychrometer, in meters below land surface.

Water potentials determined from TCP's at depths of 4, 6, and 11 m in the shaft, and from TCP's at depths of 10 and 11 m in the borehole, were not used in the analysis of water movement because variations in the TCP readings appeared beyond what was considered reasonable. These instrument readings fluctuated widely or had not equilibrated as of September 1988.

Water potentials measured with TCP's installed in the sediments, either in the shaft or in the instrument borehole, did not agree with water potentials measured on core samples. Field measurements of water potential ranged from -3.1 to -5.8 MPa, whereas measurements on core samples ranged from -4.0 to below -8.0 MPa. Usually, water potentials measured on small volumes (1 $cm^3$) of a larger core sample were equal to or lower than those measured in the field, suggesting that water in the small core samples was lost to evaporation either during the drilling and coring process or during sample handling and analysis. The already dry sediments would need to lose only a small amount of water to lower the water potential significantly. Therefore, water potentials measured on the core samples were inadequate for determining actual gradients in the sediments.

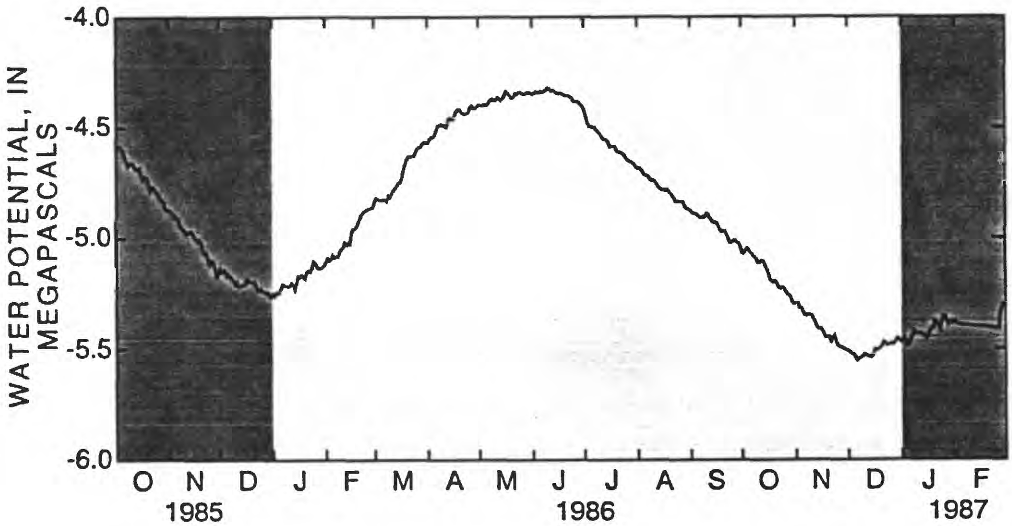



FIGURE 20.--Water potential determined from thermocouple psychrometer L1 installed at a depth of 3 meters below land surface in monitoring shaft from October 1985 to March 1987. Location of thermocouple psychrometer is shown in figure 9.

### *Seasonal Trends*

Water potentials determined from TCP's buried within 2 m of land surface varied seasonally and showed rapid responses to rainfall. Between November 1986 and March 1988, a period of little precipitation, water potentials in the upper meter of sediments were generally beyond the calibration range of the TCP's (less than about -7 MPa). For two TCP's installed in the instrument borehole at depths of 1.2 and 1.6 m below land surface, water potentials generally were highest (about -3.8 MPa) during July and August and were lowest (less than -5.0 MPa) between December and March (fig. 17A). This trend was not initially expected because evapotranspiration was anticipated to dry the sediments in the upper 2 m during the summer. However, the observed trend at depths of 1.2 and 1.6 m seems more related to changes in soil temperature than to evapotranspiration.

Changes in temperature produce changes in water content in the sediments, which affects the water potential in the sediments. Moisture in the sediments moves from warm areas to cooler areas. Thus, during the summer, moisture tends to move downward into the sediments from the surface because temperatures in the near-surface sediments are higher than at depth. In the winter, moisture tends to move upward because temperatures in the near-surface sediments are lower than at depth.

In the upper meter of sediments, water potentials increased rapidly to more than -0.5 MPa following precipitation. Water potentials between depths of 1 and 2 m also increased rapidly following precipitation as shown during November 1987 (fig. 17a). In contrast, neutron-moisture probe readings taken immediately after precipitation did not indicate a change in water content below a depth of 1 m. Apparently, small changes in water content, not detectable with the neutron-moisture probe, result in large changes in water potentials. Whatever the cause, for differences between the observed changes in depth between water potentials and water content measurements, the gravel layer at a depth between 0.75 and 2.2 m effectively retards water movement and acts as a capillary barrier. (Capillary barriers stop the flow of water through interconnected pore spaces, but not the movement of vapor.) This gravel layer causes the upper meter of sediments to act as a storage unit for the small amounts of precipitation that fall at the site. Moisture redistribution does occur, but a lack of a response to infiltration below a depth of 1 to 2 m indicates that most, if not all, of the stored water is lost by evapotranspiration.

Water potentials varied seasonally between the depths of 1 and 7 m (fig. 17) and seem related to changes in temperature. The seasonal variations were observed from TCP's in both the monitoring shaft and instrument borehole. The magnitude of seasonal variations generally decreased with depth, ranging from -4 to -5.5 MPa at a depth of 3 m and from -4.1 to -4.8 at a depth of 7 m. Seasonal maxima and minima did not occur at the same time at all depths. Both the range in water potentials and the seasonal variations were similar to observations at depths of 3 and 4 m in a nearby alluvial valley (Kearl, 1982).

Water potentials between the depths of about 8 and 11.5 m did not vary seasonally. Instead, water potentials increased slowly from November 1986 until March 1988 (fig. 18). Whether the observed increase in water potentials is natural or in response to air injected into the formation during drilling of the access holes is not known. Water potentials ranged from about -4.9 to -3.7 MPa in this zone. In contrast, water potentials at depths of 12 and 13 m remained relatively stable during the study period at -3.7 and -3.2 MPa, respectively (fig. 18D).

Water potentials below the depth of 13 m can be inferred from estimates of water content as determined from neutron-moisture readings in the 31-m deep access tube (N2) and from moisture-retention data determined from core samples. Below the depth of 13 m, estimates of gravimetric water content were generally between 5 and 6 percent by weight, except at depths of about 14, 25, and 28 m, where estimates were more than 7 percent by weight (fig. 9). Using these estimates and the relation of water content to water potential determined from moisture-retention data, water potentials between the depths of 14 and 31 m were estimated to range from -2.5 to -5.5 MPa. The higher water potentials (less negative) are associated with intervals of higher water content that correspond to fine-grained sediments. Water potentials below the zone of seasonal variations may, therefore, be a function of the presence or absence of fine-grained sediments.

## Estimates of Liquid-Water Flux

Both hydraulic conductivity and water-potential gradient are needed to determine an estimate of liquid-water fluxes in the unsaturated zone. A hydraulic gradient averaged over each month in 1987 was calculated at selected depths using mean daily water potentials and distances between selected TCP's (table 5). Uncertainty of the calculated gradients is approximately 20 cm/cm for estimates below a depth of 5 m. Above a depth of 5 m, uncertainty is greater, probably around 60 cm/cm, because water potentials varied considerably and values at similar depths did not always correspond.

The water-potential gradients (table 5) above a depth of 9 m vary seasonally whereas between the depths of 9 and 13 m, the gradient is always upward. Most of the calculated gradients above a depth of 5 m are within the estimated uncertainty and, therefore, were not used to calculate fluxes. The gradient varied in magnitude between the depths of 5 and 6 m, but was usually upward toward the surface. The strongest upward gradient, over -50 cm/cm, between these depths was estimated from February through July 1987. The gradient between the depths of 6 and 7 m was always downward and usually in excess of 40 cm/cm. Water-potential gradients between the depths of 7 and 8 m and 8 and 9 m were within the estimated accuracy of 20 cm/cm for the first 7 months of the year. From August to December, however, the gradients slowly increased from -20 to -50 cm/cm (upward) between the depths of 7 and 8 m and from 24 to 36 cm/cm (downward) between the depths of 8 and 9 m. Water-potential gradients between the depths of 9 and 11 m, 11 and 12 m, and 12 and 13 m were upward throughout the study period with gradients ranging from -16 to -54 cm/cm.

The normally upward gradient between the depths of 5 and 6 m and the downward gradient between the depths of 6 and 7 m suggests that during most of 1987, water was moving away from sediments at a depth of 6 m. A similar condition was observed from August to December at a depth of 8 m where the gradient between depths of 7 and 8 m was upward and the gradient between depths of 8 and 9 m was downward (table 5). Conversely, during August through December, the downward gradients between the depths of 8 and 9 m and an upward gradient between depths of 9 and 11 m suggests that water was moving toward the sediments at a depth of 9 m. Upward gradients below a depth of 9 m throughout 1987 suggests that water movement was consistently upward. General vertical movement of water below the depth of seasonal variation (8 m) may be related to water content of the sediments. The implied vertical movement of water away from the depth of 14 m corresponds to a zone of higher water content; whereas the implied movement of water toward the depth of 9 m corresponds to a zone of lower water content (fig. 21).

A range of hydraulic conductivities and hydraulic gradients were used to obtain estimates of liquid-water flux. The unsaturated hydraulic conductivity at the prevailing water potential and water content is variable and ranges from $1 \times 10^{-5}$ to $5 \times 10^{-9}$ cm/d (table 3).

Water-potential gradients in the upper 9 m range about 200 cm/cm to near 0 cm/cm. Thus, calculated fluxes in the upper 9 m range from $2 \times 10^{-3}$ to $1 \times 10^{-9}$ cm/d and are both upward (-) and downward (+) depending on the time of year. Water-potential gradients between the depths of 9 and 13 m did not vary considerably and were consistently upward during 1987 (table 5). The fluxes over this interval are estimated to range from $1 \times 10^{-2}$ to $2 \times 10^{-8}$ cm/d assuming the same range in hydraulic conductivity as used to estimate fluxes in the upper 9 m of sediments.

The flux estimates (particularly in the upper 9 m) may or may not be reasonable because of the transient nature of the water-potential gradients, because of the uncertainty in the unsaturated hydraulic conductivity, and because of possible errors associated with using eq. 1 in these dry sediments. Nonetheless, the analyses indicate that liquid-water fluxes above a depth of 9 m are small, of short duration, and of approximately equal value upward and downward at different times of the year. Under current climatic conditions, this implies water movement in the upper 9 m at the study site is in approximate equilibrium and water moves from one depth to another depending only on seasonal changes in water potential and temperature. These results, as well as the observed upward gradient below a depth of 9 m, indicate limited potential for transport of radionuclides by liquid-water flow.

TABLE 5.--*Mean monthly water-potential and vapor-density gradients during 1987 for selected depth intervals*

[Negative gradients indicate flux is upward; positive gradients indicate flux is downward. Gradients were calculated using monthly average of mean daily water potentials and temperatures from measurements taken every 4 hours. Uncertainty in water-potential gradients is estimated at ±20 centimeters per centimeter for depth intervals below 5 meters and ±60 centimeters per centimeter for depth intervals between 1.2 and 5 meters. Uncertainty in vapor-density gradients is estimated at about ±0.6 x $10^{-9}$ grams per cubic centimeter per centimeter]

| Depth interval[1] (meters) | Month | | | | | | | | | | | | Average for 1987 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | January | February | March | April | May | June | July | August | September | October | November | December | |
| | *Water-potential gradient, in centimeters per centimeter* | | | | | | | | | | | | |
| 1.2-1.6 | -191.36 | -84.48 | -30.93 | 51.50 | 61.30 | 46.93 | 34.95 | 2.59 | 68.31 | 112.66 | -- | -- | 13.77 |
| 1.6-2.8 | 8.11 | -5.96 | -16.32 | -18.44 | -.13 | 14.90 | 33.28 | 40.47 | -.68 | -54.12 | -- | -- | 44.65 |
| 3.0-4.0 | -8.68 | 45.30 | -17.53 | 16.08 | 71.26 | 99.03 | 192.07 | 155.42 | 214.31 | -- | -- | 267.88 | 149.71 |
| 4.0-5.0 | -190.37 | -40.60 | 4.79 | 69.01 | 12.05 | -7.29 | -90.45 | -52.23 | -78.75 | -125.45 | -99.55 | -68.34 | -55.60 |
| 5.0-6.2 | 70.99 | -50.03 | -58.38 | -108.37 | -78.63 | -65.45 | -52.36 | -34.06 | -18.49 | 1.06 | -5.64 | -23.99 | -35.28 |
| 6.2-7.0 | 34.85 | 44.43 | 53.90 | 58.42 | 50.09 | 46.94 | 51.43 | 56.03 | 60.47 | 63.43 | 71.78 | 74.14 | 55.49 |
| 7.0-7.9 | -3.02 | -9.14 | -2.66 | 10.20 | 14.21 | 8.03 | -2.79 | -20.39 | -34.73 | -54.38 | -- | -49.91 | -5.52 |
| 7.9-9.0 | 5.27 | 2.49 | 2.28 | 6.23 | 6.13 | 11.35 | 17.64 | 23.95 | 27.00 | 29.30 | -- | 36.31 | 7.35 |
| 9.0-11.1 | -18.47 | -16.53 | -16.08 | -18.02 | -18.45 | -19.69 | -20.59 | -20.96 | -20.59 | -19.53 | -16.30 | -18.89 | -18.68 |
| 11.1-12.0 | -42.27 | -35.45 | -31.37 | -27.05 | -24.73 | -23.31 | -23.47 | -23.27 | -23.54 | -23.36 | -30.04 | -17.61 | -27.12 |
| 12.0-13.0 | -53.68 | -53.07 | -48.22 | -47.88 | -49.27 | -48.97 | -48.87 | -49.27 | -48.63 | -48.49 | -49.52 | -50.01 | -49.65 |
| | *Vapor-density gradient, in grams per cubic centimeter per centimeter (multiply values by $10^{-9}$)* | | | | | | | | | | | | |
| 1.2-1.6 | -53.56 | -37.29 | -30.00 | 1.68 | 31.31 | 40.35 | 50.06 | 44.19 | 38.28 | 14.90 | -28.58 | -62.88 | 0.705 |
| 1.6-2.8 | -17.51 | -17.05 | -13.86 | -6.77 | 5.99 | 11.21 | 16.93 | 17.17 | 15.49 | 8.02 | -.39 | -14.70 | .377 |
| 3.0-4.0 | -7.34 | -10.89 | -10.93 | -8.59 | -1.94 | 5.56 | 5.58 | 12.41 | 14.19 | 15.83 | 9.54 | -.32 | 1.923 |
| 4.0-5.0 | -4.04 | -5.30 | -6.01 | -5.30 | -3.92 | -2.03 | 5.76 | 5.25 | 6.31 | 6.04 | 4.91 | 1.51 | .266 |
| 5.0-6.2 | .02 | -5.10 | -7.76 | -9.94 | -7.24 | -4.43 | -1.39 | 2.25 | 4.36 | 6.08 | 5.35 | 2.24 | -1.300 |
| 6.2-7.0 | .48 | -1.03 | -2.08 | -2.93 | -4.76 | -3.91 | -2.16 | .27 | 2.12 | 4.13 | 4.56 | 3.90 | -.116 |
| 7.0-7.9 | 2.86 | 1.88 | -.72 | -4.20 | -2.08 | -2.71 | -2.77 | -1.98 | -1.13 | .10 | 1.32 | 1.93 | -.624 |
| 7.9-9.0 | 3.39 | 1.14 | .96 | 2.40 | -1.24 | -1.48 | -1.29 | -.97 | -.50 | .41 | 1.13 | 1.61 | .463 |
| 9.0-11.1 | -3.67 | -4.07 | -5.09 | -2.39 | 1.44 | 1.05 | -.15 | -.88 | -.64 | -.33 | -.08 | .34 | -1.208 |
| 11.1-12.0 | -3.38 | -3.91 | -3.88 | -- | -- | -1.33 | 2.09 | -3.75 | -3.52 | -2.90 | -2.31 | -2.50 | -2.539 |
| 12.0-13.0 | -4.38 | -4.37 | -3.84 | -3.64 | -3.35 | -3.05 | -2.82 | -2.67 | -2.53 | -2.44 | -2.35 | -2.46 | -3.158 |

[1]Depth intervals between psychrometers referenced to land surface. Psychrometers used to estimate gradients include 0D1, 0D2, 0D3, L1, L2, L3, 0D6, M5, 0D7, M7, 0D10, L10, and L11.

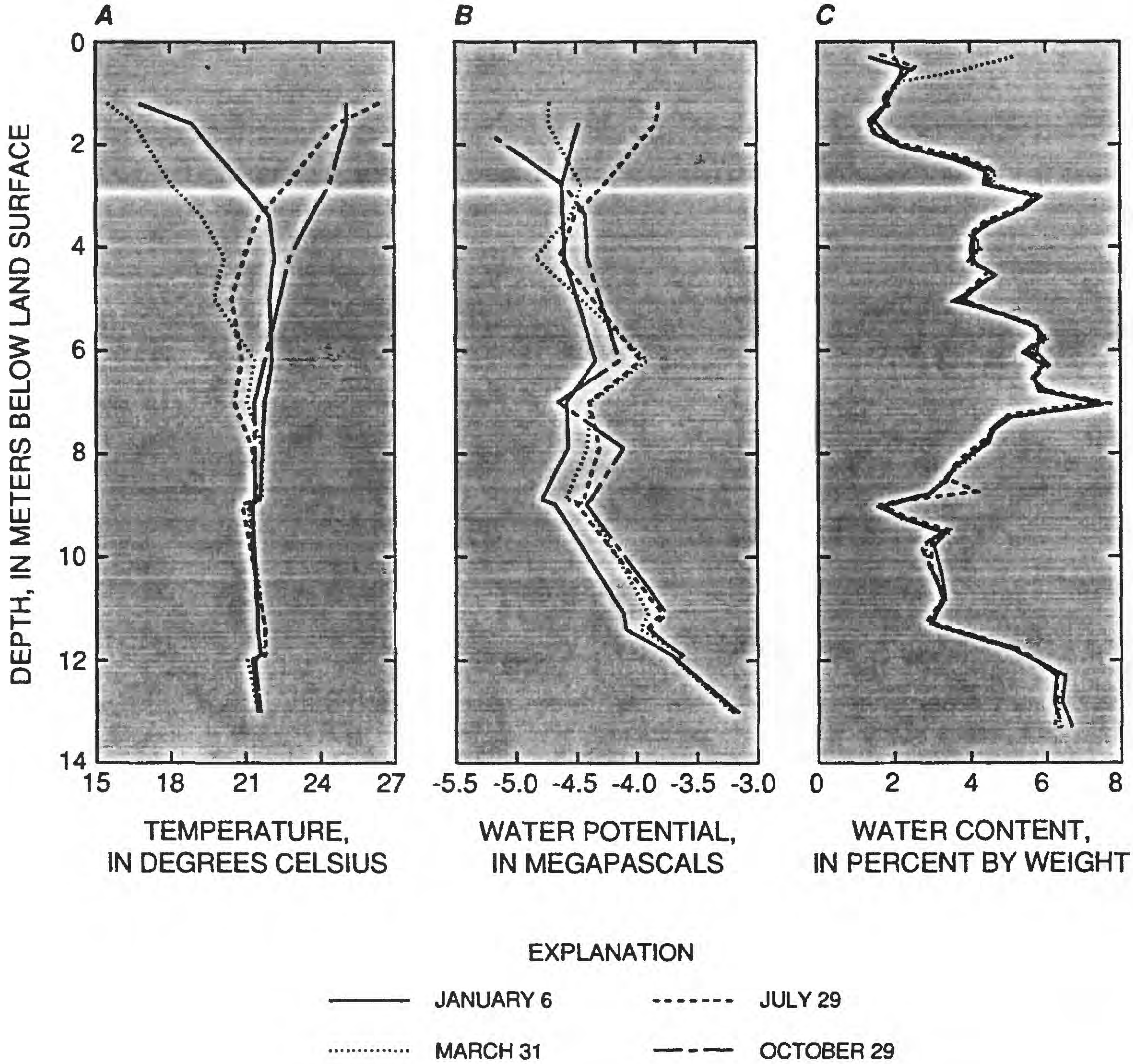



FIGURE 21.--Profiles of (*A*) temperature, (*B*) water potential, and (*C*) water content with depth at study site for selected dates in 1987. Temperature and water-potential data are from thermocouple psychrometers installed at monitoring shaft and instrument borehole. Water content is from neutron-probe access tube N1. Location of monitoring shaft, instrument borehole, and neutron-probe access tube, N1, is shown in figure 2.

An estimate was made of the amount of precipitation needed to reverse the observed upward gradient and produce a downward flux. Infiltration was assumed limited by the saturated hydraulic conductivity of the surficial deposits of sandy silt (2 cm/d). In addition, the distribution of moisture to a depth of 13 m was assumed instantaneous. With these assumptions, 8 cm of rain falling over a 4-day period at a constant rate of 2 cm/d is needed to reverse the observed gradients. Even under these unlikely conditions, downward flux would probably be less than $10^{-2}$ cm/d. For more realistic conditions of moisture redistribution and with capillary barriers retarding the wetting front advance, it may take several weeks of constant rainfall for such a reversal to occur. In addition, high evaporative demand and plant transpiration at the site could be expected to remove much of the available water in the sediments and again produce upward gradients. Perhaps the most likely scenario for significant amounts of downward percolation at the study site would be a large storm during the winter months that eroded the surficial deposits of sandy silt exposing the gravels below. Under such conditions, infiltration would not be limited by the surficial sandy silt and considerable percolation might occur through the highly permeable gravels.

## Estimates of Water-Vapor Flux

Water vapor can move in response to barometric pressure changes (convection) and in response to vapor-density gradients (diffusion). Water-vapor convection, although it may be an important factor in water movement in desert alluvium, is not considered in this report because measurements of subsurface air-pressures and *in situ* air-permeabilities were not made during the study. Thus, convective fluxes could not be calculated. For this reason the following discussion only considers water-vapor diffusion.

Diffusion is fairly independent of sediment texture and depends for the most part on total volume and tortuosity of interconnected pores. Water-vapor movement can occur both around and through liquid-water barriers. The diffusion equation for vapor can be written:

$$q_v = -D_v \frac{d\rho_v}{dz} \tag{14}$$

where $q_v$ = vapor flux, in grams per square centimeter per second;
$D_v$ = diffusion coefficient for water vapor, in square centimeters per second;
$\rho_v$ = vapor density, in grams per cubic centimeter; and
$z$ = depth, in centimeters.

The diffusion coefficient for water vapor is a function of temperature and pressure. Factors that influence $D_v$ are discussed by Sherwood and others (1975, p. 8-17); however, only minor error is introduced if $D_v$ is considered a constant (Hanks and Ashcroft, 1980, p. 93).

Vapor density ($\rho_v$) is a function of water potential and temperature. In isothermal systems, water potential is the force driving water-vapor movement (Hillel, 1982, p. 124-125). When temperature differences do occur, they cause changes in vapor density that far exceed changes resulting from water-potential gradients. For example, a change in temperature of 1°C has nearly the same effect upon vapor pressure as a water-potential change of 10 MPa (Hillel, 1982, p. 125).

Both water potentials and temperatures were used to calculate water-vapor flux at the study site. Temperature measurements from TCP's installed in the monitoring shaft were used for calculating vapor flux below a depth of 3 m. Temperature measurements from TCP's installed in the instrument borehole were used above a depth of 3 m, because no TCP's were installed above that depth of the monitoring shaft.

The mean temperature in the sediments for 1987 increased slightly with depth (table 4). Above a depth of 3 m, the mean temperature was generally less than 21°C, whereas below a depth of 10 m, the mean temperature was generally more than 21.5°C. Mean temperatures for psychrometers installed in the instrument borehole were slightly greater than those at similar depths in the monitoring shaft. The reason for the temperature difference is unknown, but it could be related to the different methods of installation. However, the general increase in temperature with depth in both the monitoring shaft and the instrument borehole suggests that the potential for vapor movement is, on the average, slightly upward.

Mean monthly vapor-density gradients were determined for 1987 between selected TCP's using monthly averages of the daily water potentials and temperatures. Vapor densities for a given water potential and temperature were calculated as described by Philip and De Vries (1957, p. 224, eq. 3). Mean monthly vapor-density gradients for selected depth intervals are listed in table 5. The uncertainty of the calculated vapor-density gradient is estimated at ±0.6 $(g/cm^3)/cm$ and is based on the uncertainty of the TCP temperature measurement of ±0.2°C. The use of mean monthly temperatures and water potentials to calculate vapor-density gradients may cause an overestimation or an underestimation depending on the month. The direction of monthly, vapor-density gradients determined from TCP's between depths of 1.2 and 9 m varied seasonally during 1987 (table 5).

Water-vapor flux was calculated using eq. 14, assuming 1 g of water was equal to 1 $cm^3$ of water, the gradients in table 5, and an assumed diffusion coefficient for water vapor of 0.2 $cm^2/s$ (Hanks and Ashcroft, 1980, p. 93). The maximum water-vapor flux calculated was $-10^{-3}$ cm/d (upward) in December and $10^{-3}$ cm/d (downward) in July between the depths of 1.2 and 1.6 m. The maximum water-vapor flux was considerably less than $10^{-3}$ cm/d at depths greater than 1.6 m below land surface. Also the maximum water-vapor flux lagged with depth depending on the time of year (table 5). For example, the maximum upward vapor flux between the depths of 1.6 and 2.8 m was during January, whereas the maximum upward vapor flux between the depths of 4 and 5 m was during March, and the maximum between the depths of 8 and 9 m was during June. Times of maximum upward water-vapor flux and maximum downward water-vapor flux were usually 6 months apart. Water-vapor flux between the depths of 11 and 13 m is in response to the water-potential gradient, because temperature did not vary significantly. Water-vapor flux was always upward in this interval and about $5 \times 10^{-5}$ cm/d.

The monthly water-vapor fluxes were summed to obtain an annual water-vapor flux for each selected depth interval. At depths between 1.2 and 5 m, annual water-vapor flux was slightly downward and between depth of 5 and 13 m, the annual flux was slightly upward. Average annual water-vapor fluxes were calculated between the depths of 3 and 4 m ($3 \times 10^{-5}$ cm/d downward), 5 and 6 m ($2 \times 10^{-5}$ cm/d upward), 9 and 11 m ($2 \times 10^{-5}$ cm/d upward), 11 and 12 m ($4 \times 10^{-5}$ cm/d upward), and 12 and 13 m ($5 \times 10^{-5}$ cm/d upward). These results suggest that water-vapor flux during 1987 was toward a depth of 5 m.

## SUMMARY AND CONCLUSIONS

A commercial facility for disposal of low-level radioactive wastes is in the Amargosa Desert about 17 km south of Beatty, Nev. Since 1962, approximately 113,000 $m^3$ of wastes have been placed in trenches that range from about 90 to 240 m long, 2 to 15 m deep, and as much as 35 m wide. The trenches have been excavated into unsaturated sediments consisting mostly of unconsolidated alluvial-fan, debris-flow, and fluvial deposits composed of a poorly sorted mixture of cobbles, sand, and gravel with minor amounts of silt and clay. These deposits extend to a depth of 70 to 80 m below land surface. Underlying these deposits is an areally extensive clay layer about 20 m thick. Alternating layers of sand and silty sand predominate below the clay layer to a depth of about 125 m and overlie a gravelly sand that extends to a depth greater than 200 m. Water levels in wells range from a depth of 85 to 115 m below land surface. The general direction of ground-water flow is to the south-southwest with a gradient of approximately 0.06 m/m.

Thick unsaturated zones in arid environments offer potential advantages for disposal of radioactive wastes, but little is known about the natural movement of water near such facilities. A study was begun in 1982 to better define the rates and direction of water movement through the unsaturated deposits in the vicinity of the disposal facility near Beatty, Nev.--one of the most arid disposal sites in the United States. This report discusses the analyses of data collected between 1983 and 1988. For this study, a small site was chosen outside the southwest corner of the waste-disposal facility.

Climate at the study site is characterized by hot summers, cool winters, and little precipitation. Mean daily temperatures from January 1985 through December 1987 ranged typically between 25 and 34°C in the summer to between 2 and 15°C in the winter. Maximum summer temperatures approached 45°C and minimum winter temperatures were between -5 and 5°C. Total precipitation from October 1984 to February 1988 was 326 mm--much of this precipitation fell between the months of November and March. Only six storms produced precipitation in excess of 25 mm. Both temperature and precipitation measurements were within the historic norms for the area.

Most of the sediments at the study site are composed of a poorly sorted mixture of sand and gravel in a fine sand and silt matrix. The surficial deposits are composed of a sandy silt similar in texture and thickness to that found at the surface of the waste-disposal facility. Downward percolation of precipitation is impeded by this relatively impermeable layer. Beneath the sandy silt and extending to a depth of 2.2 m is a moderately sorted cobbly sand and gravel with little silt and clay, which acts as a capillary barrier to unsaturated flow. Between the depths of about 6 and 8.5 m, the deposits are a moderately sorted coarse sand underlain by a moderately sorted cobbly sand and gravel to a depth of 9 m. This cobbly sand and gravel also acts as a capillary barrier to unsaturated flow.

Porosity and bulk density were determined from core samples collected during the drilling of test holes. Porosity of the samples ranged from 25 to 43 percent and bulk density ranged from about 1.4 to 1.8 $g/cm^3$. Generally, porosities were higher and bulk densities lower for the moderately sorted cobbly sand and gravel beneath the surficial sandy silt and for the moderately coarse sand and cobbly sand and gravel between the depths of 6 and 9 m.

Chloride concentrations in the sediments were highest between the depths of 2 and 8 m; maximum concentrations were measured between the depths of 2 and 4 m below land surface. The distribution of salts and chloride in the sediments suggests that water from previous precipitation events may have percolated to a depth between 2 and 8 m but that the infiltrating water was then transpired or the water continued to move as a vapor, leaving the salts behind.

Water content of the unsaturated sediments was measured using a neutron-moisture probe. Gravimetric water content between the depths of 1 and 30 m ranged from 4 to 8 percent (0.04 to 0.08 g/g). Higher water content was observed at depths of about 7, 14, and 25 m, which correspond to zones of finer grained deposits. Lower water content was observed at depths of about 1.5 and 9 m, which correspond to zones of cobbly sand and gravel. The gravimetric water content of the surficial sandy silt ranged from almost air dry in the summer to 21 percent (0.21 g/g) following precipitation. Redistribution of moisture following precipitation was not observed below a depth of 1 m.

Saturated hydraulic conductivity was determined from selected core samples using either an air permeameter or a constant-head permeameter. The hydraulic conductivity of the surficial sandy silt was determined to be about 2 cm/d. The saturated hydraulic conductivity of six samples collected between the depths of 0.6 and 7.3 m ranged from 1 to 48 cm/d. Unsaturated hydraulic conductivities were also determined for the selected core samples at water potential of -1.5 MPa and at estimated field water potential and water content. The values ranged from $10^{-4}$ to $10^{-20}$ cm/d.

Two methods of installing thermocouple psychrometers (TCP's) were used to determine water potentials in the sediments during the study. The first method involved installation of a large diameter (1.52-m) vertical shaft to a depth of 13.7 m below land surface. Horizontal holes 5 cm in diameter were drilled a minimum of 3 m outward beneath an undisturbed area from prefabricated access ports in the shaft. A 2.5-cm-diameter pipe made of polyvinyl chloride (PVC) was installed in each horizontal hole and sealed with a polyurethane foam. A Delrin plug was installed at the end of each pipe. The TCP was then inserted into the pipe and a matching plug coated with rubber cement was used to seal the end of the TCP from the pipe and shaft. Psychrometers were installed between the depths of 3 and 13 m below land surface. Determination of water potentials using this method indicates that reasonable estimates of water potentials can be obtained beneath an undisturbed area but the method is expensive and may best be suited for long-term studies. The method could be improved by finding an easier way to isolate the TCP at the end of the horizontal access tube.

The second method involved installation of TCP's in a borehole. The borehole was filled with drill cuttings between the depths of 6 and 12 m. Above the depth of 6 m, thin layers of dry and wet bentonite were used to seal the hole between TCP's. The disadvantage to this method is that TCP's could not be retrieved for recalibration or replaced. In addition, unexplained anomalies in TCP readings may be caused by inadequate sealing in the borehole. Perhaps reliable estimates of water potential could be obtained from TCP's in a borehole by designing a method for retrieving the TCP's and developing a better technique for sealing the borehole between TCP's.

Water potentials were most variable near the surface. Water potentials were often beyond the range of the TCP's above a depth of 1 m (less than about -7 MPa) but quickly increased to more than 0.5 MPa following precipitation. Water potentials measured between depths of 3 and 7 m ranged from -3.7 to -5.5 MPa and varied seasonally. Water potentials between the depths of 9 m and 11.5 m did not vary seasonally, but have steadily increased since the TCP's were installed in 1986. Water-potential measurements at depths of 12 and 13 m in the monitoring shaft have remained stable at -3.7 and -3.2 MPa, respectively. Water-potential gradients were averaged monthly for selected depth intervals using mean daily water potentials for 1987. Water-potential gradients varied during the year and were both upward and downward between the depths of about 2 and 9 m. Gradients below a depth of 9 m were consistently upward.

Temperatures within the upper 0.6 m of sediments varied diurnally and in response to infiltrating water following precipitation. Temperatures varied seasonally to a depth of about 8 m. Largest seasonal variations were observed near the surface and decreased with depth. Temperatures in sediments between the depths of 9 and 13 m were nearly constant at about 21°C. Maximum and minimum temperatures lagged in time with depth. The movement of heat through the sediments was estimated from changes in temperature as a result of seasonal changes in air temperature; the rate of heat flow was estimated to be about 5 cm/d. From seasonal variations in temperature, the thermal diffusivity of the sediments was estimated to be 0.67 $mm^2/s$, indicating the sediments are good thermal conductors.

Water fluxes within the upper 9 m of sediments, both as water vapor and as liquid, were estimated to be small, of short duration, and of approximately equal rates upward and downward at different times of the year. Liquid-water fluxes were estimated to range from about $1 \times 10^{-2}$ to $1 \times 10^{-9}$ cm/d. The large range in estimated flux is the result of a large uncertainty in the unsaturated hydraulic conductivity. Water-vapor fluxes (not including movement as a result of barometric-pressure changes) were estimated to range from about $1x10^{-3}$ to $1x10^{-6}$ cm/d. Results from this study imply that water movement in the upper 9 m at the study site is in approximate equilibrium and water moves from one depth to another in response to seasonal changes in water potential and temperature.

Between the depths of 9 and 13 m, water potential and temperatures were nearly constant, and water-potential and vapor-density gradients were consistently upward. Liquid-water flux was estimated to range from about $1x10^{-2}$ to $1x10^{-8}$ cm/d, depending on the value of the unsaturated hydraulic conductivity used in the estimate. Water-vapor flux was estimated from $2x10^{-5}$ to $5x10^{-5}$ cm/d on the basis of vapor-density gradients only.

In summary, the upper 13 m of the unsaturated zone can be divided into upper, middle, and lower zones on the basis of water potential, water content, temperature, and sodium-chloride content. The upper zone, from the surface to a depth of about 2 m, is subject to rapid changes in water potentials, water contents, and temperatures in response to precipitation and air-temperature fluctuations. The middle zone, from a depth of about 2 m to 9 m, has unchanging water contents but seasonal trends in water potential as vapor moves upward and downward in response to seasonal temperature trends. This zone may also be the maximum depth that percolating water reaches, as indicated by the accumulation of sodium chloride in this interval. The lower zone, from a depth of about 9 m to at least 13 m, is characterized by relatively unchanging water potential, water-contents, and temperatures. Water-potential and vapor-density gradients in this zone indicate upward movement of water by liquid-water flow and as water vapor.

Two main factors combine to limit downward percolation of water at the study site. Of primary importance is the arid climate, which has low precipitation and potentially high evapotranspiration. In addition, the combination of surficial sandy silt, which limits infiltration, and an underlying gravel, which acts as a capillary barrier, effectively impedes downward flow. The most likely conditions for downward percolation of water seem to be either an intense storm that erodes the surficial sandy silt or a long period of increased precipitation that reverses existing upward gradients.

# REFERENCES CITED

Allison, G.B., and Hughes, M.W., 1983, The use of natural tracers as indicators of soil-water movement in a temperate semi-arid region: Journal of Hydrology, v. 60, p. 157-173.

American Petroleum Institute, 1952, Recommended practice for determining permeability of porous media: Dallas, Texas, American Petroleum Institute, API RP 27, 27 p.

Barbour, M.G., Cunningham, G., Oechel, W.C., and Bamberg, S.A., 1977, Growth and development, form and function, *in* Mabry, T.J., Hanzikes, J.H., and DiPeo, D.R., Jr., eds., Creosote bush--Biology and chemistry of Larren in New World deserts: Strousburg, Penn., Dowden, Hutchinson, Ross, p. 48-91.

Bedinger, M.S., 1989, Geohydrologic aspects for siting and design of low-level radioactive-waste disposal: U.S. Geological Survey Circular 1034, 36 p.

Bouwer, Herman, and Rice, R.C., 1984, Hydraulic properties of stony vadose zones: Groundwater, v. 22, no. 6, p. 696-705.

Brown, R.W., and Bartos, D.L., 1982, A calibration model for screen-caged peltier thermocouple psychrometers: U.S. Department of Agriculture, Forest Service, Intermountain Research Station, Research Paper INT-293, 155 p.

Brown, R.W., and Collins, J.M., 1980, A screen-caged thermocouple psychrometer and calibration chamber for measurements of plant and soil water potential: Agronomy Journal, v. 72, p. 851-854.

Brown, R.W., and Johnston, R.S., 1976, Extended field use of screen-covered thermocouple psychrometers: Agronomy Journal, v. 68, no. 6, p. 995-996.

Campbell, G.S., 1974, A simple method for determining unsaturated conductivity from soil moisture retention data: Soil Science, v. 117, no. 6, p. 311-315.

-----1977, An introduction to environmental biophysics: New York, Springer-Verlag, 159 p.

Clebsch, Alfred, Jr., 1968, Geology and hydrology of a proposed site for burial of solid radioactive waste southeast of Beatty, Nye County, Nevada, *in* Morton, R.J., Land burial of solid radioactive wastes--Study of commercial operations and facilities: Atomic Energy Commission, NTIS Report WASH-1143, p. 70-100. Available only from National Technical Information Service, U.S. Department of Commerce, Springfield, VA 22161.

Fischer, J.N., 1986, Hydrogeologic factors in the selection of shallow land burial sites for the disposal of low-level radioactive waste: U.S. Geological Survey Circular 973, 22 p.

Folk, R.L., 1974, Petrology of sedimentary rocks: Austin, University of Texas, Hemphill Publishing Co., 170 p.

Fouty, Suzanne, 1989, Chloride mass-balance as a method for determining long-term ground-water recharge rates and geomorphic surface stability in arid and semi-arid regions--Whiskey Flat and Beatty, Nevada: Tucson, University of Arizona, unpublished M.S. thesis, 130 p.

Gardner, W.H., 1965, Water content, *in* Black, C.A., Evans, D.D., White, J.L., Ensmiger, L.E., and Clark, F.E., (eds.), Methods of soil analysis, part 1, Physical and mineralogical properties, including statistics of measurement and sampling: Madison, Wisc., American Society of Agronomy, Agronomy Series no. 9, p. 82-127.

Hammermeister, D.P., Blout, D.O., and McDaniel, J.C., 1985, Drilling and coring methods that minimize the disturbance of cuttings, core, and rock formation in the unsaturated zone, Yucca Mountain, Nevada: National Water Well Association, Conference on Characterization and Monitoring of the Vadose (Unsaturated) Zone, Denver, Colo., November 1984, Proceedings, p. 507-541.

Hanks, R.J., and Ashcroft, G.L., 1980, Applied soil physics: New York, Springer-Verlag, 159 p.

Harrill, J.R., Gates, J.S., and Thomas, J.M., 1988, Major ground-water flow systems in the Great Basin region of Nevada, Utah, and adjacent States: U.S. Geological Survey Hydrologic Investigations Atlas HA-694-C, 2 plates, scale 1:1,000,000.

Hillel, Daniel, 1982, Introduction to soil physics: New York, Academic Press, 364 p.

Isaacson, R.E., Brownell, L.E., and Hanson, J.C., 1974, Soil moisture transport in arid site vadose zones: Atomic Energy Commission contract AT(45-1)-2130, 93 p.

Johnson, M.D., and Lowery, Birl, 1985, Effect of three conservation tillage practices on soil temperature and thermal properties: Soil Science Society of America Journal, v. 49, no 6, p. 1547-1552.

Kearl, P.M., 1982, Water transport in alluvial desert soils: Reno, University of Nevada, Desert Research Institute Publication 45024, 126 p.

Kilroy, K.C., 1991, Ground-water conditions in Amargosa Desert, Nevada-California, 1952-87: U.S. Geological Survey Water-Resources Investigations Report 89-4101, 93 p.

Klute, Arnold, 1986, Water retention--laboratory methods, *in* Klute, Arnold (ed.), Methods of soil analyses, part 1, physical and mineralogical methods: Madison, Wisc., American Society of Agronomy, Agronomy Monograph 9, part 1, 1188 p.

Mark Group, The, 1989, Exploratory boring and monitoring well installation program US Ecology RCRA facility, Beatty, Nevada: Las Vegas, Nev., 82 p., 10 appendices.

Mehuys, G.R., Stolzy, L.H., Letey, J., and Weeks, L.V., 1975, Effects of stones on the hydraulic conductivity of relatively dry desert soils: Soil Science Society of America Proceedings, v. 39, no. 1, p. 37-42.

Morgan, D.S., and Fischer, J.M., 1984, Unsaturated-zone instrumentation in coarse alluvial deposits of the Amargosa Desert near Beatty, Nevada, *in* Proceedings of the Sixth Annual Participants' Information Meeting, U.S. Department of Energy Low-Level Waste Management Program: Available from National Technical Information Service, U.S. Department of Commerce, Springfield, VA 22161, CONF-8409115, p. 617-630.

Nichols, W.D., 1987, Geohydrology of the unsaturated zone at the burial site for low-level radioactive waste near Beatty, Nye County, Nevada: U.S. Geological Survey Water Supply Paper 2312, 57 p.

Philip, J.R., and De Vries, D.A., 1957, Moisture movement in porous materials under temperature gradients: American Geophysical Union Transactions, v. 38, no. 2, p. 222-232.

Richardson, R.M., 1962, Significance of climate in relation to the disposal of radioactive waste at shallow depth below ground: Colloque International sur la Retention et la Migration des Ions Radioactifs dans les Sols (Health Physics Society and Commissariat a l'Energie Atomique), Saclay, France, October 1962, Proceedings, p. 207-211.

Schulz, R.K., Ridky R.W., and O'Donnell, Edward, 1987, Control of water infiltration into near surface low-level waste disposal units, annual report, October 1985-September 1986: U.S. Nuclear Regulatory Commission Report NUREG/CR-4918, v. 1, 20 p.

-----1988, Control of water infiltration into near surface low-level waste disposal units, task report--A discussion: U.S. Nuclear Regulatory Commission Report NUREG/CR-4918, v. 2, 24 p.

Sherwood, T.K., Pigford, R.L., and Wilke, C.R., 1975, Mass transfer: New York, McGraw-Hill Book Company, 677 p.

Spiegel, M.R., 1961, Schaum's outline series of theory and problems of statistics: New York, McGraw-Hill Book Company, 359 p.

Stone, W.J., 1984, Recharge in the Salt Lake coal field based on chloride in the unsaturated zone: New Mexico Bureau of Mines and Mineral Resources, Open-File Report 214, 64 p.

Weibe, H.H., Campbell, G.S., Gardner, W.H., Rawlins, S.L., Cary, J.W., and Brown, J.W., eds., 1971, Measurement of plant and soil water status: Logan, Utah State University, Utah Agricultural Experiment Station Bulletin 484, 71 p.

Winograd, I.J., and Thordarson, William, 1975, Hydrogeologic and hydrochemical framework, south-central Great Basin, Nevada-California, with special reference to the Nevada Test Site: U.S. Geological Survey Professional Paper 712-C, 126 p.

Wollenberg, H.A., Wang, J.S.Y., and Korbin, G., 1983, Nuclear waste isolation in the unsaturated zone of arid, *in* Mercer, J.W., Rao, P.S.C., and Marine, I.W., eds., Role of the unsaturated zone in radioactive and hazardous waste disposal: Ann Arbor, Mich., Ann Arbor Publishing, p. 195-210.