REVIEW OF MECHANISMS, METHODS, AND THEORY FOR DETERMINING RECHARGE TO SHALLOW AQUIFERS IN NORTH DAKOTA

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SELECTED FACTORS FOR CONVERTING INCH-POUND UNITS TO METRIC UNITS

For those readers who may prefer to use metric (International System) units rather than inch-pound units, the conversion factors for the terms used in this report are given below.

Multiply inch-pound unit	By	To obtain metric unit
Acre	0.4047	hectare
Bar	100,000	pascal
Calorie	4,184	joule
Foot	0.3048	meter
Inch	25.40	millimeter
Kilowatt hour per acre	889.5	joule per square meter
Ounce, avoirdupois	28.35	gram
Square mile	2.590	square kilometer

To convert degrees Fahrenheit (°F) to degrees Celsius (°C), use the following formula: °C = (°F-32)x5/9.

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

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ABSTRACT

Effective management of ground-water resources requires knowledge of all components of the water budget for the aquifer of interest. Efforts to simulate ground-water flow prior to development and the effects of proposed pumping in several of North Dakota's shallow glacial aquifers have been hindered by the lack of reliable estimates of ground-water recharge. This study was done to (1) review the methods that have been used to measure recharge, (2) review the theory of unsaturated flow and the methods for characterizing the physical properties of unsaturated media, (3) consider the relative merits of a rigorous data-intensive approach versus an estimation approach to the study of recharge, and (4) review past and current agronomic research in North Dakota for applicability of the research and the data generated to the study of recharge.

Direct, quantitative techniques for evaluating recharge are rarely applied. The theory for computing fluxes in unsaturated media is well established and numerous physics-based models that effectively implement the theory are available, but the data required for the models generally are lacking. Many parametric approaches have been developed to avoid the large data requirements of the physics-based approaches for analyzing flow in the unsaturated zone. However, the parametric approaches normally include fitting coefficients that must be calibrated for every study site, thereby detracting from the general utility of the parametric approach.

The functional relation of matric potential to moisture content is required for physics-based soil-water models, whether analytic or numeric. Laboratory methods to determine these relations are tedious, costly, and may not give results representative of the soils as they occur in the field. Many models have been proposed to estimate the moisture-characteristic curve and hydraulic-conductivity function from basic soil properties, but none yield results that are universally satisfactory. In situ methods, because they require minimal disturbance of the soil profile and may be used repeatedly on the same soil mass, have become the preferred means for acquiring physical data, especially hydraulic conductivity. Hydrologic investigations, except for recent studies of hazardous-waste disposal sites, rarely have included physical characterizations of unsaturated media.

Any of four phenomena could hinder attempts to simulate unsaturated flow in settings typical of North Dakota; variability of soil properties, hysteresis, frozen ground, and macropore development. The spatial and temporal variability of soil properties probably is the greatest complicating phenomenon and must be dealt with by detailed characterization of the properties. Hysteresis can detract from the accuracy of flow calculations

for some soils under certain conditions but, for the present, our scant knowledge of soil physical properties is a greater hindrance to reliable soil-water modeling than is the hysteresis phenomenon. Although seasonally frozen ground undoubtedly affects hydrologic processes in North Dakota, much more research is needed before meaningful quantitative treatment is possible. Finally, macropores can influence soil-water movement significantly, but macropore development may not be common on the intensively farmed, coarse-textured soils that typically overlie North Dakota's glacial aquifers. Lysimetry currently is the only reliable means of analyzing macropore flow.

The soil-related research that has been conducted in North Dakota to date (1983) provides little of the type of information required to estimate ground-water recharge. Useful data could be developed by systematically evaluating the hydraulic characteristics of the prominent soil types overlying North Dakota's shallow glacial aquifers. These data would be required to enable use of a physics-based approach to estimating recharge. The size of the aquifer under study, its economic value, and the resources available for data collection should be considered when choosing between parametric or physics-based methods.

INTRODUCTION

North Dakota's ground-water resources have been developed at an accelerated rate in recent years, primarily to supply water for irrigation. Applications for new ground-water-use permits are evaluated with regard to the effect of the proposed use on prior users and on the overall water budget of the aquifer. When the application involves aquifers that already are extensively developed, the North Dakota State Water Commission normally uses mathematical flow models to evaluate the effect of the proposed withdrawals. The purpose of the modeling is to learn enough about the aquifer flow system to make responsible and equitable decisions concerning individual use requests as well as the overall management of the aquifers.

Most of the aquifers either currently developed or projected for large-scale development are of Quaternary age and consist mainly of glacial outwash and alluvial deposits. Generally, the aquifers are overlain by a soil mantle only a few feet thick and have water tables less than 10 feet below the land surface. The aquifers are recharged directly by infiltration of precipitation or by percolation through thin surficial deposits. Much water, both before and after it reaches the saturated zone, is lost to evapotranspiration. Also, occurrences of rejected recharge and surface discharge, evident as bogs and sloughs, are common.

Because reliable data concerning recharge and evapotranspiration magnitudes for North Dakota aquifers are lacking, a typical modeling approach has been to treat these two components of the water budget as unknowns and to make estimates of their values as part of the model calibration process. These estimates generally are not reliable due to the nonuniqueness of any calibrated best-fit combination of recharge and evapotranspiration. Also, uncertainties in the values of other variables affecting the calibrated balance can further diminish the reliability of the recharge and evapotranspiration estimates.

Errors in model projections that result from the use of these estimates decrease the effectiveness of the aquifer management process. The accuracy of model projections may not be a critical factor for aquifers having little development or where large possible errors in model results would not result in the over appropriation of water in the aquifer. However, if appropriation decisions for an intensively developed aquifer are based on the projections of a model having questionable validity, the water supplies of existing users could be adversely affected, or, conversely, opportunities for additional safe development of the aquifer could be denied for no good reason.

To provide data for ground-water-flow models is not the only worthwhile reason to measure recharge or to study the mechanisms by which it occurs. Solutions to many water-resources problems that involve natural flow systems, whether above or beneath the land surface, could be achieved more readily with some knowledge of the water-transmitting characteristics of the unsaturated zone. For example, an important related application of unsaturated-zone properties is in accounting for the moisture in the soil for rainfall-runoff modeling. Determining recharge magnitudes and processes would, in fact, benefit a variety of water-resource concerns, such as agricultural water management (irrigation scheduling, design of surface and subsurface drainage systems), and the selection, management, or cleanup of hazardous waste disposal sites.

Toward the goal of improved management of the State's water resources, the North Dakota State Water Commission entered into a cooperative project with the U.S. Geological Survey to make a literature study of the methods appropriate for a determination of recharge for shallow glacial aquifers. This report describes the findings of that literature study.

Objectives

Reliable estimates of ground-water recharge are needed for North Dakota's shallow glacial aquifers. Accordingly, the objectives of this investigation are to: (1) Review the hydrologic literature dealing with the determination of ground-water recharge for shallow water-table settings; (2) review the pertinent theory of flow in unsaturated or variably saturated media, including the characterization of the physical properties that appear in the governing equations; (3) consider the relative merits of a rigorous data-intensive approach versus a parametric approach to the study of recharge; and (4) review the agronomic research, past and present, in North Dakota for applicability of the research and the resulting data to the study of recharge.

The Recharge Problem

A review of the hydrologic literature indicates that little precedent is available for direct, quantitative approaches to studying recharge. Hydrologists typically have investigated the physical and hydraulic properties of aquifers in great detail, but generally have ignored similar properties of the surficial media that constitute the unsaturated zone. Yet these media receive precipitation and store the water while a variety of forces affect its movement and generally diminish its quantity, and finally convey some of the precipitation to the water table. In short, ground-water recharge is a result of dynamic processes that occur in the unsaturated zone. Consideration of these unsaturated-zone processes, however, had not been a prominent issue in hydrogeology until the last few years.

Early theoretical developments in hydrogeology dealt with the effects of pumping from areally extensive, confined aquifers. Theis's (1935) solution to the equation for unsteady radial flow of confined ground water stimulated what has become a very large amount of literature on well hydraulics. Although this literature now includes analytical models pertaining to a variety of assumed initial and boundary conditions for confined and unconfined aquifers (Walton, 1979), these models generally do not account for the vertical flux of water through the unsaturated zone.

With the widespread availability of digital computers in the 1960's, hydrologists began developing computer models that could simulate ground-water flow in two or three dimensions. The computer models are able to encorporate much more realistic boundary conditions and variable aquifer properties than are the analytic models. Recharge, however, is simulated by the typical ground-water-flow model simply as a prescribed flux, applied uniformly or spatially varying throughout the aquifer area. This flux is simulated by the addition of a general source or sink term that also may include fluxes due to evapotranspiration, leakage, and pumpage. (See, for example, Trescott and others, 1976.) Typically, few or no data are available concerning recharge magnitudes, so the investigator must use estimated values and then adjust the estimates by model calibration. If the distributions of hydraulic head, hydraulic conductivity, specific storage, and the magnitudes and locations of boundary fluxes (including evapotranspiration and leakage) and pumpage are known fairly well, this estimate-calibrate technique may give satisfactory results. If one or more of the preceding components are poorly defined, however, as often is the case, the simulation results will be of dubious validity, particularly if the model is sensitive to the component.

A possible solution to the common dilemma just described would involve direct study of the process of ground-water recharge. Such a study, however, requires examination of the unsaturated zone. Although the general equation describing the flow of water through unsaturated porous media was developed more than 50 years ago (Richards, 1931), hydrologists did not seem to take notice of its potential for application to the study of ground water until the 1960's. Even since then, field applications of the soil physics theory in hydrologic studies have been rare. The theory has not been put to general practical use.

Both hydrologists and soil physicists have called attention to the unnatural dichotomy of the two disciplines and have warned of the need to consolidate, or at least coordinate, their respective accumulations of knowledge. In remarks prefacing a detailed discussion of a simulation study of ground-water recharge and discharge, Freeze (1969) noted "***until very recently, ground-water hydrologists have avoided, with a singular determination, studies that included consideration of the unsaturated zone. By the same token, soil physicists working in the unsaturated domain have given very little attention to the saturated flow processes that occur below agricultural depths. If the twain must meet, and they must, an obvious subject for the early application of an integrated saturated-unsaturated approach is that of ground-water recharge and discharge."

A reminder of the continuing failure to address, by direct study, the mechanisms of natural ground-water recharge was provided by Peck (1979) in remarks to the 1979 Canberra Symposium on "The Hydrology of Areas of Low Precipitation." He stated, "It is only in the last 10-20 years that much attention has been directed towards recharge and loss from aquifers. Even now, study of the recharge process tends to fall between soil physics and ground-water hydrology. The first group has traditionally ignored the fate of water which passes beyond the rooting depth of agricultural plants, often only 1-2 m (meters), while the second group usually treats a water table as the upper boundary of the flow domain for their consideration. ***limitations on our existing knowledge of flow processes in particular circumstances emphasize the need for further caution in applying any method for estimating recharge. Further field studies of recharge are required to aid the development of sound theory."

A recent paper by Rushton and Ward (1979) reviewed some of the more commonly used water-balance techniques, most having little physical basis, for estimating ground-water recharge. The authors found that considerable adjustment to equation parameters by empirical fitting was generally necessary to achieve satisfactory results. They concluded that, "***future research should focus attention on the recharge mechanism itself, to determine whether the proposed models are on the right lines, rather than on the classical model for recharge estimation with its inherent weaknesses. It is hoped that in light of this presentation hydrologists will be encouraged to carry out critical recharge studies in their own areas."

The theme espoused by these three scientists appears repeatedly in the literature. That is, hydrologists should not ignore the strong influence of the unsaturated zone on hydrologic processes occurring both above and beneath the land surface. Soil scientists, on the other hand, should look beyond profile studies, or that part of the hydrosphere between the land surface and the base of the root zone, to evaluate the relationship of the soil zone to other parts of the hydrologic cycle.

The paucity of quantitative information regarding the movement of water in the unsaturated zone for dominant soil types and vegetative covers inhibits reliable estimates of ground-water recharge. Thus the ability to provide optimal management for our increasingly valuable water resource also is hindered.

RECHARGE MECHANISMS

Conceptually, the analysis of ground-water recharge is quite straightforward. Aquifers may be recharged by any combination of three mechanisms.

Aquifers that lie at or near the land surface may be recharged directly by rainfall and snowmelt that infiltrates the soil surface and percolates to the water table. Shallow aquifers also may receive recharge from surface-water bodies in areas where lake or stream beds are in direct contact with the aquifer. Finally, aquifers may be recharged by leakage through adjacent, less permeable materials. Aquifers that lie at depths beneath the reach of the first two mechanisms may be recharged by leakage only.

Estimation of recharge, then, requires recognition of the effective mechanism and a measurement of the prevailing water fluxes. The crux of the problem of determining ground-water recharge, of course, is in making a reliable measurement of that water flux.

Hydrologists have devoted much more attention to the latter two recharge mechanisms than to the first--probably because the latter two generally are considered to be processes of the saturated zone. Analytic models of the exchange of water between streams and confined aquifers (Ferris, 1951) or unconfined aquifers (Rorabaugh, 1960) were introduced in the early fifties and sixties. Walton (1979) listed many of the unsteady-state analytic models that treat stream-aquifer interaction and the leaky artesian condition regarding the well hydraulics of confined aquifers. Developments or reviews of stream-aquifer and aquifer-confining bed relationships also may be found in the works of Ferris and others (1962), Bentall (1963), Lohman (1972), Glover (1977), and Bouwer (1978).

Recent papers that deal specifically with estimating ground-water recharge from streams include Gelhar and others (1979), Flug and others (1980), and Abdulrazzak and Morel-Seytoux (1983). Field studies of the interaction of lakes or sloughs and ground water were reported by Meyboom (1966) and Sloan (1972). Winter (1976) addressed the same topic based on numerical simulations of a steady-state system.

The mechanism of recharge by direct infiltration and ensuing downward movement to shallow aquifers is the dominant mechanism by which shallow glacial aquifers in North Dakota are recharged. It is, however, the mechanism for which the techniques of quantitative analysis are least defined or standardized. Most studies that have involved a quantitative analysis of water-table fluctuations have approached the recharge problem from the perspective of ground-water theory only. Freeze (1969) pointed out that such approaches ignore the influence of the unsaturated zone in transmitting infiltrating water to the water table, and suggested unsaturated-flow theory as the appropriate approach to studying natural ground-water recharge and discharge processes.

Other workers have tried a variety of approaches to develop an estimate of the amount of water that moves through the earth between the land surface and the water table, or some subset of that interval. The following section reviews most of the more common approaches that could have some application to a determination of ground-water recharge.

METHODS OF STUDYING RECHARGE

The movement of water through unsaturated natural media is a very complex process. If the media is within the biologically active zone (that is, the soil zone), the movement of water can be an extremely dynamic process. That process has been the subject of intensive study, but until recently the research was conducted almost exclusively by soil physicists. Because the investigative realm of the soil physicist normally ends at the base of the root zone, their literature pertains to "soil water" rather than water in the unsaturated zone. Therefore, it can become an acquired tendency for a researcher of unsaturated-flow processes to think in terms of flow phenomenon in soils rather than in the broader context of the unsaturated zone. In areas where very shallow water tables occur, such as in many of North Dakota's glacial aquifers, the two zones are nearly coincident. Thus, to maintain compatibility with most of the literature from which this report was drawn, the spatial reference for discussions of unsaturated flow usually will be the soil, but may alternate between the soil and the unsaturated zone, generally without regard to the depth involved. Where a distinction is important, the distinction will be made.

Water-Balance Techniques

Precipitation falling to the Earth's surface may flow overland to become streamflow, evaporate from the land surface or plant surfaces, or it may infiltrate the soil surface to be stored in the subsurface--perhaps later to be drawn from the soil by plant roots and transpired. If the pathways for precipitation are written as an equation, the equation would specify that the quantity of water falling on a given area of the Earth's surface may be accounted for by a summation of the quantities representing runoff, evapotranspiration, and water remaining in the subsurface. This equation is the basis for the water-balance approach to estimating ground-water recharge.

The water balance may be formulated rigorously as a combination of equations based on demonstrated physical principles or it can be represented very informally by equations that express simple concepts or empirical observations. The latter approach has been applied more commonly in hydrologic studies, mainly because the data requirements increase in proportion to the level of physical basis in the study approach.

The long-term water balance for the zone between the soil-atmosphere boundary and the water table may be written as

$$R = P - RO - ET, \tag{1}$$

where R = ground-water recharge,
P = precipitation,
RO = runoff, and
ET = evapotranspiration.

All quantities are expressed in units of length per unit time.

Each component in equation 1 should be represented by long-term (several years) average values. Seasonal changes in soil-moisture storage then may be ignored because the average of yearly soil-moisture changes will tend toward zero. Most studies involving a determination of ground-water recharge have approached the problem through some form of solution of this equation.

The simple appearance of the water-balance equation belies the difficulty in applying it credibly. The major drawback to the water-balance method is that errors in the known components, or the components that appear on the right-hand side of the equation, are accumulated in the unknown component, e.g., recharge. As pointed out by Kitching and others (1977), standard precipitation gages may err by up to 20 percent of the catch for any given precipitation event. Measurement error for snowfall, particularly if it is wind driven, may be much greater. Additional uncertainty in the water-balance equation arises because evapotranspiration estimates based only on meteorological data and crop coefficients are unverifiable without empirical data from lysimeters or soil-moisture studies. Unfortunately, the empirical data rarely are available. The runoff component of the water-balance equation generally can be determined within 10 percent if the study area boundaries coincide with drainage basin boundaries and the runoff is confined to a defined channel so that accurate discharge measurements can be made. But, if a defined drainage course does not exist within the study area, or if internal drainage to lakes or sloughs predominates, a reliable estimate of runoff can be problematic. Freeze and Cherry (1979, p. 207) noted, in fact, that the practical application of the water-balance equation is fraught with problems and pointed out some of the potential sources of error.

Water balances for short time periods, especially of less than 1 year, must account for changes in soil-moisture storage (ΔSM). The water-balance equation then becomes:

$$R = P - RO - ET \pm \Delta SM. \qquad (2)$$

Addition of the ΔSM term necessitates some method of continuously accounting for the varying amount of moisture stored in the soil profile.

The time-variant processes of soil-moisture storage and redistribution have been studied by soil physicists in the laboratory and in field-soil profiles for decades--as evidenced by the large amount of literature on the subject. Hydrologists have used at least the rudiments or approximations of the soil-moisture theory in combination with one of the various methods for estimating soil-water loss by evapotranspiration to develop water budget techniques for use in estimating ground-water recharge. A typical application of the budget technique is described by Meyboom (1967) in his pioneering analysis of ground-water recharge on the Canadian prairies. He used Thornthwaite's (1948) method of calculating potential evapotranspiration in conjunction with a simple soil-moisture budget technique developed by Holmes and Robertson (1959). Meyboom's work (also see Meyboom, 1966) indicated that recharge indeed does occur on the northern prairies.

Rushton and Ward (1979) stated that the conventional method of estimating ground-water recharge in Great Britain is based on a combination of the Penman (1948, 1949) potential evapotranspiration method and the Grindley (1967) method of estimating soil-moisture deficits. Rushton and Ward (1979) found that the Penman-Grindley method tends to underestimate the recharge, apparently because it wrongfully assumes that no recharge can occur when a soil-moisture deficit exists (that is, when the moisture content is less than the ill-defined field capacity). They also concluded that "Unless recharge calculations are made on a daily basis, a significant underestimate of the recharge may result." Spink and Rushton (1979) reinforced those conclusions by simulating the regional ground-water flow with a computer model. Howard and Lloyd (1979) examined the potential cumulative effect of errors in data used in the Penman (1950) evapotranspiration-soil-moisture balance method for the estimate of recharge and found that the method was very sensitive to variations in several of the evapotranspiration energy parameters. They emphasized the importance of using a daily water balance rather than a 10-day or monthly average water balance and questioned whether the method adequately represented the recharge mechanisms involved. They also admonished that "***the estimates will always be limited by the accuracy of the available input data."

The various water-balance schemes that have been used by hydrologists differ mainly in the level of detail or complexity incorporated in the determination of each balance component. One of the more detailed treatments was developed by Eagleson (1978). His one-dimensional, vertical water-balance model provides for the dynamic interaction of climate, vegetation, soil, and water through deterministic process equations, but uncertainty is incorporated by formulating the input variables in terms of probability density functions rather than the more conventional mean values. Although the ground-water and soil-moisture flow representation in the model is highly idealized and only approximately physically based, Caro and Eagleson (1981) determined that, with minor modification, the model provided "***a reasonable basis for estimating ground-water recharge***" due to rainfall for an arid zone in the Middle East.

An approximate treatment of the soil-water balance for large study areas, such as a drainage basin, generally is one of the functions of watershed or rainfall-runoff models. These models typically represent the soil or unsaturated zone as a series of storage reservoirs. An upper reservoir might represent the storage arising from small depressions at the land surface and interception by vegetation canopies. Other reservoirs could represent storage in various intervals of the soil profile, each with a given soil-water-holding capacity and a fractional contribution to the total evapotranspirational demand. Moisture accretions in excess of the available storage capacity of the top reservoir cascade to the second reservoir--and so on, down to the water table. Bear (1979, p. 38) described the soil-moisture accounting process in some detail for the two-layered Stanford Watershed Model. Crawford and Linsley (1966) and Larson and others (1982) detailed the operational characteristics, including the soil-moisture scheme, of three other frequently used general watershed models.

A large number of water-balance models have been developed by soil scientists to simulate the movement of water between the atmosphere-soil boundary or the top of the plant canopy and the base of the root zone. They offer a great variety of methods for treatment of soil-moisture storage and boundary conditions, including moisture uptake and dissipation by plants. Although these models are designed for application to agronomic problems and generally are not concerned with water that passes beyond the root zone to become deep percolation, many of them could be used for the determination of ground-water recharge as a component in a model with greater scope.

Even a cursory review of the published models would be a major undertaking and was not attempted as part of this study. De Jong (1981) provided a good review of many of the models and compared the types of approaches that have been used. Methods of simulating soil-water uptake by plant roots was emphasized particularly in De Jong's (1981) review.

Lysimetry

The ideal water balance would be one in which all, or all but one, of the components in the balance equation actually could be measured. This ideal could be achieved only in a closed system where all water entering and leaving the system could be controlled and monitored. The closed-system concept is the basis for open-topped boxes, called lysimeters, that have been used in many agronomic studies to measure the components of the near-surface water budget. The lysimeter is filled with soil representative of that in the study area and is placed in the ground so that its surface is flush with the surrounding land surface. Local crops or native vegetation usually are grown in the lysimeters in order to represent soil-water-plant relationships under prevailing land uses. Ground-water recharge may be determined as the quantity of rainfall or applied water that passes through the root zone and percolates to the base of the lysimeter. A variation of the box-type lysimeter is the monolythic lysimeter, which is an undisturbed block of soil that is enclosed only laterally--no floor is provided--in an impermeable barrier. Bouwer (1978, p. 266) described the typical lysimeter construction and operation and cited several studies involving lysimetry. Hillel (1980b, p. 202) listed 10 publications that document the use of lysimeters for measurement of the field water balance. Reichman and others (1979) described the construction and performance of 12 automated, non-weighing lysimeters used for crop production studies near Oakes, N.Dak. Kitching and Bridge (1974) described the construction and operation of two very large (33 feet square) monolythic lysimeters used for a recharge study in England.

The major advantage of the lysimetry method is that it provides a direct measurement of deep percolation that, in many settings, can be equated directly with ground-water recharge. Another advantage is the ability to account realistically for heterogeneous hydraulic phenomenon such as flow along root channels and desiccation cracks. The major disadvantage is the expense of installation and operation. Also, the lysimeter provides data for only one combination of soil and microclimatic conditions. If the study area is very heterogeneous, the lysimeter data may have limited transfer value. In addition, installation of most lysimeters requires disturbing the soil and subsoil so that the resulting infiltration may not be representative of natural infiltration.

Time-Series Analysis

Another technique that has been used to estimate ground-water recharge rates is the time-series model. This technique is a relatively simple statistical approach that relates variations in water-table levels at a given point. Eriksson (1970), for example, analyzed 30 years of ground-water-level variations using a first-order linear Markov process of the type

$$W_t = aw_{t-1} + U_t, \tag{3}$$

where W_t = the water level at time = t,

a = a constant whose value can be no greater than unity,

w_{t-1} = the water level at t minus one time lag (such as 1 day or 1 hour), and

U_t = a random independent variable that accounts for the stochastic nature of the water-level variations.

Rennolls and others (1980) used a first-order autoregressive model to describe the relationship between water-level variations in a well and a series of daily rainfall magnitudes. Because their model assumed no delay in the aquifer's response to rainfall, the total response of the water table to any rainfall event was assumed to occur on the day of the event. Viswanathan (1983), on the other hand, used a model that represented water levels in bore holes as a function of the previous day's water level, rainfall on the current day and 8 previous days, and a random variation factor. Unknown components in his model were fit by least squares regression analysis.

The major shortcoming of time-series models is their lack of physical basis. Equation variables and coefficients are fit to historical water levels for a particular well and could not be expected to have validity for locations significantly removed from the calibration site. Where historical data are lacking, the calibration of time-series models is impossible. Also, as pointed out by Freeze and Banner (1970), the method of evaluating ground-water recharge on the basis of observation-well hydrographs alone can be misleading because too many other factors that bear no relation to recharge can affect the water levels. Even if the water-level fluctuations can be attributed to recharge, estimates of the specific yield of the aquifer are needed to compute the recharge flux.

Mathematical Models

In recent years, the major emphasis in the study of hydrologic problems has been the use of computer models that solve mathematical expressions for the hydrologic processes of interest. The governing equations may be based on empirical observations or on accepted theoretical principles. It should be noted, however, that the distinction is somewhat vague because much of the hydrologic theory has some empirical foundation. The differential equation of ground-water flow is an expression of Darcy's law--an empirical equation. The important difference is that the theoretical principles have been tested under a multitude of hydrologic environments and have been proven generally applicable. On the other hand, purely empirical expressions are derived to compare observed causes with observed effects for a particular study area. Empirical equations are not universally applicable, so trial and error adjustments of the equation components almost always are required.

Computer models of hydrologic processes use numerical techniques to approximate solutions to the governing equations. Such models may be classified as empirically based or theoretically based (also commonly referred to as physics based) depending on the nature of the constituting equations. Empirically based models, due to the ordinarily liberal use of fitting coefficients, generally must be calibrated against observed phenomenon. Physics-based models use exact (or as exact as current knowledge will permit) formulations of the hydrologic processes and include few, if any, fitting coefficients. In concept then, the physics-based models may be used without calibration for a given study site. In practice, they nearly always are calibrated by adjusting values of the system properties (such as hydraulic conductivity) to compensate for the modeler's imperfect knowledge of the actual values.

Just as a distinction sometimes may be difficult between empirical and theoretical equations of hydrologic processes, so it is with hydrologic models. As noted by Woolhiser and Brakensiek (1982, p. 8), "Theory and empiricism are generally so intermeshed that in actuality most or all watershed hydrology models are hybrids that include both theoretical and empirical components." Their statement could be generalized to include all types of comprehensive hydrologic models.

As used in this report, theoretical or physics-based models will refer to those models that are based primarily on the differential equations that express continuity relationships for the various segments of the hydrologic cycle. Empirical or parametric models simulate water movement through the conversion of an input to an output by some transfer function. The transfer function generally is an empirical expression but also could be based on some physical principle (Lappala, 1980). Data requirements for the parametric models are not as stringent as those for the physics-based models and often can be derived from generally available data or by estimation. The parametric approach, therefore, is less data intensive than the physics-based approach. However, since the transfer functions usually must be adjusted by calibration to observed field data, generality of the models may be impaired. In fact, the lack of strong physical basis is their major weakness.

Many models of both types that could be used to study ground-water recharge have been developed in the last two decades. Freeze (1969) reviewed publications that provided rigorous numerical mathematical treatments of one-dimensional, vertical, unsteady, unsaturated flow in what generally were assumed to be homogeneous, isotropic soils. Although a variety of initial and boundary conditions were considered, the total potential gradient was assumed to consist only of the hydraulic gradient in each case. Nearly all of the papers involved assumptions of a semi-infinite medium (i.e., no lower boundary) or of a static water table. Neither assumed condition allowed for a realistic interaction of the unsaturated and saturated zones.

Since Freeze's (1969) review, there has been some unification of the theoretical treatments of saturated and unsaturated flow. Several transient models, the earliest of which were developed by Rubin (1968) and Freeze (1969), treat the movement of water in both saturated and unsaturated media so that flow may be simulated from the land surface to and beneath the water table. Taylor and Luthin (1969) and Cooley (1971) developed numeric models of the radial flow to pumping wells in unconfined aquifers that incorporated treatment of the unsaturated zone.

The most physically realistic of the models that simulate both saturated and unsaturated flow treat the entire subsurface as a continuous-flow domain, with hydraulic head as a continuous function of depth without an artificial boundary at the water table. Thus, from a value of zero at the water table, pressure head (total hydraulic head = pressure head + elevation head) becomes increasingly positive with depth in the aquifer and increasingly negative with distance toward the land surface. The three-dimensional model described by Freeze (1971) is one of several models that uses a single equation and the continuous-head concept to simulate the flow of water. Vauclin and others (1979) reported development of a two-dimensional, single-equation model of saturated-unsaturated flow which, unlike most of its predecessors, was tested and verified with a two-dimensional sandbox model.

Other investigators have distinguished between saturated and unsaturated flow processes and have imposed a mathematical discontinuity (i.e., a flow boundary) at the water table. One such model is that of Hornberger and Remson (1970). Another group expressed preference for the unified or continuous-flow concept but, for reasons of computational efficiency, developed subsurface flow models by coupling a one-dimensional form of the Richards equation to some form of the unconfined ground-water equation (Pikul and others, 1974; Laat, 1980; Gilding, 1983). Models of this type are purported to yield satisfactory results when used within their prescribed limitations and are computationally efficient enough to simulate subsurface flow for basin-size study areas.

Many more specialized models have been written, each describing the flow of water or energy through unsaturated or variably saturated media under a particular set of assumed conditions. Lappala (1980), for example, reviewed more than 250 reports dealing with solutions to the equations for transport of mass and energy in variably saturated media. Unfortunately, a trait common to almost all of the models with strong physical basis is that field verification is lacking. Consequently, very few have been put to any practical use.

Because parametric models require fewer or more generally available data than the physics-based models commonly used for research projects, they frequently are chosen for production use--such as guiding the water-related management decisions in agricultural operations. Skaggs (1978) described a model consisting of several components that evaluates, each by approximate methods, infiltration, surface and subsurface drainage, potential and actual evapotranspiration, subirrigation, and soil-water redistribution. The model was designed for compatibility with readily available data, a design criteria for most parametric models. Austin and others (1982) briefly described the components of a model with similar capabilities that was designed for use in improving irrigation efficiencies.

King and Lambert (1976) designed a parametric model to simulate the movement of precipitation to deep water tables. Their model included treatment of the processes of surface runoff, surface storage, potential and actual evapotranspiration, soil-water redistribution, and water-table accretion and ablation. Redistribution was simulated by transferring water from layer to layer according to Darcy's law. Knoch and others (1983) developed a one-dimensional model to predict ground-water recharge from precipitation or irrigation on a daily basis. Their model differs from that of King and Lambert (1976) mainly in that it includes a simple routine to treat snowmelt over frozen ground.

In conclusion, little doubt remains as to the value of computer models for simulating real hydrologic processes. It has been demonstrated that physics-based models may be made to essentially duplicate the hydraulic behavior of fairly simple systems. Model performance diminishes with increasing complexity of the system being modeled, but results sufficiently accurate for the intended purpose often can be achieved if the physical characteristics of the prototype are known in relative detail. Thus, the availability of data generally is the key to successful modeling of hydrologic processes. Models are available for application to most practical problems, but the data required for a minimal characterization of the area under study often are lacking. This is particularly true for problems involving flow in unsaturated media.

THEORY OF FLOW IN THE UNSATURATED ZONE

Physical Properties

Three physical properties of soil must be known to enable a quantitative treatment of soil-water movement--soil-moisture content, soil-moisture potential, and hydraulic conductivity. The volume of water contained in the pores of a unit volume of soil is known as the soil-water content or soil-moisture content or, as preferred by Hillel (1980a), simply, wetness. The soil-water content also may be expressed as a mass ratio, but the volume basis generally is used in the analysis of soil-water movement. Both fractional quantities are dimensionless and usually are multiplied by 100 and reported as percentages.

Soil-moisture potential is a measure of the total energy status of soil water at a particular point in space. The standard expression for potential is energy per unit mass, having units of $(\text{length})^2/(\text{time})^2$. It usually is more convenient, however, to express the potential as energy per unit weight of water--which is the familiar quantity hydraulic head and has the units of length. The ratio of the difference in potential between two points in the soil matrix to the distance between the points is the potential gradient. A finite gradient is a prerequisite to soil-water movement.

The soil-moisture potential has three components: pressure potential, gravitational potential, and osmotic potential. The pressure potential is assigned a positive value if it is greater than atmospheric pressure and a negative value if it is less. Negative pressure potentials generally are referred to as capillary or matric potentials in the soil-science literature. In reporting actual values of matric potential, the terms soil suction or soil tension normally are used so that the negative sign may be omitted. Therefore, soil suction or tension increases (becomes more positive) as matric potential decreases (becomes more negative).

Matric potentials result from the affinity pore water has for soil particles, by adsorption, and by capillary forces. The drier the soil, the greater the tenacity with which it holds a given unit of water. Thus, soil suction and soil wetness are inversely proportional. At the low end of the field-soil wetness scale, the soil suction may be many atmospheres, whereas the suction at saturation is zero. A graphic representation of the relationship of soil wetness and soil suction is called the soil-moisture characteristic (Childs, 1940). Due to a phenomenon called hysteresis, the correspondence between wetness and suction during the wetting of a soil is different than it is during the drying process. Hysteresis and its implications for soil-moisture flow are described in a subsequent section of this report.

The gravitational potential results from a difference in elevation between some specified reference elevation and the point in question. It represents the potential energy that a unit of water possesses due to its elevation in the soil column and must be accounted for in comparing total potentials for points not on the same elevational plane.

The osmotic potential arises from the presence of solutes in the soil pore water, a condition that lowers the potential energy of water in the solution relative to that of pure water at the same pressure and gravitational potential. The osmotic potential, however, is manifest in soil-water flow behavior only if a solute concentration gradient exists across the equivalent of a semipermeable or selective membrane. Water then may move by osmosis through the membrane, from the dilute to the concentrated solution, to raise the total soil-water potential of the concentrated solution. Because normal field soils, even those with restrictive clay layers, emulate selective membranes only poorly at best, the osmotic effect on liquid water movement normally is overwhelmed by the matric and gravitational potential. Thus, the osmotic potential generally may be disregarded, especially for clay-free soils (Hillel, 1980a, p. 247).

The third essential property for the analysis of soil-water movement is the hydraulic conductivity. Unlike its invariant nature at a given point below the water table, hydraulic conductivity in the unsaturated zone varies with the wetness. Because liquid water may flow only in the hydrated layer around solid particles in the soil matrix, unsaturated hydraulic conductivity is directly proportional to wetness and is very sensitive to even minor changes in wetness. Hydraulic conductivity, in fact, usually varies over several orders of magnitude through the normal range of wetness for a given field soil (Amerman, 1973, p. 174; Hillel, 1980a, p. 197). Due to the inverse relationship of wetness and soil suction, it follows that hydraulic conductivity and suction also must be inversely related.

The relationship of hydraulic conductivity to suction may be visualized most readily by reference to the classic capillary tube model (see, for example, Hillel, 1980a), whereby the soil pore space is idealized as a collection of variously sized capillaries. Experiments show that a water-filled capillary of radius r will not drain until the suction (ψ) acting on it attains a value given by

$$\psi = \frac{2\delta \cos\alpha}{r\rho g}, \tag{4}$$

where δ = surface tension between liquid and air (mass/time^2),
α = contact angle between liquid and capillary wall (degrees),
r = radius of the capillary (length),
ρ = liquid density (mass/length^3), and
g = gravitational acceleration (length/time^2).

The equation provides that the large capillaries or, by the analogy, the large soil pores are the first to drain as the suction increases from an initially low value. Not only do the large pores allow the most efficient movement of soil water, but their loss (by drainage) requires that subsequent water movement take a much more tortuous route to avoid the large air-filled pores. Thus, unsaturated hydraulic conductivity decreases with increasing suction--quite abruptly in uniform-sized soils but more gradually in well-graded ones.

Equations for Unsaturated Flow

Various aspects of the physics of fluid movement in unsaturated or variably saturated media are described in several reference works including Miller and Klute (1967), Stallman (1967), Childs (1969), Philip (1969, 1970), Hillel (1971, 1980a, 1980b), Bear (1972; 1979, chapter 6), Nielsen and others (1972), Bruce and others (1973), Morel-Seytoux (1973), Corey (1977), Freeze and Cherry (1979, chapter 2), and Amerman and Naney (1982). Rather than attempt to duplicate the exhaustive coverage of these references, the following treatment will abstract those facets of the theory that directly apply to the study of ground-water recharge.

The equations used to solve practical problems involving flow through porous media employ the concept of a macroscopic continuum. The concept tacitly acknowledges the futility of approaching field-scale problems from the microscopic perspective by using equations that describe the flow of water through individual soil pores. The macroscopic continuum approach assumes that the physical properties of the soil and water are continuous functions of space and time and that flow behavior may be represented adequately as an average of the fluid flow process that actually occurs in a labyrinth of individual soil pores. The matter of an appropriate volume over which to average the relevant soil and water properties has received considerable attention in recent literature, but a concensus of opinion still is lacking. Bear (1972, p. 28) discussed the continuum approach at length and provided a mathematical definition for what he called the representative elementary volume.

The movement of water through porous media may occur by any of several different mechanisms. Klute (1973) listed the mechanisms for nonrigid media as (1) bulk, convective transport in either the liquid or the gas phase, or both; (2) diffusion and dispersion in the liquid or the gas phase, or both; and (3) convective transport of the water in the liquid or the gas phase, or both, due to the motion of the solid phase. If the media is rigid, as generally is assumed for field flow problems, the third mechanism may be ignored. In fact, the bulk of the literature concerning unsaturated-flow phenomenon considers only the liquid phase and further assumes that the hydraulic gradient is the sole gradient causing flow (i.e., convective transport).

As for the movement of water in the gas phase, convective transport may occur in soil very near the surface in response to wind gusts or large, abrupt changes in atmospheric pressure. Such phenomena are effective only at shallow depths, however. Vapor movement by diffusion, on the other hand, is induced by vapor pressure gradients. Under isothermal conditions and in the absence of osmotic or solute effects, vapor pressure gradients generated by suction variations within the normal range for field soils are not likely to result in appreciable water movement relative to that which occurs in the liquid phase (Hillel, 1980a). Temperature induced vapor gradients, however, can be appreciable in the near-surface soils during periods of wide diurnal variation of air temperatures. Nonetheless, at least under the wetter moisture conditions conducive to vigorous plant growth, it generally is assumed that water movement in the vapor phase is negligible relative to that in the liquid phase.

The general equation governing the flow of water in unsaturated porous media is obtained by combining an expression for the conservation of mass of water flowing through a volume element of soil with the expression relating mass flux of water to the potential gradient causing flow (i.e., Darcy's law). The continuity equation, which is derived in some form in almost any text dealing with hydrology, soil physics, or fluid mechanics, may be written for the vertical dimension as

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z}, \tag{5}$$

where θ = volumetric wetness,
q = the Darcy flux,
t = time, and
z = the vertical coordinate, positive upward.

The flux equation (Darcy's law) is

$$q = -K(\psi)\frac{\partial H}{\partial z}, \tag{6}$$

where $K(\psi)$ = the hydraulic conductivity (a function of suction) and
H = the total hydraulic potential.

The minus sign indicates that the flow is in the direction of decreasing potential. Because $H = -\psi + z$, equation 6 may be written

$$q = -K(\psi)\frac{\partial}{\partial z}(-\psi + z) = K(\psi)\frac{\partial\psi}{\psi z} - K(\psi). \tag{7}$$

Combining equations 5 and 7 yields the equation for unsteady vertical flow:

$$\frac{\partial\theta}{\partial t} = -\frac{\partial}{\partial z}\left[K(\psi)\frac{\partial\psi}{\partial z}\right] + \frac{\partial K(\psi)}{\partial z}. \tag{8}$$

Extension of equation 8 to three dimensions yields the following equation:

$$\frac{\partial\theta}{\partial t} = -\frac{\partial}{\partial x}\left[K(\psi)\frac{\partial\psi}{\partial x}\right] - \frac{\partial}{\partial y}\left[K(\psi)\frac{\partial\psi}{\partial y}\right] - \frac{\partial}{\partial z}\left[K(\psi)\frac{\partial\psi}{\partial z}\right] + \frac{\partial K(\psi)}{\partial z}. \tag{9}$$

The $\frac{\partial\theta}{\partial t}$ term commonly is expanded by the chain rule to $\frac{\partial\theta}{\partial\psi}\frac{\partial\psi}{\partial t}$, where $\frac{\partial\theta}{\partial\psi}$ is defined as the soil water capacity, C. Thus, if $\frac{\partial\theta}{\partial t}$ is represented by $C\frac{\partial\psi}{\partial t}$ and K and C are expressed as functions of ψ, then ψ is the only dependent variable in equations 8 and 9. Richards (1931) apparently was the first to combine the Darcy equation with the equation of continuity for application to unsteady, unsaturated flow problems. Equation 9, with the $\frac{\partial\theta}{\partial t} = C\frac{\partial\psi}{\partial t}$ substitution, essentially is the same as Richards' (1931) equation 15. In recognition of his pioneering work, most formulations of the one-, two-, or three-dimensional unsteady, unsaturated flow equation are referred to simply as the Richards equation.

Another adaptation of equation 8 or 9 has been used to facilitate an analytic solution. The suction gradient may be expanded by the chain rule to

$$\frac{\partial\psi}{\partial z} = \frac{\partial\psi}{\partial\theta}\frac{\partial\theta}{\partial z},$$

so that the Darcy flux term becomes

$$q = -D(\theta)\frac{\partial\theta}{\partial z},$$

where the hydraulic diffusivity, $D(\theta)$, is defined as

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

and both D and K are functions of wetness (θ). Equation 8 then may be written

$$\frac{\partial\theta}{\partial t} = -\frac{\partial}{\partial z}\left[D(\theta)\frac{\partial\theta}{\partial z}\right] + \frac{\partial K(\theta)}{\partial z}. \tag{10}$$

As stated by Hillel (1980a, p. 205), equation 10 is in a form analogous to the equations of diffusion and heat conduction, and solutions have been documented for a wide range of boundary conditions.

Determination of the Physical Properties

The main problem in applying physics-based theory to the study of unsaturated flow is defining the functional relations of matric potential and hydraulic conductivity to moisture content or wetness. Numerous papers in the soil-physics literature have described methods for the laboratory determination of the moisture-characteristic, hydraulic-conductivity, and diffusivity relations, but the methods are tedious and costly and are not applied often to actual field-scale soils problems, much less to hydrologic problems. Reviews of available methods, or of papers that originally documented the methods, are given by Gardner (1965), Klute (1965a, 1965b, 1973), Richards (1965), Childs (1969), and Hillel (1980a). Because these reference works are available, no further mention will be made in this report of the laboratory methods for physical characterization.

Much effort has been expended in recent decades to develop theoretical and empirical models that would diminish or even eliminate the need for the cumbersome laboratory methods. Another impetus for such models has been the need to define simple functional relationships among the variables in the Richards equation to facilitate its' analytic solution. The latter objective probably is not as critical now as it was before the development of computer-assisted numerical techniques for solution of differential equations.

Soil Moisture

Simple power functions relating moisture content to suction were used by Brooks and Corey (1964, p. 4), Gardner and others (1970), Campbell (1974), and Clapp and Hornberger (1978) to approximate the moisture-characteristic curve. Equations developed in each of the four papers are empirical, and most are of the form $\psi = a\theta^{-b}$, where a and b are constants that must be determined by empirical fitting for each locality where the equation is to be applied. The need for empirical evaluation of the equation components is the major deterrent to use of these equations. White and others (1970) developed equations relating moisture content to suction based on a physical interpretation involving the pore geometry, but their equations also include components that must be evaluated experimentally.

Another approach has been to fit regression equations to data relating moisture content at a prescribed suction to readily measurable soil properties, such as percent sand, silt, and clay, percent organic matter, and bulk density. Gupta and Larson (1979) derived equations for moisture content at each of 12 different suctions ranging from 0.04 to 15.0 bars to approximate the moisture characteristic for several different soils. Rawls and others (1982) developed a similar equation using data from more than 2,500 soil horizons. They were able to improve the regression fit by adding the 1/3- and 15-bar water-retention values as regression parameters. Although the regression method still requires enough suction and wetness data to fit a least-squares line, published results such as those of Rawls and others (1982) represent a very broad cross section of soil types and may offer a convenient means of approximating the moisture characteristic for a given soil with a minimum of data collection.

McQueen and Miller (1974) briefly reviewed several other papers that dealt with the approximation of the soil-moisture characteristic from limited data, and Hillel (1980a, p. 149) listed several papers that have proposed equations for the same purpose. Hillel noted, however, that "As yet, no satisfactory theory exists for the prediction of the matric suction versus wetness relationship from basic soil properties. The adsorption and pore-geometry effects are often too complex to be described by a simple model. Several empirical equations have been proposed which apparently describe the soil-moisture characteristic for some soils and within limited suction ranges."

Hydraulic Conductivity

Estimation models

Alternatives to the traditional laboratory methods for determining the unsaturated hydraulic-conductivity functions $K(\psi)$ and $K(\theta)$ have been intensively researched. According to Mualem (1976), most models developed to predict unsaturated hydraulic conductivity use one of two general approaches. The first approach assumes that K may be expressed as a power function of a relative moisture-content index. Probably the best known of these models was published by Brooks and Corey (1964):

$$K_r = \left[\frac{S - S_r}{1 - S_r}\right]^{\alpha}, \tag{11}$$

where K_r = relative hydraulic conductivity = $\frac{K}{K_{sat}}$,
K_{sat} = saturated hydraulic conductivity,
S = saturation = (volume of water)/(volume of interconnected pore space),
S_r = residual saturation,
α = an experimentally determined coefficient.

The $\frac{S - S_r}{1 - S_r}$ term may be thought of as the active or participating part of the soil-moisture reservoir, or the amount of moisture in excess of that part that is immobile because of great attractive forces between it and the soil particles.

The other general approach relies on an idealization of the soil pore space as a collection of parallel capillary tubes. Various models have been developed around the capillary tube concept (Brutsaert, 1967). Probably the best known is the model of Childs and Collis-George (1950) and subsequent modifications by Marshall (1958) and by Millington and Quirk (1959). The soil pore space is represented by groupings of parallel, variously sized, randomly spaced capillary tubes that are randomly joined, end-to-end, with other such groupings. Dead-ended (no connection with other tubes at the adjoined face) tubes are considered nonconducting. Poiseuille's law and the equation of capillarity (equation 4 in this report) are used to derive an expression for hydraulic conductivity. A measured moisture-characteristic curve is used to estimate the volume of pore space associated with each pore-radius increment. A summation of the discharge through each of the size classes of pores that remain water filled at a prescribed suction yields the overall conductance (which is converted to hydraulic conductivity) for the media at that suction and corresponding wetness. Hillel (1980a, p. 209) and Brutsaert (1967) described the concepts of the capillary models in considerable detail.

Tests of various hydraulic-conductivity models using experimental data were made by Nielsen and others (1960), Jackson and others (1965), Kunze and others (1968), Green and Corey (1971), Bruce (1972), and Jackson (1972). A recent paper by Gureghian and others (1982) reviewed four different models that have been developed to calculate unsaturated hydraulic conductivity, and compared the performance of each model against the measured hydraulic conductivity for two different soil types. No concensus is evident in the cited literature as to which of the models is best for general usage. Apparently, each of the models has been used to advantage to derive hydraulic-conductivity values for specific types of soils within common ranges of moisture content.

In situ methods

As indicated earlier, the interest in models used to estimate the moisture-characteristic and hydraulic-conductivity functions arose primarily because of the cumbersome or tedious nature of the laboratory methods for their determination. The development of field techniques for the measurement of soil hydraulic properties was spurred, in part, for a similar reason. An equally important impetus, however, was the need for hydraulic data obtained by measurements taken *in situ* on soils undisturbed by coring or other extraction procedures used to obtain samples for laboratory analysis. In recent years, many workers have espoused the merits of *in situ* measurements over laboratory methods. In fact, there seems to be some evidence in the literature that the laboratory methods, especially for hydraulic conductivity, are falling from favor. Hillel's (1980a) comments are typical of the concerns expressed by many soil scientists and hydrologists--"***it seems basically unrealistic to try to measure the unsaturated hydraulic conductivity of field soil by making laboratory determinations on discrete and small samples removed from their natural continuum, particularly when such samples are fragmented or otherwise disturbed. Hence, it is necessary to devise and test practical methods for measuring soil hydraulic conductivity on a macroscale *in situ*."

The problems inherent to laboratory determinations include the unavoidable disturbance of the soil sample during extraction, transport, and testing, and the inability of the small, discrete samples to account for field-scale heterogeneity. Because soil properties such as hydraulic conductivity, porosity, bulk density, and pore-size distribution vary in space, the values determined experimentally at any location will vary according to the volume of soil sampled. A large sample should be more representative of the average conditions of a field soil than a small sample. Therefore, it may be assumed that *in situ* measurements, because they may be designed to sample much greater volumes of soil than is feasible by laboratory methods, more closely approximate the average conditions of a given expanse of heterogeneous soil.

Another important advantage of *in situ* methods is that experimental plots may be established for regular, frequent data collection by nondestructive techniques. The same soil mass may be monitored indefinitely without loss of structural integrity.

The earliest studies designed to determine unsaturated hydraulic conductivity by *in situ* methods apparently were made by Richards and others (1956) and Ogata and Richards (1957). They measured matric potentials by tensiometry and moisture content by laboratory gravimetric methods. Nielsen and others (1964) measured matric potentials by tensiometry and moisture content with the neutron meter as well as by gravimetric methods. They concluded that the *in situ* hydraulic-conductivity technique was preferrable to laboratory methods because of the greater sampling volume and the ability to test undisturbed soil horizons in their natural position. Rose and others (1965) used *in situ* moisture-content data (from the neutron meter) in combination with matric potentials derived from laboratory-measured moisture-characteristic curves to calculate hydraulic conductivity. Brust and others (1968) admonished that the reliability of Rose's procedure depends upon the accuracy with which the field moisture-characteristic curve can be reproduced in the laboratory, and Klute (1973, p. 29) pointed out that "It seems *a priori* better to measure the hydraulic gradient [*in situ*] rather than infer it from a water retention curve." Klute (1973, p. 29) also noted: "The choice at the present time seems to be whether it is better to take a few samples to the laboratory and make measurements on them under more controlled conditions, or to make measurements *in situ* under less controlled conditions. The laboratory measurements may give more accurate and precise results on the samples as they exist in the laboratory than the field methods. However, the field methods are likely to give results that are more representative of the soil as it exists in the field. Measurements in the laboratory on 'undisturbed' core samples suffer from the adverse sampling situation imposed by the small size and number of the core samples that can be processed, and the possible and probable changes in soil properties that can occur when a sample is removed from the field to the laboratory. For such reasons, *in situ* measurement of the properties would seem to offer promise as a means of hydraulic characterization of a field site."

Watson (1966) demonstrated formally, by laboratory column experiments, the validity of Darcy's law for unsteady, unsaturated flow through fine sand at low suction. The technique, which he called the instantaneous profile method, involved the simultaneous determination of moisture content and matric potential at several column elevations at various times during drainage of the column from an initially saturated condition. Plots of moisture content (θ) versus time (t) for the various depth intervals and of total soil-water potential (*H*) versus depth (*z*) at various times may be constructed from the data. From these two plots, values for soil-moisture flux and hydraulic conductivity may be derived. The hydraulic conductivity may be defined as a function of either moisture content or suction.

A study reported by Van Bavel, Brust, and Stirk (1968) and Van Bavel, Stirk, and Brust (1968) apparently was the first to use Watson's instantaneous profile method entirely in the field. Hillel and others (1972) later gave a detailed, step-by-step description of a field version of Watson's method because they surmised that "***the reason why a measurement which is so important and useful has not yet been accepted and applied as widely as it should be***" was its inadequate exposure and clarification. Watson's method has been used since, reportedly with good success, by many investigators (Nielson and others, 1973; Stone and others, 1973; Jeppson and others, 1975; Rice, 1975; Stoner, 1983).

Attempts have been made to obtain satisfactory estimates of the hydraulic-conductivity function while simplifying the data requirements of the instantaneous profile method. Libardi and others (1980) are among several workers that have implemented the observations of Black and others (1969) and Davidson and others (1969) that the hydraulic gradient during drainage from uniform, wetted soil profiles is often unity--that is, the suction gradient is negligible. The need for suction data is, therefore, eliminated. Ahuja and others (1980), on the other hand, collected *in situ* suction data, but eliminated the need to measure moisture content by assuming it could be represented adequately as a simple function of suction.

Other field methods for the determination of hydraulic conductivity include a uniform sprinkling technique (Youngs, 1964) and a method involving infiltration through an impeding layer (Hillel and Gardner, 1970; Bouma and others, 1971; Hillel, 1980a). Neither method, however, is as generally applicable or as readily accomplished as the instantaneous profile method.

Diffusivity

An evaluation of the diffusivity $\left[K(\theta)\frac{\partial\psi}{\partial\theta}\right]$ term is of interest mainly to those seeking analytic solutions to the moisture content-diffusivity form of the flow equation. Diffusivity may be determined simply as the product of $K(\theta)$ and $\frac{\partial\psi}{\partial\theta}$ if the hydraulic conductivity and suction versus moisture-content relations are known or it may be determined by solving the flow equation explicitly for diffusivity and experimentally measuring the other terms in the equality. Methods for the determination of diffusivity are described by Doering (1965), Klute (1965b), and Gardner (1970).

Surface Boundary Conditions

The exchange of water across the surface boundary of the soil or unsaturated zone is of vital importance to the maintainence of agricultural productivity and to the replenishment of ground-water supplies. This exchange also is a major determinant of the magnitude and duration of streamflows for virtually all watersheds. It is not surprising then that the processes of infiltration and evapotranspiration are of interest to those in a wide range of scientific and engineering disciplines and that much research has been devoted to developing an understanding of these processes. Some of the fundamentals of that accumulated understanding that might apply to ground-water recharge are discussed in the following pages.

Infiltration

Infiltration is the process by which water flows from the soil surface into the soil profile, following a tortuous route within the hydration envelope surrounding individual soil particles that make up the solid matrix. Study of the three-phase infiltration process has been a prominent issue in soil physics for more than 70 years (Fleming and Smiles, 1975). Hydrologists also have sought a practical means of predicting infiltration for use in partitioning the water balance between runoff and subsurface storage accretion.

The physical principles involved in the infiltration process are well understood and have been reviewed comprehensively by Childs (1969), Philip (1969), Hillel (1971, 1980b), Nielsen and others (1972), Morel-Seytoux (1973), and Skaggs and Khaleel (1982). Fleming and Smiles (1975) considered some applications of the infiltration theory to hydrology and identified some associated problems that as yet are unresolved.

Infiltration is the initial phase of the soil-moisture flow process and is governed by the same physical principles and soil properties as the entire flow process. It differs from normal redistribution in that flow is induced by extreme gradients in the matric potential. These gradients result from the establishment of a thin layer of saturation at the soil surface at the outset of infiltration, overlying what may be very dry near-surface soil if appreciable time has passed since the last wetting. Infiltration generally is characterized by either (or both) of two factors--infiltration capacity (or infiltrability as suggested by Hillel, 1971) and cumulative infiltration. The former is the rate at which a soil is capable of transmitting water into the profile, assuming the water supply is not limiting. The latter term is the total amount of water passing into the profile since the initiation of infiltration.

Infiltrability is greatest at the start of rainfall or irrigation and diminishes with time. For early time, infiltrability is proportional to $1/\sqrt{t}$. Cumulative infiltration is proportional to $\sqrt{t}$ because it is the time integral of infiltrability. With the passage of time, infiltrability asymptotically approaches a constant value and cumulative infiltration approaches a constant rate of increase. These relationships are expressed in the physically based equations describing the infiltration process.

Infiltration may occur with or without water ponded on the surface. If the water application rate does not exceed the capacity of the soil to accept the water, ponding will not occur and the infiltration rate will be numerically equal to the application rate. As the near-surface soil profile progressively becomes wetter, the hydraulic gradient inducing flow into the soil lessens and the infiltration rate diminishes. If the application of water continues long enough and at a sufficiently large rate, ponding eventually will occur. From that time forward, infiltration will proceed at a rate limited by the permeability of the soil and not by the application rate.

A surface boundary condition must be specified to enable a solution of the Richards equation. The previous discussion indicates that infiltration may be provided for either by specifying a boundary head of positive matric potential (indicative of ponded water) or by specifying a head gradient at the boundary (which, in effect, fixes the boundary flux) such that the resulting boundary flux is equal to the application rate. The determination of infiltration rates, therefore, requires knowledge of the water application rates, the hydraulic properties of the soil, and the antecedent soil-moisture content. Given this information, it is possible to calculate the time to ponding for any precipitation or irrigation event, thereby signaling the conversion of the surface boundary condition from a prescribed flux to a prescribed head.

Much of our knowledge of the infiltration process has resulted from analytic solutions to the one-dimensional Richards equation for problems having simple initial and boundary conditions (see the review by Philip, 1969). Although the analytic techniques offer a rapid means of analyzing simple, idealized flow problems, digital models generally are required for the analysis of real flow systems involving nonuniform initial conditions and boundary conditions that vary in space and time. Models that have multidimensional capability are useful particularly for small-scale problems involving infiltration and soil-moisture redistribution in the vicinity of pits, ditches, mounds, and drains. One-dimensional analysis of infiltration often is adequate for large-scale problems involving watersheds and areas underlain by extensive aquifers.

Many empirical models of the infiltration process were developed prior to the time when high-speed computers and efficient numerical methods became widely available. Other infiltration models represent approximations to the physical theory. The references listed at the beginning of this section provide reviews of the various infiltration models in much more detail than is possible here.

The approximate model originally proposed by Green and Ampt (1911) is the most widely used infiltration model in hydrologic studies. Assumptions involved in the equation derivation included a ponded surface (that is, a constant, unlimited water supply) and a deep homogeneous soil profile with initially uniform water content. It was further assumed that moisture moved through the profile as an intact slug and a sharp wetting front separated the uniformly wetted zone from the as-yet unwetted zone ahead of the front. The hydraulic conductivity was assumed to be constant in the uniformly wetted zone and the suction at the wetting front also was assumed to be constant, regardless of time or position of the front. Using these assumptions, the Darcy equation could be written to represent the infiltration flux through the wetted zone:

$$inf = K_w\left[\frac{H_p + S_f + L_f}{L_f}\right], \tag{12}$$

where inf = infiltrability (length/time),
K_w = hydraulic conductivity of the wetted zone (length/time),
H_p = head due to ponding (length),
S_f = suction at the wetting front (length), and
L_f = length of the wetted zone (length).

As indicated by Skaggs and Khaleel (1982), Philip (1954) showed that K_w may be some value less than the saturated hydraulic conductivity because the wetted zone need not necessarily be saturated.

Cumulative infiltration, I, should equal the product of the increased moisture content in the wetted zone ($\theta_w - \theta_i$, where θ_w is the wetted zone moisture content and θ_i is the nonwetted zone moisture content) and the wetting front depth, L_f (recall that θ_w and θ_i are assumed to be uniform with depth). Thus,

$$I = (\theta_w - \theta_i)L_f$$

or

$$L_f = \frac{I}{\theta_w - \theta_i}. \tag{13}$$

Assuming that $H_p \simeq 0$, equations 12 and 13 may be combined to yield

$$inf = K_w + \frac{K_w S_f(\theta_w - \theta_i)}{I}. \tag{14}$$

With the substitution $inf = \frac{dI}{dt}$, equation 14 may be integrated and rearranged to give

$$I = K_w t + S_f(\theta_w - \theta_i)\,ln\left[1 + \frac{I}{S_f(\theta_w - \theta_i)}\right]. \tag{15}$$

As suggested by Hillel (1980b, p. 15), the time rate of change of the second term on the right side of equation 15 becomes inconsequential relative to that of $K_w t$ at large times, so that the equation eventually may be approximated by $I \simeq K_w t$ + constant.

Skaggs and Khaleel (1982) listed several studies in which the Green-Ampt model was used with satisfactory results for profile conditions other than those assumed in the original derivation. They also described some of the methods that have been used to estimate the Green-Ampt components (K_w, H_f, and $\theta_w - \theta_i$). Most workers have used some type of manipulation of the moisture-characteristic and hydraulic-conductivity functions to derive this estimate. Recent papers by McCuen and others (1981) and Rawls and others

(1983) drew on much of that earlier work to develop regression equations relating soil-texture indices to each of the Green-Ampt components.

The availability of computer models that solve the unsaturated-flow equation has eliminated the need to rely on the common estimation methods for the determination of infiltration rates and volumes. Most rainfall-runoff models, however, do not include provisions for the solution of the unsaturated-flow equation and continue to use the estimation techniques. Because nonuniform and intermittent boundary conditions may be simulated readily with the unsaturated-flow models, treatment of the infiltration process may be accomplished by specifying the rainfall (and irrigation, if applicable) timing and intensity. As long as the rainfall intensity does not exceed the infiltration capacity, the boundary condition is a specified flux equal to the application rate. If at any time during the simulation the application rate exceeds the infiltration capacity, ponding occurs and the boundary condition is converted to a specified potential that depends on the depth of ponding allowed. Some models (Lappala and others, 1987, for example) can be programmed to calculate the time of ponding so that the boundary conversion is done independent of user control.

Evapotranspiration

The economic importance and scientific interest attributed to the evapotranspiration process has resulted in a very large body of literature devoted to the topic. Brutsaert (1982) included passages in his book on evaporation that indicate that the rudimentary theories of evaporation were recorded in Greek antiquity. Research within the last several decades has firmly established the interrelationship of processes involving moisture movement within the soil-plant-atmosphere system and has developed theories on the physics of those processes. Much remains to be learned, however, particularly in regard to the quantitative treatment of the exchange of water between plants and the soil-moisture reservoir and atmosphere.

An extensive treatment of the principles of evapotranspiration and the methods available for measuring it are beyond the scope of this study, but many such treatises are available. Although most are written from an agricultural perspective, the physical principles governing the evapotranspiration process are applicable universally. A partial list of these reference works includes Jensen (1974), Burman and others (1980), Hillel (1980b), Brutsaert (1982), and Saxton and McGuinness (1982).

Evapotranspiration is a collective term that includes evaporation from bare soils and transpiration from plant tissues. It is a major component of the soil-water balance for all vegetated soils and for all but the coarsest of bare soils. Evapotranspiration can affect the magnitude of ground-water recharge in either or both of two ways. In areas where the water table is relatively deep, evapotranspiration diminishes the amount of water stored in the soil-moisture reservoir and decreases the likelihood that ensuing precipitation events will contribute to ground-water recharge. Where the water table is shallow (less than 10 to 20 feet), evapotranspiration may draw water directly from the water table. The latter situation may not affect the amount of water stored in the soil profile, particularly if the depth to water is no more than a few feet.

Three conditions must be satisfied for evapotranspiration to occur. First, there must be an energy supply to meet the latent heat requirement. About 590 calories are required to evaporate 1 gram of water at 60 °F. The average pan evaporation of 0.29 inches for a July day at Fargo, N.Dak. (Farnsworth and Thompson, 1982), requires the energy equivalent of about 20,400 kilowatt hours per acre. Second, there must be water available at the soil surface or within the profile. Third, a vapor pressure gradient must be maintained between the plant or soil surface and the atmosphere--that is, there must be a vapor sink to accept the evaporating water and allow the evapotranspiration process to continue. The processes of advection, convection, and diffusion transport the vapor away from the evaporating surfaces so that the surrounding air generally does not become vapor saturated.

During the wet conditions typical of springtime, the availability of energy is the limiting factor for evapotranspiration. For much of the rest of the growing season, particularly in nonirrigated arid or semiarid regions, the availability of water limits the evapotranspiration process. At such times, the permeability of the progressively drying soil controls the amount of water that is made available for evapotranspiration at the soil surface. Incoming radiant energy that is not consumed in the evaporation process is dissipated in warming the soil, the vegetative matter, and the air. Evapotranspiration seldom is limited due to the lack of a vapor pressure gradient at the evaporating surfaces.

The only direct method of measuring evapotranspiration is to maintain an accurate account of water inflow and losses using lysimeters. This method is impractical for most hydrologic studies because of the expense involved and the fact that the measurement applies only to the lysimeter site and to the soil type and structure within the lysimeter.

Another technique that has been shown to compare favorably with lysimeter measurements of evapotranspiration involves very accurate measurements of temperature and vapor pressure at two or more heights above the ground (or crop canopy) and of vertical wind-velocity profiles. The aerodynamic profile method (as termed by Saxton and McGuinness, 1982) uses temperature and moisture gradients and wind transport data to calculate the mass transfer of moisture away from the land surface. Very sensitive equipment is required and the resulting accumulation of data is voluminous. The aerodynamic profile method (see, for example, Swinbank, 1951; Dyer, 1961) is useful for research applications but seldom is used in water-resources studies (Burman and others, 1980).

The data-intensive methods are becoming more practical with the advent of electronic environmental sensors and data loggers that allow the collection and computer entry of large amounts of climatic and hydrologic data with much less human intervention than was required previously.

The most commonly used methods for the estimation of evapotranspiration involve formulas that require climatic data. An assumption inherent to each of the methods is that the soil-moisture supply does not limit evapotranspiration. The equations are written in terms of potential evapotranspiration, the rate at which water would be transferred from soil and plant surfaces under conditions of vigorous plant growth (with several inches of emerged foliage) and ideal moisture availability, and are expressed as the latent heat transfer per unit area or an equivalent depth of water per unit area (Burman and others, 1980). Some of the equations have a physical basis, but others are based only on empirical observations. The availability of climatic data normally dictates which method will be used for a given study.

Probably the two most widely used methods for determining potential evapotranspiration are those introduced by Penman (1948, 1956) and by Jensen and Haise (1963). Penman determined potential evapotranspiration by combining a vertical energy budget with an aerodynamic theory for turbulent transport. His method has been extensively tested and refined, and now is one of the most widely used and reliable methods for determining potential evapotranspiration from climatic data (Saxton and McGuinness, 1982). Data requirements for the Penman method include net radiation, air temperature, air humidity, and horizontal wind travel. The Jensen-Haise method involves less computation and requires only air-temperature and solar-radiation data. However, the Jensen-Haise method neglects vapor transport by convection or turbulence and consequently underestimates potential evapotranspiration for consistently windy areas.

The values determined by any of the potential evapotranspiration methods must be adjusted to represent actual evapotranspiration. The potential evapotranspiration value represents the upper limit of evaporative potential under ambient atmospheric conditions and must be reduced if the ideal conditions (vigorous plant growth and unlimited soil-moisture supply) assumed in deriving the potential evapotranspiration equation are not met. The reduction usually is accomplished by multiplying the potential evapotranspiration by experimentally developed evapotranspiration crop coefficients. In its most comprehensive form, the crop coefficient accounts for plant growth state, available soil moisture, and accelerated surface evaporation following rainfall and irrigation applications. Because it is an empirically derived coefficient, a hinderance to its general use for determining actual evapotranspiration from potential evapotranspiration is that any given crop coefficient applies only to a specific crop in a particular locality. Crop curves (plots of crop coefficient versus time since emergence of the crop) designed for the adjustment of potential evapotranspiration calculated by the Jensen-Haise method are available for several crops, including corn and alfalfa, for southeastern North Dakota (Stegman and others, 1977).

POTENTIAL CAUSES OF NONIDEAL FLOW BEHAVIOR

Significant advancements have been made during the last two decades in our ability to solve practical problems involving flow in saturated and unsaturated media. Physical theories that were developed over a period of more than 50 years are now the basis for a large array of computer models that may be applied to problems that were considered insoluble just a few years earlier. A common feature of most of the applications to real problems, however, has been the idealization of the systems under study.

Idealization generally has been necessary, either because the data to properly characterize the system are not available or because the theory relating to the nonideal behavior has not been developed. Regarding the movement of water in unsaturated media, particularly with respect to soil-water processes of interest in North Dakota, four potentially troublesome phenomena could contribute to nonideal behavior or otherwise impair the ability to reliably simulate moisture movement between the land surface and the water table. The four phenomena that are addressed in the following section are: (1) Variability of soil hydraulic properties, (2) hysteresis, (3) frozen ground, and (4) macropore effects on soil-water flow.

Variability of Soil Hydraulic Properties

The task of measuring soil hydraulic properties is complicated by the inherent variability of those properties, both spatially and temporally. In the development and applications of the theories of soil-water movement, the variation of hydraulic properties with time generally has been ignored--probably with adequate justification for all except the surface layers that are physically disturbed by raindrop and traffic compaction, wind and water erosion, and tillage effects. Areal variation of hydraulic properties, however, is an undisputable reality, even within the most narrowly defined soil taxonomic unit.

Soil heterogeneity and its effects on water movement has received much attention in recent years, probably as an outgrowth of the general recognition of the stochastic or probabilistic behavior of natural processes. Since the early 1970's, many researchers have examined the statistical variation of soil-water properties over variously sized plots (see, for example, Nielsen and others, 1973). Warrick and Nielsen (1980), in summarizing the results of eight such studies, listed the mean, standard deviation, and coefficient of variation for several different soil physical properties. Freeze (1975) pointed out that most evidence points to a log-normal probability density function for unsaturated hydraulic conductivity, whereas a normally distributed probability density function consistently is reported for porosity.

An important consideration is how to account for the observed variability. Rogowski (1972) contended that as long as the coefficients of variation for moisture content and hydraulic conductivity for a given area are within a prescribed range, the soil properties could be assumed uniform within that area and could be represented in models by their mean values. Often, however, the variances are not within the prescribed range and some other method must be used to deal with the variability. An approach introduced by Miller and Miller (1956) assumes that the pore geometries of media throughout a given area differ only in scale and that the magnitudes of the hydraulic properties vary in proportion to the ratio of the scales. In principal, the concept of similar media allows the determination of soil hydraulic properties at any site, based only on the value of a single physical characteristic (such as porosity or particle size) at the site. Miller (1980) provided a detailed discussion of scaling of soil-water phenomena. Other papers describing applications of scaling theory include Reichardt and others (1975), Peck and others (1977), Warrick and others (1977), Sharma and Luxmoore (1979), Simmons and others (1979), Sharma and others (1980), and Clapp and others (1983).

The effects of soil heterogeneity on hydrologic response may be evaluated by Monte Carlo analysis (Warrick and others, 1977; Smith and Hebbert, 1979). Numeric values for properties (such as hydraulic conductivity) used for hydrologic process models are generated randomly from a prescribed probability density function for each of many simulations. The model results then are used to define the probability distribution for a resultant property (such as hydraulic head) at a specific point in the system. Thus, the researcher may evaluate the probable range of hydrologic responses and the likelihood of any particular response, given a realistic range (defined by the probability density function) for the inherent physical characteristics of the system (the soil hydraulic properties).

Until very recently, studies of the spatial variability of soil properties used the conventional statistics of completely random variables. By ignoring the spatial dependence of soil properties, researchers overestimated the number of samples required to achieve a desired precision of estimation (McBratney and Webster, 1983). Statistical techniques that address the spatial structure of soil properties now are being used to improve sampling design and to derive more reliable interpretations from the acquired data (Vauclin and others, 1983).

Hysteresis

A complication in the relation between matric suction and wetness may introduce significant error in flow calculations that make use of the relation. If the wetness versus matric-suction relation is plotted for a given soil, first as the soil is gradually dried from an initially saturated state and again as it is gradually resaturated, two distinct curves will result. The curves show that the soil wetness at any value of suction is greater during drying than during wetting. If the wetting or drying processes had been reversed prior to completion, the wetness-suction values would not plot on the primary curves but would plot along any of an infinite number of possible curves within the primary loop. The variation of the wetness-suction relation according to whether the soil is wetting or drying is called hysteresis. Hysteresis and its probable causes are detailed in most texts that deal with soil physics, including Nielsen and others (1972), Klute (1973), Bear (1979), and Hillel (1980a). A particularly comprehensive coverage is given in Childs (1969).

Hysteresis can be a problem in flow calculations because of the uncertainty of the capacity term, $C = \frac{d\theta}{d\psi}$. Situations exist, such as vertical drainage from an initially saturated, covered plot, where moisture content changes in a monotonic fashion and hysteresis does not occur. The moisture content in most natural soil profiles, however, fluctuates over fairly short time spans in response to infiltration events and subsequent redistribution accompanied by evapotranspiration. As the infiltration wetting front moves downward, progressively deeper soil zones become first wetter and then drier as the front passes. Moreover, evapotranspiration may desiccate the near-surface soil quite rapidly after infiltration ceases. Thus, it may be anticipated that hysteresis could affect the redistribution process as the various soil zones undergo alternating wetting and drying cycles. It also could be expected that most of the process would occur under wetness-suction relationships defined by the intermediate curves rather than the primary curves. Hillel (1980b, p. 63) provided a thoughtful discussion of the hysteresis-affected redistribution process.

Many attempts have been made to describe hysteresis mathematically, but apparently no theory has emerged that generally is effective in analyzing and predicting the behavior of hysteretic flow systems (Klute, 1973). Reviews of the work devoted to methods for estimating the moisture-characteristic scanning curves from minimal data are provided by Nielsen and others (1972), Klute (1973), Mualem (1973), and Hillel (1980b). The most promising means of satisfactorily treating hysteresis is by numerical methods (Hillel, 1980b). Even then, scant knowledge of the physical properties of the various soil types, rather than the inability to refine the moisture characteristic to reflect hysteretic behavior, will continue to limit the degree to which actual soil-water processes may be simulated.

Most studies involving analyses of flow in unsaturated media have ignored hysteresis. Apparently the general thought was that either the process was not of consequence in the system under study or, even if hysteresis was considered important, the process was so complicated as to defy treatment. In any event, there is little evidence in the literature of concern for the validity of interpretations that have not considered the effects of hysteresis on unsaturated flow. An appropriate course of action regarding the hysteresis problem was offered by Klute (1973, p. 26): "It is desirable to establish the extent of the deviations and perturbations caused by hysteresis and to search for and develop simplified means of coping with hysteresis in the analysis of flow systems. Evaluations (on the basis of the intended application of flow theory) must be made of the errors introduced either by neglect of hysteresis or by the use of an approximate treatment of it. In some cases of flow, one can be quite sure, from knowledge of the boundary and initial conditions, that hysteresis either cannot be involved or if it is, the extent to which it would be displayed would be minimal. In field soils, with nonuniform initial conditions and cyclic boundary conditions, hysteresis must play a role in the behavior of the flow. The question to be answered clearly is 'how much?' "

Frozen Ground

There seems to be a widely held premise that water is immobile in frozen ground and that infiltration into such soils cannot occur. The hydrologic implication is that spring snowmelt cannot infiltrate the soil but either must be stored in the snowpack or it must run off until the soil has been warmed sufficiently for the frozen pore water to become liquid. Only then, the concept holds, may water infiltrate the soil and percolate to the water table to become recharge. Unfortunately, little empirical basis exists to prove or refute this premise. Experimental and theoretical research has, however, provided some knowledge of the hydraulic attributes of frozen ground. A few of the most relevant findings of that research relating to ground-water recharge will be mentioned here. The interested reader may refer to Dingman (1975) for a comprehensive review of the hydrologic effects of frozen ground.

Two characteristics of the freezing process in wetted porous media assure some permeability even at subzero temperatures. As natural soils begin to freeze, ice forms first in the central part of the larger water-filled pores. Water remains fluid near the soil particles, maintaining a continuous route for flow, albeit more tortuous than prior to freezing. The thickness of this fluid film between the ice and soil particles rapidly decreases with decreasing temperature, but the film may persist until the soil temperature falls to many degrees below zero (Harlan, 1973). Also, experiments with variously sized capillary tubes indicate that increasingly smaller pores require increasingly lower temperatures to initiate freezing (Dingman, 1975). Therefore, the smaller soil pores remain ice free and offer unconstricted flow even after the commencement of freezing in the larger pores.

Although appreciable moisture flux may occur in partially frozen soils, that flux varies appreciably with the stage of the freezing process. Decreasing soil temperatures advance the freezing process and reduce the potential rate of flow through the soil. For soils that are saturated at the onset of freezing, the relative reduction in hydraulic conductivity with decreasing temperature is very abrupt (Burt and Williams, 1976), and liquid flux rates within the frozen zone may become insignificantly small as the soil cools to well below zero. Soils that are relatively dry when frozen, however, contain considerable pore space that is air filled and able to conduct water when wetted. Thus, the dry soils retain much of the ice-free hydraulic conductivity.

Although soil-water flow generally is associated with gradients in hydraulic potential, flow induced by temperature gradients also is a common phenomenon. A familiar hydrologic result of the tendency of soil water to migrate from warm regions to cold regions is the gradual decline through the winter months of water levels in shallow (on the order of 10 feet or less) water-table wells. The near-surface frost zone acts as a strong sink during the redistribution of water from the warmer soil beneath. Redeposition of the upward migrating water as ice in the frost zone actually may cause the total water content of the near-surface soil to exceed the normal saturated level. Willis and others (1964) attributed the redistribution largely to vapor movement, but Miller (1980) indicated that freezing-induced redistribution can occur much too rapidly to be explained by vapor transport.

Redistribution of soil water into the frost zone can have a a significant affect on the hydrologic properties of the near-surface soil. Kane (1981) observed freezing-induced redistribution in a seasonally frozen silt loam soil and found that, for relatively wet soils, the accretion of moisture (as ice) near the soil surface reduced the infiltration rate and the hydraulic conductivity. He found that little moisture redistribution occurred in relatively dry soils, thereby largely preserving the ice-free hydraulic conductivity and, thus, the infiltration capacity to accommodate spring snowmelt. Results of infiltration tests on frozen soils with varying moisture contents showed that generally the soil-moisture content at the time of freezing was inversely related to the ensuing infiltration rate. The implication is that soils that are dry at freeze-up will have large infiltration capacities when snowmelt begins, even though still frozen.

Much remains to be learned regarding the hydrologic characteristics of frozen ground. The flow process is complicated due to simultaneous transport of heat and water and the major effect that minor temperature changes can have on the water flow rates due to freezing and thawing. The development and widespread use of models capable of reliably simulating the simultaneous transport of heat and water will greatly enhance our practical knowledge of the subject. Meanwhile, it is apparent that grain size and antecedent moisture content affect the ability of a soil to accommodate moisture supplied by snowmelt and early spring rains.

Macropore Effects on Soil-Water Flow

The classic theory of soil-water movement describes flow through micropores between the particles of the solid matrix. The assumed laminar flow is in response to gradients in the total potential--in which the Bernoulli velocity term is inconsequential. However, within the last few years, there has been a growing awareness that much of the vertical water flux in some field soils takes place via openings much larger than the normal intergranular pore spaces. The large openings, popularly known as macropores, may extend continuously to a depth of several feet or more and enable infiltration fluxes many times greater than would otherwise occur in their immediate vicinity. These large fluxes through macropores may occur even when most of the soil matrix is unsaturated.

The implications of macropore flow on the infiltration and redistribution of water in soils have been addressed often in recent literature. Beven and Germann (1982) provided a comprehensive review of this literature. Their paper was drawn upon freely for the following information.

Macropores may be formed by various earth-dwelling animals and by plant roots, as well as by the crack- or fissure-producing effects of desiccation and freeze-thaw cycles. Land use is an important determinant of the type of macropores formed and their longevity. Macropores formed in some soils may last for decades if left undisturbed by man's activities. Tillage and compaction by heavy machinery tend to destroy macropores.

Soil-water flow via macropores is difficult to measure because the macropores occur irregularly and because the mechanics of flow within the macropores are poorly defined. The irregularity imparts great significance to the scale of investigation. If a small study plot is selected that contains either no macropores or an uncharacteristically large concentration of them, scaling of the results from flow determinations on the small plot may result in a poor representation of flow rates for the entire study area.

Thus, selection of a representative area for field testing of hydraulic properties is critical where macropores significantly influence flow.

A common assumption is that flow within macropores, rather than conforming to Darcy's law, approaches open-channel or pipe-flow conditions. Whether or not this assumption is valid, potentiometric gradients measured in the soil matrix may differ greatly from the gradients inducing flow in nearby macropores. Discontinuity of soil-water potentials in the vicinity of macropores limits our ability to make reliable flow calculations in soils with abundant macropores. Although a few models have been developed to simulate flow in such soils, the macropore geometry must be so idealized that the models are of marginal utility.

More field studies are needed to determine the characteristics of macropores and their effect on the movement of soil water.

SOIL RESEARCH IN NORTH DAKOTA

Studies of the physical properties of North Dakota's soils have been sponsored mainly by four agencies or institutions: North Dakota State University, through both the Agricultural Experiment Station and the Soils Department; the Agricultural Research Service and the Soil Conservation Service, both of the U.S. Department of Agriculture; and the Bureau of Reclamation, U.S. Department of the Interior. Although most of the effort devoted to soils studies by these agencies has not involved acquisition of the quantitative physical data required for a rigorous evaluation of water movement through the unsaturated zone and ensuing ground-water recharge, the previous work does offer valuable insights into the nature of soil-water-plant relationships for many soil types across North Dakota.

Probably the most familiar of any soils studies that have been conducted in North Dakota are the soil surveys done cooperatively by the Soil Conservation Service and the North Dakota Agricultural Experiment Station. Soil surveys have been completed for about half of North Dakota's counties, and surveys currently (1983) are in progress in 11 other counties. Surveys have not been started, however, in eight counties in southeastern North Dakota. As a part of the soil-survey effort, referred to as the soil characterization study, 15 to 20 profiles are studied and sampled each year for laboratory analysis to determine selected physical and chemical properties of the soils. At least one profile for most soil series in the State has been included in the study. Each horizon in the selected profiles, down to a depth of about 60 inches, is analyzed. The physical characteristics determined for each sample include particle size; bulk density; water content or wetness at 1/10-, 1/3-, and 15-bar suctions; and the water content or wetness at saturation. Most of these data have not been published, but are available for inspection at the Soils Department at North Dakota State University.

The U.S. Bureau of Reclamation has conducted soils studies in areas of North Dakota that are a part of the Garrison diversion project. The work involved determination of the physical properties that affect the suitability of the soil for irrigation and drainage. The data include saturated lateral hydraulic conductivities determined by auger-hole and pump-in tests and saturated vertical hydraulic conductivities determined by ring permeameter tests (U.S. Department of the Interior, 1978), laboratory determined available water capacity, and descriptive lithologic logs for shallow auger holes. Saturated lateral hydraulic conductivities are available for several hundred sites and saturated vertical hydraulic conductivities for about 50 sites in eastern North Dakota. Descriptive lithologic logs number about 30 per square mile in the Bureau project areas. These data are not available in computerized files nor have most of the data been published. The data are available for inspection at the U.S. Bureau of Reclamation office in Bismarck, N.Dak.

A series of experiments was conducted by the Agricultural Research Service, U.S. Department of Agriculture, to evaluate the effects of water-table depth and irrigation application rates on crop yields in medium- to coarse-textured soils of glacial origin (Follett and others, 1970; Follett and others, 1974; Benz and others, 1978; Benz and others, 1981). The field experiments were conducted near Oakes, N.Dak., on loamy sand and sandy loam soils, mostly of the Hecla series. The research made important contributions to the knowledge of plant-soil-water relationships and raised interesting questions regarding the traditional approach to artificial drainage of croplands with high water tables, but the study was not designed to provide new quantitative hydraulic information on North Dakota soils.

Shay (1980) evaluated the influence of morphological and site characteristics on the hydrologic regime of a small basin in western McHenry County. A dense basal till restricted vertical water movement from the sandy and coarse loamy soils to the extent that artificial drainage was deemed necessary for the site to be irrigated. Particle size, bulk density, porosity, and the moisture characteristic were determined on laboratory samples extracted from selected sites in the study area. Saturated lateral hydraulic conductivities were determined *in situ* by auger-hole and pump-in tests.

Two reports have been published regarding the available water capacity of selected North Dakota soils. The available water capacity of a soil is defined as the amount of water that the soil will retain between field capacity and the water content at which plants will permanently wilt. Field capacity is defined as the amount of water the soil profile will retain against the force of gravity (or the water content at such time that drainage becomes negligible).

Rivers and Shipp (1972) reported on a Bureau of Reclamation study of the available water capacity for glacially derived soils in six different areas in central and eastern North Dakota. They found that the standard laboratory determination of field capacity for sandy soils, the moisture content at 1/10-bar suction, was consistently less than the value obtained by sampling the in-place field soil after freely draining for 48 hours following a thorough wetting. Available water capacities were found to range from 2.8 inches per 48 inches for a coarse sand to 5.8 inches per 48 inches for a loamy fine sand. Regression analysis indicated that percent silt content generally was not a reliable index of the available water capacity.

Cassel and Sweeney (1974) used an *in situ* monitoring technique to determine the available water capacity, generally in 6-inch intervals, for 28 soil profiles representing 26 different soil types and 22 soil series. Data also were published for bulk density, particle size, *in situ* field capacity, and a laboratory determined (four- or five-point) moisture characteristic.

Apparently only two studies have been conducted in North Dakota that involved a determination of unsaturated hydraulic conductivity. Cassel (1974) reported unsaturated hydraulic-conductivity data as a function of water content for 11 North Dakota soils having glacial and lacustrine parent material and various textural compositions. Essentially, he used the instantaneous profile method, but determined the moisture-characteristic curves from cores tested in the laboratory. Carvallo and others (1976) evaluated the spatial variability of unsaturated hydraulic conductivity on a Maddock sandy loam soil within a 0.025-acre area near Oakes. Seven depths were tested within each of five 8-foot square plots. The plots were diked and irrigated, then covered to prevent evaporation during the ensuing monitoring period. Potential gradients were determined *in situ* by tensiometry, and the moisture-characteristic curves were derived from laboratory core samples. Unsaturated hydraulic conductivities were found to vary, at the 1 percent level of significance, with depth as well as areally.

Other investigations of soil-water processes in North Dakota include research on the feasibility and potential effects of irrigation planned for the Garrison diversion project. The Soils Department at North Dakota State University in cooperation with the U.S. Bureau of Reclamation is conducting research at Oakes to determine the quality of water that may be expected to return to the James River as drainage and deep percolation following irrigation applications to adjacent project areas. Data are obtained from four intensively instrumented lysimeters (nonweighing type).

The Agricultural Research Service, U.S. Department of Agriculture, is involved in research to evaluate nitrogen movement and the degree of salinity accumulation under various irrigation regimes in medium- to fine-textured soils that have restrictive layers at shallow depths. The work is being done on two small study sites in the Apple Creek area of Burleigh County, N.Dak. Key installations at each site include a climate station and 36 nonweighing, bottomless lysimeters--each equipped with tensiometers and a neutron moisture-access tube. The Burleigh County Water Resources District and the North Dakota State Water Commission are project cooperators.

In a study unrelated to the Garrison diversion project, researchers at the Soils Department, North Dakota State University, are evaluating the effects of various cropping and management alternatives on the soil-water balance for a silty clay soil at Fargo, N.Dak. Four precision weighing lysimeters and neutron moisture-access tubes are used to monitor the water balance. A complete climatological station provides the data required for the calculation of potential evapotranspiration.

PARAMETRIC VERSUS PHYSICS-BASED APPROACHES TO STUDYING RECHARGE

Methods for the determination of ground-water recharge may vary from very simple parametric techniques to rigorous physics-based techniques involving numerical solution of the differential equation for flow in unsaturated media. A parametric model may be as simple as:

Annual recharge = [annual precipitation] x [transfer coefficient],

where the transfer coefficient is some value between zero and unity. The parametric model also may be a much more complex relation involving many variables or it may be a series of coupled relations. A typical physics-based model would solve the Richards equation for one-, two-, or three-dimensional flow between the land surface and the water table and would incorporate treatment of infiltration, evapotranspiration, and seepage faces, if applicable, in the assignment of boundary conditions.

Data requirements for the parametric models normally are minor relative to those for the physics-based models. In fact, as a general rule, the quantity of data required by any of the models increases in proportion to its physical basis. Also, as a general rule, the cost of otherwise similar studies varies in direct proportion to the intensity of original data collection. Finally, the accuracy or reliability of a given model generally is thought to increase in proportion to its physical basis. From these generalities it may be inferred that approaches to studying recharge that have strong physical basis require the most data and are the most costly to conduct, but also may be expected to give the best results.

Assuming the inference that increasingly reliable information can be obtained at increasing greater cost is valid, the water-resources manager should be able to select a study approach based on the economic and social value of the water resources in question. Water resources of great value presumably would warrant a large investment in reliable information, so that the probability of a costly or harmful mistake in the management of the resource would be minimized. By similar reasoning, water resources of lesser value would warrant less investment in the reliability of information collected.

Unfortunately, the relationship between cost and reliability for the various possible approaches is not straightforward. For any given study area, an approach involving more data collection at greater cost than some other approach would not necessarily result in a better estimate of recharge. A claim may be made with some certainty, however, that the flow of water through a given soil profile may be estimated more reliably using a physics-based model and the relevant physical data than by almost any parametric model. The uncertainty arises in extrapolating the profile results to entire study areas with little or no knowledge of the degree of inhomogeneity of soil properties in the study area. If the validity of physics-based model simulations frequently is in doubt due to the inhomogeneity problem, then it may be advisable to employ the much less costly parametric approaches to achieve results of equivalent validity. Little progress has been made to resolve the validity question since the introduction of the physics-based models nearly two decades ago. The main reason is that very limited data have been available to determine the degree of spatial variability of soil hydraulic properties and to evaluate how severely that variability limits our ability to make valid simulations of the soil-water flow process.

For aquifers of small areal extent that supply or have the potential to supply water of large economic value, the intensive data collection needed to use a physics-based approach may be justifiable. For aquifers of very large areal extent, it may not be feasible to collect sufficient data to represent the natural variations of the hydraulic properties of the soil. Parametric models for those aquifers probably would give results as reliable as physics-based models that have insufficient distributed data to characterize the properties of the field media. Thus, the size of the aquifer under study, its economic value, and the resources available for data collection are the primary influences on the choice of an approach for estimating recharge.

DISCUSSION AND CONCLUSIONS

A review of the hydrologic literature indicates that relatively few attempts have been made to measure ground-water recharge based on established methods for determining the flow of water through the unsaturated zone. Most recharge studies have used approximate water budgets or have relied on the analysis of changes in ground-water levels. The water-budget and water-level methods have little basis in the physics of flow between the land surface and the water table. As a result, empirical relations developed from these studies generally have little transfer value to other areas.

Physical theory of the flow of water through unsaturated media has been available for several decades. Soil physicists have documented many analytical solutions to the flow equation for simple boundary conditions of infiltration, evaporation, and drainage toward tiles and seepage faces. Analytical solutions are not available, however, for the more complex conditions that normally occur and that commonly are of interest to soil scientists, hydrologists, and engineers. To address real flow problems that have nonideal boundary and initial conditions, researchers have developed computer models of the unsaturated or "variably" saturated flow process to simulate flow in systems of virtually any geometry. Little practical use, however, has been made of these models by hydrologists.

The reason that such a potentially useful technology has not received more general use is that the appropriate data are not available. Any physics-based model of unsaturated flow requires data on the relation of matric potential and hydraulic conductivity to wetness. The determination of these relations is tedious and costly, and rarely has been done outside the realm of soil physics research. Data, therefore, are available for several small research plots, but systematic appraisals of the pertinent hydraulic characteristics for areal hydrologic studies are almost nonexistent.

Many attempts have been made to estimate the required relationships based on limited data, and many models have been developed that are claimed to give results comparable to measured values. None of the models are universally applicable though, and fitting coefficients used in most of the models must be evaluated against measured values. Thus, a prerequisite to widespread use of quantitative physics-based approaches to modeling the ground-water recharge process is the development of a greatly expanded data base for the hydraulic properties of soils.

Laboratory methods probably provide more accurate values for the moisture characteristic and unsaturated hydraulic conductivity of a small discrete sample than do *in situ* methods. However, the small size of laboratory samples and the unavoidable disturbance of the samples during field extraction, transport, and laboratory handling lessen the representativeness of laboratory values relative to those actually occurring in the field. The apparent consensus of opinion now is that the hydraulic data, especially for hydraulic conductivity, should be determined *in situ* whenever possible. Methodology for the *in situ* measurement of matric potential and wetness, and the ensuing development of the moisture-characteristic and hydraulic-conductivity relationships are well established, but seldom applied.

Several factors could influence the reliability of recharge estimates derived from a physics-based analysis of water movement through the unsaturated zone. The most troublesome factor is the areal variability of soil hydraulic properties and, related to that, the appropriate means of scaling data gathered at selected test plots so as to represent the entire study area. Recent studies demonstrate the potential for extreme variability of hydraulic properties and indicate the need to define the statistical nature, including the spatial structure, of the variations. Deterministic flow models then could be made to reflect realistic spatial variations of the property values. If data analysis indicated the property values to be so random as to preclude mapping them at some practical scale, the flow process could be simulated with a stochastic model. Values for such a model could be generated randomly from a population of values with a known mean and standard deviation. Either modeling approach, therefore, requires the collection of a substantial quantity of data.

Other factors may complicate the analysis of unsaturated flow even for a media that is thoroughly characterized with regard to the moisture-characteristic curve and the hydraulic-conductivity function. One of the factors, hysteresis, is known to occur in most soils and may be especially prevalent in coarse-textured soils. Because the moisture-characteristic curve actually includes an infinite number of possible scanning curves, each associated with some combination of wetting and drying cycles, even the most sophisticated computer models cannot determine an exact wetness from a given suction. Even if that determination were possible, the cost of computer time and storage required to accomplish it would be prohibitive to most studies. Extreme accuracy in the suction-wetness relation, however, may not be essential to achieve satisfactory solutions to practical problems involving the simulation of unsaturated flow. Perhaps for most applications, the primary wetting and drying curves and some simple routine for estimating intermediate values of suction corresponding to some brief wetness history would suffice. Sensitivity analyses using measured ranges between the primary characteristic curves should indicate whether hysteresis would jeopardize the validity of flow simulations for a particular soil or association of soils.

Seasonally frozen ground affects the processes of infiltration, runoff, and soil-moisture redistribution in temperate climatic areas such as North Dakota. A widely accepted premise is that most of the annual ground-water recharge occurs during the spring snowmelt period when the soil profiles are subject to deep and prolonged wetting and evapotranspiration is minimal. The major obstacle to the determination of infiltration and redistribution under partially frozen conditions is the lack of data relating hydraulic conductivity to soil temperature. Experimental work is needed to establish the conductivity-temperature relationship for the representative grain-size classes of soils. Heat and moisture flow models then could be applied to solve problems dealing with moisture movement during freeze-thaw periods.

Macropore development and the associated difficulties in defining moisture flux rates may not pose a great hindrance to the determination of recharge for many of North Dakota's aquifers. The glacial aquifers typically are overlain by coarse-textured soils that are farmed (tilled) intensively--a soil and land-use combination that is not conducive to macropore formation. In those situations where macropore flow cannot be disregarded, field measurements could be made to gauge the macropore contribution to the total moisture flux. Lysimetry currently is the only plausible means to make such measurements. Macropore flow models either are not yet sophisticated enough or require too much nonexistent data to produce accurate estimates of flow rates.

The phenomena that contribute to nonideal flow behavior should not be considered barriers to the use of physics-based approaches for the study of recharge. They should be viewed as complicating factors for which the investigator must compensate. An analysis of the sensitivity of the flow system to observed or estimated values of the unsaturated-zone hydraulic properties could be done concurrently with the data-collection activities. Then if, for example, hysteresis is found to affect soil-water movement only to a minor degree, further effort would not be expended to define the hysteretic nature of the characteristics curve. Coordination of the data collection with the model application would help to ensure that the proper data and the appropriate quantity of data are made available for whatever quantitative analysis is required to meet the project objectives.

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