The Connecticut Agricultural Experiment Station, New Haven

Ground Water in the Northeast

CHARLES R. FRINK, EDITOR

Bulletin 876

December 1989

Northeastern Regional Research Publication

Technical Committee of Northeastern
Regional Research Coordinating Project
NEC-28, Soil Research

Administrative Advisor

Connecticut (New Haven) .... C.R. Frink

Technical Committee

Connecticut (New Haven) .... C.R. Frink
Connecticut (Storrs) .... H.D. Luce
Delaware .... J.T. Sims
Maine .... R.V. Rourke
Maryland .... D.S. Fanning
Massachusetts .... J.H. Baker, P.L.M. Veneman
New Hampshire .... R.D. Harter
New Jersey .... L.A. Douglas
New York .... D.R. Bouldin, N.H. Peck, R.J. Wagenet
Pennsylvania .... D.E. Baker
Rhode Island .... W.R. Wright
Vermont .... R.J. Bartlett, F.R. Magdoff
Washington, DC .... J.R. Preer
West Virginia .... R.F. Keeber, R.N. Singh

Cooperating Federal Agencies

Agricultural Research Service, USDA .... H.B. Pionke
Cooperative State Research Service, USDA .... B.L. Schmidt, C.M. Smith
Soil Conservation Service, USDA .... K.H. Langlois, Jr.
Tennessee Valley Authority .... P.M. Giordano, K.R. Kelley

Preface

The Northeast Soil Research Committee (NEC-28) commissioned the three papers presented here to stimulate new and innovative approaches to understanding the movement of agricultural chemicals and other contaminants to ground water. The papers were written by specialists to assist soil scientists in related disciplines in designing and conducting meaningful experiments in both the laboratory and field. They were also written to aid in making the best and most efficient use of additional funds for research on ground water quality available under the Water Quality Initiative within the Cooperative State Research Service, USDA.

The importance of the topic is underscored by our increasing awareness of the extent of ground water contamination and by Congressional amendments to the Clean Water Act requiring increased control of nonpoint sources of pollutants in ground water.

A synopsis of the three papers follows, along with some of my own observations and conclusions.

Concepts and Models of Water Flow in Macropore Soils

As Wagenet and Germann point out, classical concepts consider the soil as a homogeneous porous media through which water moves uniformly in response to gradients in potential energy. Recently, there have been sufficient observations in the field to demonstrate that the flow of water often differs from these classical concepts. Rather, most of the water tends to move rapidly along preferred pathways that have come to be known as macropores. While there are differences of opinion as to what constitutes a macropore, the consequences are clear: water and dissolved solutes can move rapidly downward through a small fraction of the total cross-sectional area of the soil. This pattern of flow has at least two practical consequences. First, contaminants may reach ground water long before predicted by classical concepts. Second, large amounts of solute will remain in the soil in areas by-passed by the rapid flow in macropores.

The implications of this kind of flow on the extrapolation of studies conducted in the laboratory are also substantial. Almost without fail such studies are conducted with homogeneous soils under "equilibrium" conditions. A sorption maximum measured in this fashion probably has little relationship to reality in the field. Thus, Wagenet and Germann recommend that future studies focus on soil morphology and macropore flow dynamics.

Estimating Parameters Controlling Saturated and Unsaturated Flow in Soils

Despite the uncertainties in modeling water flow in soil, we need to know how to describe water flow and transport of solute at the field, landform and watershed scale. As Rogowski states, we need to know over what volume of soil and interval of time hydraulic and chemical properties of soil should be averaged.

Both well tried and novel statistical methods need to be encouraged. Hierarchical analysis of variance and geostatistical tools of structural analysis, kriging, probability kriging, and risk analysis should be emphasized, along with methods exploring applicability of fractals, time series, and spectral processes. The difficulties center primarily around the concept of homogeneity and the need for adequate numbers of samples.

Variograms, employed to ascertain the structure of distributions on a field or watershed scale, also hold promise. Additional effort is needed to apply predictive and extrapolative methods of geostatistics to extremely large data sets, such as those of DEM and associated digitized soils, cover, and geologic information. GIS and geostatistical approaches must also be integrated on a large watershed scale.

In conclusion Rogowski offers a brief quotation from Burges: "Progress is made by those who take as broad a view as possible and recognize the totality of a problem, rather than a fragment."

Ground Water Flow in the Northeast

The deterioration of the quality of our surface waters, and the lack of adequate storage revealed during the drought of the mid-60's has caused us to look increasingly at ground water as an alternative source of supply. As Gburek points out, however, legal and political aspects of subsurface flow systems can hinder our ability to manage ground and surface water supplies wisely.

Historically, the riparian doctrine of water use has prevailed in the Northeast, providing for unrestricted use of ground water by each landowner. Continuing legal tests have caused a shift toward a doctrine of reasonable use which restricts the water rights of the landowner in relation to the needs of others. This transitory stage in water law tends to impede a rational approach to management of our water resources.

Many contend that a conjunctive use program is the most appropriate for the Northeast. This involves balancing the use of ground water and surface water reservoirs based on supply and demand. During high flow periods, excess surface water is used to recharge ground water, while during low flow, ground water is pumped to augment surface supplies. This concept has traditionally been applied to the problem of water supply, but there is no reason why it cannot be extended to the problem of waste disposal as well. However, the control of water supply and waste disposal in the Northeast is commonly vested within different agencies, departments, or authorities which tends to impede rational resource management.

Because aquifers provide a high percentage of flow to streams, the effects of an individual aquifer may be observed on a much larger scale. For instance, nutrient input to Chesapeake Bay is currently a concern. Here, regulation of ground water quality at the state level may not be sufficient because the large scale subsurface flow system produces a large scale surface water system transcending state boundaries.

Ground water management affects the near-stream environment as well. Wetlands are generally zones of ground water discharge and removal. Ground water contamination will thus have a direct effect on wetlands and their outflow. The riparian zone can also serve as a site for water quality management, because, like wetlands, it provides ideal conditions for denitrification. Such controls must be quantified since little research has been conducted to date.

Finally, agricultural impacts on ground water must generally be analyzed in the context of a watershed system. The typical watershed in the Northeast consists of intermixed urban, suburban, crop, and forest lands all contributing contaminants to the same subsurface flow system.

Gburek concludes that the humid climate of the Northeast assures us of continuing abundant supplies of water. The nature of our shallow ground water aquifers insures that they will be refilled by precipitation annually. The challenge is to prevent further deterioration of our ground water in an area where agriculture and industry are coupled with the highest population density in the country.

Charles R. Frink

Administrative Advisor, NEC-28

Vice Director
The Connecticut Agricultural Experiment Station
New Haven, CT 06504

Concepts and Models of Water Flow in Macropore Soils

R.J. Wagenet
Professor
Department of Agronomy
Cornell University
Ithaca, NY 14853

P.F. Germann
Professor
Institute of Geography
University of Bern
Bern, Switzerland

Joint Contribution from the New York Agricultural Experiment Station (Department of Agronomy publication 1642) and the New Jersey Agricultural Experiment Station (publication E-15235-2-88).

CONTENTS

INTRODUCTION

Classical concepts of water movement consider the soil as a homogeneous porous media composed of both large and small pores. Water is assumed to flow in response to gradients in potential energy which are determined by a variety of forces acting upon the water molecules. The principal forces are matrix forces including gravity, adhesion and cohesion, and hydrostatic forces resulting from the pressure of overlying water. Water status and flow in soils have been measured by numerous techniques and models have been derived to predict the flux of water and chemicals through soil.

Within the last ten years, observations in the field have convincingly demonstrated that flow of water in soils rarely follows predictions based on these classical concepts. In fact, in well-aggregated soils, a relatively few large openings can result in movement of water and solutes at rates far exceeding predictions based on classic concepts. Also, there can be substantial movement of water through such soils without displacement of resident chemicals. These non-homogeneous soils can be characterized by a bulk soil matrix, but they also contain large openings such as cracks, fissures, and channels that are not described by classic concepts of flow.

It is interesting that such deviations from ideality were recognized over a century ago, for as Beven and Germann (1982) pointed out, Schumacher (1864), wrote: "...the permeability of a soil during infiltration is mainly controlled by big pores, in which the water is not held under the influence of capillary forces." In fact, the "big pores" are of several types including large cracks, channels, and even wormholes, that can act as preferred pathways for flow. In such soils, there will be only partial displacement of resident water and solutes by incoming water. The consequences include increased opportunity for transport of water and solutes to deeper depths than predicted using classical concepts, as well as decreased opportunity for leaching of resident chemicals even in the presence of substantial fluxes of water.

These conditional flow events cannot be predicted unless the geometry and continuity of the preferred pathways, or macropores, is known. Even then, the quantitative influence of these macropores in any particular soil depends in part upon conditions at the soil surface. However, their importance was recognized early in experiments at Rothamsted, when Lawes et al. (1882) reported: "The drainage water of a soil may thus be of two kinds: it may consist (1) of rainwater that has passed with but little change in composition down the open channels of the soil or (2) of the water discharged from the pores of a saturated soil." Although these peculiarities of flow in soils in the field have been recognized for over 100 years, few techniques have been developed to study and describe them.

This discussion first outlines the well-established classical approach to description of water movement in soil based on numerous references in the soils literature. Excellent reviews are provided by van Genuchten and Alves (1982) and Nielsen, et al. (1986). The balance focuses on water and chemical movement in macropore systems which are less well understood.

CONCEPTS

Classical Approach in Homogeneous Soils

The basic relationship describing water flow through saturated materials was postulated by Henry Darcy (Darcy 1856), and an understanding of the forces of soil-water retention was first provided by Buckingham (1907). The relationship between water content and the potential energy status of soil water followed about 15 years later (Gardner 1920), with the basic equations describing water flow in unsaturated soils being presented in 1931 (Richards 1931). By the 1950's, the formulation of these concepts was the basis for description of water movement in homogeneous soils.

During the period that immediately followed, these principles were tested, evaluated, and refined and a substantial technology was developed to measure water

water if they are continuous throughout a soil sample (Bouma et al. 1979). Perhaps macropores are best defined as pores that are significantly larger than those resulting from the simple packing of the individual soil particles. The collective result, the macropore volume, has been identified from a sharp change in hydraulic conductivity as a saturated soil drains (Germann and Beven 1981). For large undisturbed blocks of a soil derived from Oxford Clay, these authors reported macroporosities of 0.01 to 0.045 $cm^3\ cm^{-3}$ which is approximately 5-10% of the total pore volume.

Table 1. Definitions of Macropores and Macroporosity (Beven and Germann 1982).

References	Capillary Potential, kPa	Equivalent Diameter, μm
Nelson and Baver (1940)	>-3.0	
Marshall (1959)	>-10.0	>30
Brewer (1964)		
Coarse macropores	5,000	
Medium macropores	2,000-5,000	
Fine macropores	1,000-5,000	
Very fine macropores	75-1,000	
McDonald (1967)	>-6.0	
Bullock and Thomasson (1979)	>-5.0	>60
Reeves (1980)		
Enlarged macrofissures	2,000-10,000	
Macrofissures	200-2,000	
Luxmoore (1981)	>-0.3	>1,000
Beven and Germann (1981)	>-0.1	>3,000

Size alone is not sufficient to define a macropore. Pore structure and geometry (Bouma 1981a, b; Beven 1981), and soil morphological features are routinely used by soil surveyors to distinguish cracks, voids, fissures and channels. However, these distinctions are generally qualitative. The flow of water down these openings has resulted in the popular application of the terms "preferred pathway," short-circuiting," or "bypass flow" as descriptions of flow external to the bulk soil media.

Formation of Macropores

Macropores have also been grouped according to their method of formation. As presented by Beven and Germann (1982), the major groups are:

Pores formed by soil fauna. These are primarily tubular in shape but may range in size from less than 1 mm to over 50 mm in diameter for holes formed by burrowing animals such as moles, gophers, and worms. Macropores formed by soil fauna are often concentrated close to the soil surface. Earthworm channels of 2-10 mm diameter with 100 channels/m^2 have been reported (Omoti and Wild 1979) in a loam soil, with nearly all continuous to 14 cm depth and about 10% continuous to 70 cm. Ants have been reported (Green and Askew 1965) to produce macropore networks of 2-50 mm diameter to a depth of at least 100 cm. The formation of these macropores can be influenced by the effects of moisture and pH on soil fauna. In acid soils insects tend to dominate, whereas earthworms prefer less acid to neutral soils. A comprehensive review of the activities of animals in soils is given by Hole (1981).

Pores formed by plant roots. Pores in this category are also tubular. Macropores may derive from either live or decayed roots, but the distinction may be difficult, as new roots tend to follow the channels of previous roots. The bark of tree roots sometimes resists decay longer than the xylem, and a hose-type macropore is formed, partially sealed by the bark (Gaiser 1952; Aubertin 1971). Such macropores may comprise up to 35% of the volume of a surface forest soil, but they decrease rapidly with depth (Aubertin 1971). The structure of macropores derived from roots depends on the plant species and the conditions of growth. They may be very effective in channeling water even through unsaturated soils (Aubertin 1971; Beasley 1976; Mosley 1982).

Pores formed by cracks and fissures. These macropores are formed either by chemical weathering of bedrock material (e.g., Reeves 1980) or by shrinkage resulting from desiccation of clay soils (e.g., Blake et al. 1973; Lewis 1977). Cycles of freezing and thawing, drawing of mole drains, and subsoiling may also produce cracks and fissures. Shrinking and swelling of clay soils depends on changes in soil moisture, and, once a crack is formed, it may recur at the same location through a series of wetting and drying cycles. In fact, in drained heavy clay soils, cracks between structural peds may not close even after prolonged wetting (Beven 1980). Interpedal macropores may be utilized by roots and soil fauna, and Reeves et al. (1980) demonstrated how soil cracking can be influenced by the pattern of moisture extractions by roots.

These definitions of macropores are not necessarily related to their ability to conduct water. Consequently, other definitions of the functional macroporosity of a soil are required. Topp and Davis (1981) studied the infiltration of water into soil cracks using time-domain reflectometry and found that the water content increased much faster close to the cracks than at some distance from them. In many cases, dyes have been mixed with the water to indicate the path of the water and the depth to which it has penetrated (Ritchie et al. 1972; Anderson and Bouma 1973; Starr et al. 1978; Omoti and Wild 1979).

content and movement under carefully controlled laboratory conditions. Concurrently, models were derived to predict both steady-state and transient water and chemical movement under laboratory conditions. The success of these efforts resulted in acceptance of the flow theories, and raised hopes that field soils could be described in similar fashion. Since about 1970, description of water and solute movement under field conditions has been attempted using the basic relationships developed in the laboratory. However, numerous field studies have shown that spatial variability of flow properties and non-homogeneity of soils preclude the direct application of these laboratory techniques. In fact, new, complementary approaches (e.g., Nielsen et al. 1981; Beven and Germann 1981) will be required if the fundamental relationships developed in the laboratory can be modified to describe fluxes in non-homogeneous soils in the field.

The classical approach assumes that the steady water flux in a vertical, one-dimensional soil system can be expressed according to Darcy's law as

$$q_w = - K\,(dH/dz) \tag{1}$$

where q_w is the soil water flux, H is the hydraulic head, K is a proportionality factor called the hydraulic conductivity, and z is vertical distance or depth. When the water flow regime is not constant with space and time, Eq. (1) is combined with the continuity equation to become Richards' equation

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left[K(\theta)\,\frac{\partial H}{\partial z}\right] \tag{2}$$

where θ is the volumetric soil water content (L^3/L^3), $K(\theta)$ now represents the dependence of hydraulic conductivity upon the water content, and t is time. In this conceptualization, θ represents all water present in the bulk soil volume, irrespective of whether it occupies large or small pores. The flux of water at depth z during nonsteady flow can then be generally expressed as

$$q_w(z) = q_w(z_o) - (1/\Delta t)\int \Delta\theta\, dz \tag{3}$$

where q_w is the volumetric flux of water, assuming that no other sources or sinks of water exist. In this approach, the flux of water is considered to be a function of the bulk water content θ, without reference to particular and separate contributions that may arise from water residing in "large" or "small" pores.

The classic approach summarized in Eq. (3) is used to describe homogeneous soils characterized by a range of pore sizes which establish a smooth and continuous relationship between water flow and water content. This approach recognizes that water moves most rapidly through the soil when all pores are saturated. However, the concept also presumes that water does not move through pores of such excessive size that their properties overwhelm the smooth and continuous relationship between hydraulic conductivity and water content. In the classic approach, there are no pipes or conduits through which water may bypass the bulk of the soil. Rather, there are large and small pores, all of which participate in each flow event, although the degree of participation of the large pores decreases as water content decreases. More quantitative distinction between homogeneous and non-homogeneous soils is possible by recognizing and defining the role of large pores.

Water Movement in Macropore Soils

Water flow in the presence of large water-conducting openings, often called macropores, has been comprehensively reviewed by several authors, including Beven and Germann (1982) and White (1985b). The following discussion draws heavily upon these articles.

Several methods have been used to classify pore size and to distinguish macropores. The most common has been to interpret the soil moisture retention curve in terms of pore size classes, where a measure of the effective pore size is developed by relating pore radius to the matric potential through the Laplace equation. This involves an analogy between the macroscopic retention characteristics of the soil and the microscopic concepts of the behavior of a bundle of capillary tubes. Under this approach, the minimum dimensions of macropores can be estimated from the relationship between the radius R of a cylindrical pore and the matric potential with which water is held in it. Thus

$$h = -2\,\sigma/(R\,\rho_w\,g) \tag{4}$$

where σ is the surface tension at the air-water interface, ρ_w is the density of water, g is the acceleration due to gravity, and h is the matric potential energy.

From Eq. (4) a pore radius corresponding to a matric potential of zero is undefined. An arbitrary value of $h = -1.0$ cm has been used (Germann and Beven 1981) to indicate a boundary between macropores and micropores. This gives an equivalent radius of approximately 0.15 cm as the radial dimension of a macropore.

A variety of other definitions of macropore size have been proposed (Table 1). Brewer (1964) reviewed different classifications and defined macropores as having a diameter of more than 75 μm. This followed a proposal by Johnson et al. (1960), who reported an apparent relationship between percolation rate and air content in soil at a pressure head of -40 cm (which corresponds to a pore diameter of 75 μm). It now appears that size as such is less important than pore continuity: small pores with a diameter of 40 μm can conduct considerable quantities of

The area of pore walls wet by a dye in the infiltrating water has been measured (Bouma and Dekker 1978), from which, if the pore widths are known, the volume of functional macropores can be calculated. Bouma and Wosten (1979), after microscopic examination of two swelling clay soils, calculated that the macropore volume was 0.014 and 0.024 $cm^3\ cm^{-3}$, of which about one-third was stained by methylene blue. Thus, even in a soil with a large number of macropores, relatively few of them may conduct water.

CONSEQUENCES OF MACROPORE FLOW

Infiltration into Soil with Continuous Macropores

Rainwater has been shown to move rapidly downward along continuous vertical cracks in soils, even when applied at relatively low intensities (Bouma and Dekker 1978; Bouma et al. 1978). In these studies, a solution of methylene blue was sprayed at different intensities and quantities onto field plots of 0.5 m^2 each. At first, infiltration occurred vertically into the upper surface of the peds (prisms) i_1 (Fig. 1). Flow into the cracks (c_f) started when the application rate exceeded the infiltration rate (at time t_1) for the intensity i_a. High application rates (i_b) resulted in earlier flow in cracks than did low rates. Flow in cracks may not occur at very low application rates (i_c). Surface ponding of water, ignored in this simplified example, would dramatically increase flow in cracks relative to the total amount of infiltrated water, although i_1, i_2 and c_f would all increase.

It is important that in these studies, flow into the cracks occurred only along shallow bands on vertical faces of the prisms. The total width of all bands per 10 cm depth interval below the plot was called the contact area S. The area S was a function of the application method and was about 1-2% of the possible contact area considering the total surface of the vertical prism faces. Deep penetration of water in the cracks was primarily due to a small S value. At higher S values, complete lateral adsorption would occur within a depth of a few centimeters. This conclusion was confirmed by Hoogmoed and Bouma (1980) using a simulation model for flow into cracked clay soil that used S values derived

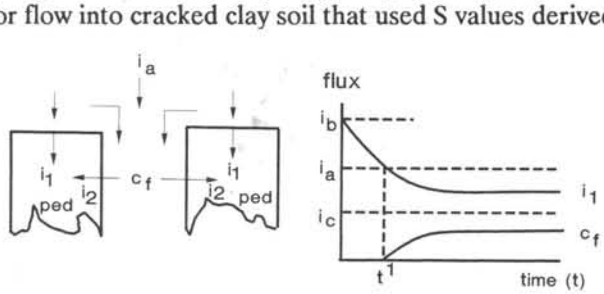

Figure 1. Schematic diagram of infiltration into cracked clay soil (from Bouma 1981).

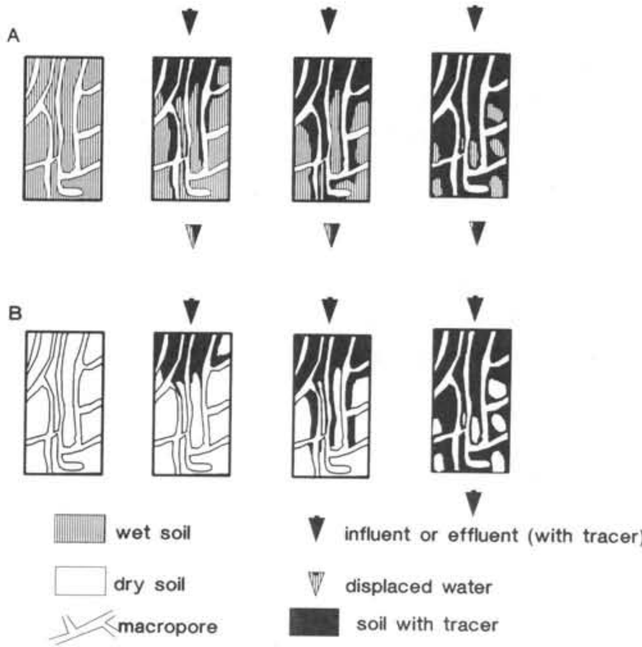

Figure 2. The effects of short-circuiting of water along air-filled macropores in unsaturated soil after application of a limited quantity of water. When the soil is initially very moist or wet (A) there is both displacement of the initially present water (gray arrow) and preferential movement of the influent along the macropores (not shown). Short-circuiting is relatively strong as compared with conditions when the soil is initially dry or slightly moist (B); then, there is no displacement but only infiltration at the soil surface and preferential vertical movement along the macropores and associated lateral infiltration (after Bouma 1977).

from independent soil morphological observations. The S values govern lateral infiltration into the peds and, therefore, the vertical depth of penetration of water into the cracks. Unfortunately, S values cannot be obtained from standard profile descriptions, but need to be measured in situ, such as through tracing of dye patterns.

When water is applied to soil in which at least some macropores are filled with air, a special case occurs when the rate of application exceeds the conductivity of the saturated matrix. The resulting pattern of flow is a complex function of the initial moisture content of the soil and the rate of application. There is essentially no mixing of inflow with the resident soil solution. This is illustrated in Fig. 2B for a dry soil with continuous macropores. Figure 2A illustrates vertical penetration of water in previously air-filled voids that is associated with vertical displacement of water initially present in the finer pores when the soil is initially moist or wet. In both cases there is rapid downward movement of "free" water under atmospheric pressure through air-filled macropores in unsaturated soil. This phenomenon has also been called "short-circuiting" (Bouma and Dekker 1978), and "non-matrix flow" (Bouma et al. 1980a). Short-circuiting in unsaturated soil is accompanied by lateral movement from the larger water-conducting pores into adjacent unsaturated soil. The rate of lateral movement is a function of the application rate, the hydraulic conductivity and the moisture content of the soil (Bouma and Anderson 1977; Bouma et al. 1978; Hoogmoed and Bouma 1980). This type of flow lends particular credibility to the conceptualization of macropore systems as being composed of two domains of flow, i.e., micropores and macropores, with exchange between them. This issue is discussed further under modeling.

Upward Flux in Clay Soils

Upward fluxes from the water table to the root zone are very important in compensating for the precipitation deficit in many clayey soils of the Northeast. Measurement of hydraulic conductivity (K) in sandy soils has allowed good estimates of these fluxes (Bouma et al. 1980b). However, results in clay soils have been poor (Bouma and De Laat 1981), due to the formation of horizontal cracks during drying of the clay. The measured conductivity is representative for the peds but not for the entire soil. Fluxes have the dimension of $m^3\ m^{-2}\ s^{-1}$, implying flow across the entire cross-sectional area. Obviously, vertical flow is not possible across a horizontal crack. To measure the effects of cracks upon flow, a block of soil (30 cm x 30 cm x 30 cm) was carved out in situ and covered with gypsum on five sides (Bouma et al. 1980a). After removal from the pit, the bottom surface was also covered and the block was put on one of its side walls. A solution of methylene blue in water was poured through the block in its new position: these surfaces were two opposite side walls in the original pit. Natural ped faces were then exposed at a given level by gently removing the peds and the stained area measured. If the area is x percent of the 900 cm^2 exposed surface, a "reduced" K value (K_r) is defined as

$$K_r\,(h = -y \text{ cm}) = K_m\,(h = -y \text{ cm}) \times \frac{100 - x}{100} \qquad (5)$$

Here, K_m is the hydraulic conductivity at the actual pressure head in the block of soil as measured with a tensiometer. The K_m measured by this technique is representative of the peds only. Several blocks were tested in identical soils with different pressure heads. Each test yielded one K_r value (Fig. 3). The degree of reduction of K_m may be different for different macrostructures and the technique can be used to characterize different types of structures in swelling clay soils. As such, the technique is an example of using soil

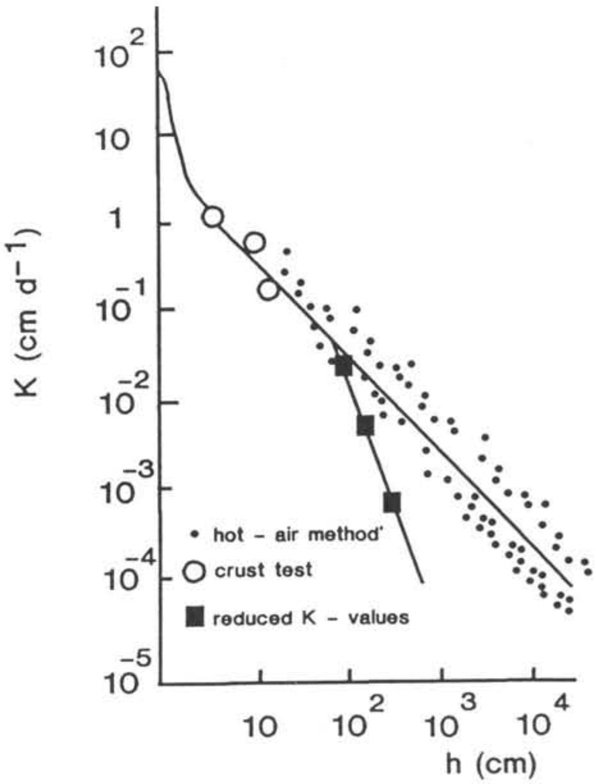

Figure 3. Measured hydraulic conductivity (K) for a heavy clay soil. The reduced K values are due to the formation of horizontal cracks upon drying. Cracks reduce the cross-sectional surface area which is available for vertical upward flow. The reduced area was estimated using a staining technique (after Bouma and De Laat 1981).

macromorphology for obtaining crucial boundary conditions for physical flow processes in soil with macropores.

Redistribution and Macropores in Deep Soils

It is not necessary that pores extend to the soil surface for macropore flow to occur, a phenomenon that has been explained by Thomas and Phillips (1979). When soils are plowed, for example, the macropores are disrupted in the upper 15 cm of soil, yet deep flow through macropores still occurs, though to a lesser extent than in an undisturbed soil (Quisenberry and Phillips 1978). Initiation of water movement in macropores at the interface of the tilled layer and the undisturbed soil below apparently is due to the relatively low conductivity of the undisturbed layer where many of the macropores are closed by shearing, smearing, and compacting by tillage. Water moving by Darcy's law through the plow layer of relatively high conductivity accumulates at the bottom of the tilled layer until the potential of the water reaches about zero. Some of the water then apparently enters a relatively few macropores through initially thin water films which become thicker as additional water moves into the macropores. This water then moves down the macropores by gravitational forces. A distinct difference in soil structure, such as that which occurs between the A and B horizons of many soils, will also cause initiation of water movement in macropores (Quisenberry and Phillips 1978). As shown by Aubertin (1971) and by Quisenberry and Phillips (1976), gravitational flow of water through macropores occurs readily in soils that are well below "field capacity." This flow has been described as "fingering." That is, the presence of macropores contiguous with the structural discontinuity results in water flow down the macropores, in the pattern of "fingers." This flow is produced by the localized unequal distribution of water at the structural interface (Starr et al. 1986; Parlange and Hill 1976), the resulting unequal distribution of pressure head, and the opportunity for flow down macropores even when the bulk soil is below saturation.

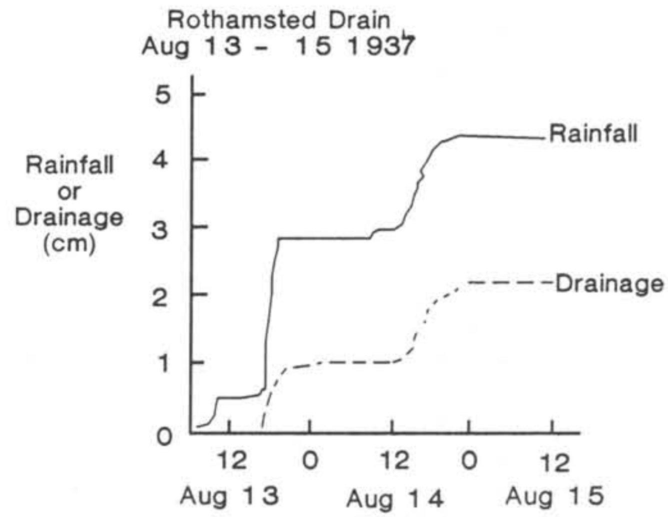

Figure 4. Cumulative rainfall and drainage from the 0.5-m Rothamsted drain in August 1937. The soil had not received rainfall for 3 weeks prior to this time yet drainage started as soon as a second storm began following a preliminary rainfall of 0.4 cm on 12 August (after Penman and Schofield 1941; as reported by Thomas and Phillips 1979).

Ground Water Recharge and Chemical Leaching

Water in macropores can move into or below the rooting depth in a matter of minutes after the addition of water to the soil surface (Quisenberry and Phillips 1976). This flow lasts no more than a few minutes or perhaps, in unusual cases, for a few hours after cessation of irrigation or rain. The effects of this type of flow on salt and water distribution have not been generally recognized. Some consequences are: (1) the value of a rain or irrigation to plants will generally not be so high as anticipated since some of the water may move below the root zone; (2) recharge of ground water can begin long before the soil reaches field capacity; (3) chemicals present in the added water will be moved to a much greater depth by a rain or irrigation than predicted by piston displacement; (4) chemicals resident in the pores of the bulk matrix may be bypassed and remain near the soil surface; and (5) it is not likely that water will carry a surge of contaminants to ground water at a time predictable by Darcian theory. Evidence for each of these consequences is presented below (from Quisenberry and Phillips 1976).

Observations at the Rothamsted drain in 1937 (Penman and Schofield 1941) demonstrated that a soil which had not received rain for 3 weeks produced no drainage after a short preliminary rain of 0.4 cm (Fig. 4). However, when rain began the second and third times, there was an immediate response in drainage. It is of note that these were the same drains used by Lawes et al. (1882). This immediate drainage has been recently observed as a common feature of well-structured soils.

Streams fed from well-structured soils in grass also show immediate responses to summer showers, even though base flow is declining and surface runoff does not occur. Flow from such a stream draining a watershed in bluegrass is shown in Fig. 5 (USGS 1971) for July 1971. The likely cause is water flow down the preferential pathway provided by a macropore. The often-heard complaint that thundershowers do not wet the soil as well as slow rains is probably valid, but not only because of surface runoff. Water can also be lost through macropores, leaving only a portion of the rain in the soil itself. While this is undesirable for plants growing in the soil, it is favorable for recharge of ground water, springs, and streams. Because all the added water does not stay in the soil, ground water aquifers are recharged before soils reach "field capacity."

If rain or irrigation water added to soil moves primarily through macropores, its interaction with solutes in the relatively immobile soil water of the bulk matrix will be limited. Evidence of Shuford et al. (1977) shown in Fig. 6 indicates that this is indeed the case. The concentration of NO_3-N after application of water is still highest near the soil surface and no increase has occurred at lower depths. This suggests that leaching occurs when NO_3 diffuses from the smaller pores to the surfaces of the macropores and then is moved through the soil profile. This is the same situation that van Genuchten and Wierenga (1976) described using the terms mobile and

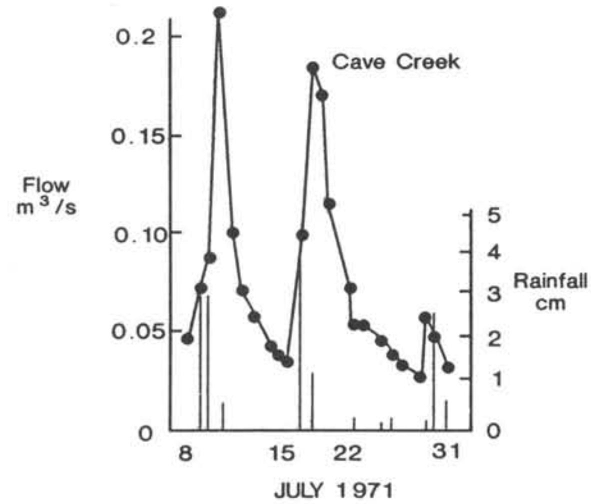

Figure 5. Drainage from Cave Creek in the central limestone areas of Kentucky during July 1979. Although the base flow is steadily dropping, each rainfall of consequence produces a peak on the hydrograph. No surface runoff occurred (after USGS 1971; as reported by Thomas and Phillips 1979).

immobile (stagnant) water. This is important in understanding chemical leaching and ground water contamination and provides one approach for models as discussed below.

Thus, the extent to which macropore flow increases or decreases the amount of chemical leached depends upon: (1) whether the chemical was recently applied or is resident on the exterior or within peds; (2) the ratio of macropore to matrix flow; (3) the saturated hydraulic conductivity of the matrix; (4) the antecedent water content of the soil; (5) the contact area between the bypass flow and the relatively static water of the matrix; and (6) the rate of chemical diffusion between the mobile and immobile water volumes.

Field Variability and Macropores

Variability of physical data collected in the field is often very large. Part of the variability in soils with macropores may be due to inappropriate measurements that assume that soils are rigid, homogeneous and isotropic. Predictions of soil moisture regimes in macropore systems are therefore often unsuccessful (Bouma and De Laat 1981). Other problems associated with the spatial variability of non-homogeneous soils and possible solutions are broadly discussed below for a number of cases. The publications cited provide additional details.

Only very large undisturbed samples yield representative measurements of the saturated conductivity of clay soils with macropores. Auger-holes may not be representative due to puddling of the hole, while small cores often yield very high values with high variability due to unrepresentative pore continuity patterns. Field variability is strongly reduced by using measurement sizes and volumes that include approximately 20 peds (Bouma 1979; Bouma et al. 1979; Lauren et al. 1987).

Soils with macropores always exhibit a strong drop in K as the water content decreases slightly below saturation. The measured K_{sat} of some Dutch clay soils was 50 cm/day, while K at h=-5 cm was only 1 cm/day (see Fig. 5). The drop is due to emptying of macropores. It is important that many existing unsteady state methods of measuring K_{unsat} do not allow determination of K_{unsat} at pressure heads higher than approximately -15 cm. Often, independently obtained values of K_{sat} are connected with those determined with unsteady state methods at h values lower than approximately -15 cm. Thus, unrealistic values are obtained for K_{unsat} near saturation which may not explain observed moisture conditions in the field. Steady-state methods such as the crust test would be more appropriate to obtain K_{unsat} values near saturation in soils with macropores.

As already discussed, preferential flow along macropores may fill unlined pores when these pores end

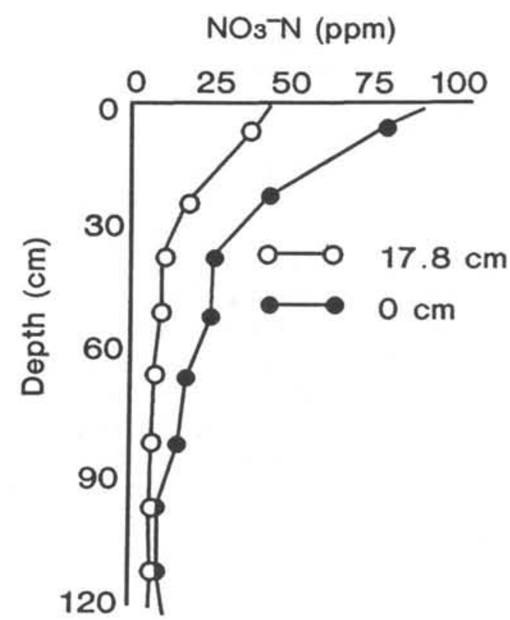

Figure 6. Distribution of NO_3-N in a Pennsylvania soil before and after leaching with 17.8 cm of water. The NO_3 was partially removed from all depths of soil by leaching, indicating a lack of complete water displacement (after Shuford et al. 1977).

inside unsaturated peds. Thus, the occurrence of perched water tables may be incorrectly suggested (Bouma et al. 1980c). The use of piezometers or tensiometers will avoid this problem.

In situ moisture contents measured with neutron or gamma probes are average values for a given volume of soil. These values do not reflect preferential patterns of water infiltration as discussed earlier. Selective gravimetric sampling of soil in a soil pit will provide more useful information.

Extraction of soil water with suction cups, which may or may not intercept continuous macropores, often yields spatially variable results (e.g., Shaffer et al. 1979). It is advisable to excavate the cups at the end of the experiments and to observe macropore patterns near the cups. Staining or tracing before excavation may be necessary. Thus, sub-populations can be distinguished among the data, which reduces the problems of interpretation.

Although spatial variability is of much concern in both monitoring and modeling studies, the effects of field variability can be reduced if the effects of macropores on flow are considered when measuring basic physical data, when monitoring soil profiles in situ, and when developing flow models for prediction purposes.

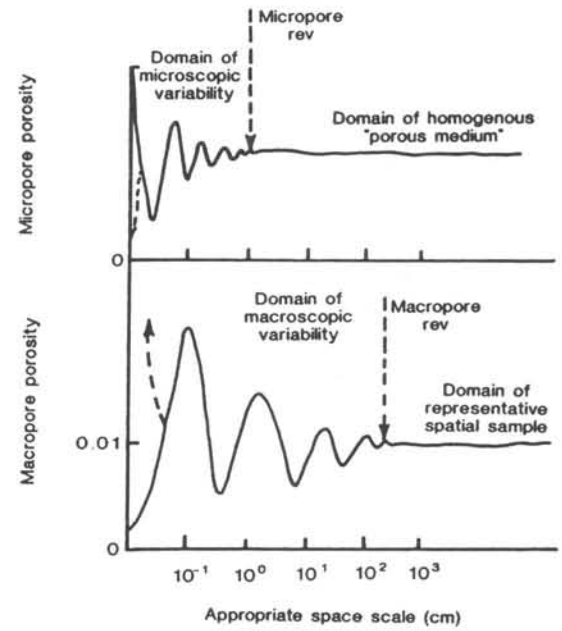

Figure 7. The variation of porosity with spatial scale and the definition of representative elementary volume (from Beven and Germann 1981).

MODELS OF MACROPORE WATER AND CHEMICAL MOVEMENT

Although staining techniques and morphological studies have provided insight into the physical structure of macropores, they have not yet led to quantitative prediction of water flow velocities through macropores. Velocities can be calculated from hypothetical macropore geometries, but these values are seldom observed in nature due to the discontinuity and irregular configuration of pore space, the effect of air entrapment on water entry into macropores, and the rather unlikely ponded conditions that are most conducive to maximum macropore flow. With such limitations, the mathematical descriptions of macropore flow that have been developed are numerous and ingenious, ranging from adaptions of the convection-dispersion equation for describing chemical movement to more empirical, though perhaps more useful, approaches. A contemporary review of modeling of macropore systems is presented by Germann (1989).

One of the most comprehensive models for describing flow of water in macropores was proposed by Beven and Germann (1981). They recognized that any model combining micropores and macropores should reduce to a Darcy-type model when no macropores are present. A domain concept was adopted (Fig. 7) with the micropore system being the domain that behaves according to classical hydraulic principles as summarized in Darcy's law. The macropores were recognized as a second domain requiring separate treatment. Linking the two domains and allowing their interaction in a physically realistic manner was also required. They approached the issue by describing the micropore system much as in Eq. (2), now written as

$$\frac{\partial \theta_{mi}}{\partial t} = - \frac{\partial}{\partial z}\left[K(\theta_{mi})\frac{\partial H}{\partial z}\right] + S \qquad (6)$$

where θ_{mi} is micropore water content and S is an exchange of water between micropores and macropores. The macropore system was assumed to be so complex that it could not be defined on theoretical grounds alone. Assumptions about macropore geometry and flow led to relationships between pore dimensions and macropore water flux. Interaction between the domains involved defining S in terms of a hydraulic gradient between micropore and macropore systems, where the gradient depends on the water contents in micropore and macropore regimes. The resulting model was capable of handling a wide range of conditions, and provides a good beginning for future efforts. Description of chemical movement was not included. A similar approach was used earlier by Edwards et al. (1979), but was less mechanistic and involved less rigorous consideration of the difference between the two domains.

The concept of two domains is implicit in other models of both water and chemical movement in macropores. Although there are a number of such models, the conceptual approaches can be contrasted with three examples. Van Genuchten and Wierenga (1976) showed that one quantitative method is to arbitrarily divide the soil water between "mobile" and "immobile" water with the proportions of each dependent upon soil properties and rate of flow. This concept was applied to chemical movement and models derived from it consist principally of analytical solutions to the convection-dispersion equation. These solutions assume a steady flow of water in both micropores and macropores, and require knowledge of a transfer coefficient between the two regimes. The approach works well for steady state conditions, so long as the transfer coefficient can be determined from the measured data. It has not been developed in detail for the transient water flow conditions of most field cases.

A similar approach was used by Addiscott (1979) to describe chloride and nitrate measured in the field in the same drain gauges used by Lawes et al. (1882). The Addiscott model assumes that the incoming water is successively partitioned between mobile and retained water. This model is much more empirical than the van Genuchten and Wierenga (1976) conceptualization, but also includes description of the water regime. The approach is useful in describing macropore flow, but as it is quite empirical, it does little to further the understanding of such systems.

A more useful approach is the model of White (1985a), which focuses on the prediction of nitrate leaching during unsaturated flow through a macropore soil. The model assumes that water infiltrating the soil surface mixes incompletely with resident water to form a mixed transport volume, the size of which can vary with time. Changes in chemical storage within the mixed transport volume can occur by diffusion and biological transformations. The mobile and immobile soil solution volumes into which the system is divided are analogous in concept to those of van Genuchten. The model utilizes a straightforward algebraic, bookkeeping approach to construct a description of chemical leaching between adjacent soil depth increments. The model was tested in experiments that describe measured nitrate concentrations in both the soil solution and in effluent from leached core samples. Although the results were generally encouraging, it was recognized that two parameters, the interfacial wetted area of the soil (ped surface wetted as water flows through interpedal voids) and the chemical diffusion coefficient, needed to be known to use the model. It remains to be seen whether this approach has long-term utility, although it appears promising.

The pronounced spatial variability of soil hydraulic properties often poses severe problems for the application of convection-dispersion and layer models to field soils. Where a large number of measurements of, for example, hydraulic conductivity and soil water content have been made at various sites in the field, the relationship has been found to vary by orders of magnitude. Identifying and condensing this variability by statistical or other means is often time consuming, requiring large numbers of samples and sophisticated models. Such limitations preclude the application of these models on a field basis, unless they are applied with recognition of field variability. A promising alternative to these models, the value of which is limited by how well the mechanisms of water flow and solute transport in macropore soils are understood, is provided by the transfer function model (Jury et al. 1982; Jury 1983). This approach requires only that the mass of solute transported be conserved and that the probability density function of solute travel times between the input and output surfaces in the soil be known. If the density function of travel times, or alternatively pore water velocities, conforms to a simple function such as a log-normal distribution, then the mean and standard deviation of that distribution are the essential parameters of the model. A log-normal distribution has been found to be appropriate for several chemicals (White et al. 1984; Biggar and Nielsen 1976; Jury et al. 1982). To extend this model further, more data are required on the distributions of pore water velocity in soils, especially those with macropores and under conditions when flow is not steady. The application of this approach to the leaching of indigenous soil solutes also needs to be explored.

SUMMARY AND CONCLUSIONS

Although classical approaches to description of water and solute movement have been proven useful and accurate in homogeneous soils, it is clear that other soils, often termed non-homogeneous soils, are characterized by two distinct domains of water flow. One is the well recognized bulk matrix, through which water moves according to Darcian principles. In some soils, due to the activities of earthworms, insects, or plant roots, or because of such morphological features as interpedal cracks, fissures or voids, a second flow domain exists. This macropore system is characterized by rapid, pipe-like water flow under certain conditions, which has led to its description as a preferred pathway, in which flow short-circuits the bulk matrix. The effect of such flow upon water status in the root zone, ground water recharge and fertilizer and pesticide movement can be substantial, but does not occur according to the mechanisms that are assumed operative in homogeneous soils. In fact, there is not yet a well established theoretical approach useful in predicting water and solute movement in such systems, although a variety of empirical and quasi-mechanistic models have been proposed. It is important that such predictive tools be developed and refined. Improved understanding of water and chemical movement, as well as better appreciation of the consequences of particular management practices, will result from the availability and application of more quantitative methods to macropore systems. It is important that research efforts in this area also continue to focus on the relationship of soil morphology and macropore flow dynamics, for this is clearly the path to not only a better understanding of non-homogeneous soils, but also will allow exploitation of the substantial data on soil morphology that have been developed over the last several years. Only through a concerted effort to pursue such issues with a balance of field observation, laboratory study and theoretical development will macropore soils be better understood and better managed.

LITERATURE CITED

Addiscott, T.M. 1977. A simple computer model for leaching in structured soils. J. Soil Sci. 28:554-563.

Addiscott, T.M., D.A. Rose and J. Bolton. 1978. Chloride leaching in the Rothamsted drain gauges: Influences of rainfall pattern and soil structure. J. Soil Sci. 29:305-314.

Anderson, J.C. and J. Bouma. 1973. Relationships between saturated hydraulic conductivity and morphometric data of an argillic horizon. Soil Sci. Soc. Am. Proc. 37:408-413.

Aubertin, G.M. 1971. Nature and extent of macropores in forest soils and their influence on subsurface water movement. For. Serv. Res. Pap. NE (U.S.) 192PS, 33 pp.

Bear, J. 1972. Dynamics of Fluids in Porous Media. Elsevier, NY.

Beasley, R.S. 1976. Contribution of subsurface flow from the upper slopes of forested watersheds to channel flow. Soil Sci. Soc. Am. J. 40:955-957.

Beven, K.J. 1980. The Grendon Underwood field drainage experiment. Rep. 65, Inst. Hydrol., Wallingford, UK, 30 pp.

Beven, K.J. 1981. Micro-, meso-, macroporosity and channeling flow in soils. Soil Sci. Soc. Am. J. 45:1245.

Beven, K.J. and P. Germann. 1981. Water flow in soil macropores. 1. A combined flow model. J. Soil Sci. 32:15-29.

Beven, K.J. and P. Germann. 1982. Macropores and water flow in soil. Water Resour. Res. 18:1311-1325.

Biggar, J.W. and D.R. Nielsen. 1976. Spatial variability of the leaching characteristics of a field soil. Water Resour. Res. 12:78-84.

Blake, G., E. Schlichting and U. Zimmermann. 1973. Water recharge in a soil with shrinkage cracks. Soil Sci. Soc. Am. Proc. 37:669-672.

Bouma, J. 1981a. Comment on "Micro-, meso-, and macroporosity of soil." Soil Sci. Soc. Am. J. 45:1244-1245.

Bouma, J. 1981b. Soil morphology and preferential flow along macropores. Agric. Water Mgmt. 3:235-250.

Bouma, J. and J.L. Anderson. 1977. Water and chloride movement through soil columns simulating pedal soils. Soil Sci. Soc. Am. J. 41:766-770.

Bouma, J. and L.W. Dekker. 1978. A case study on infiltration into dry clay soil. I. Morphological observations. Geoderma 20:27-40.

Bouma, J. and P.J.M. De Laat. 1981. Estimation of the moisture supply capacity of some swelling clay soils in the Netherlands. J. Hydrol. 49:247-259.

Bouma, J. and J.H.M. Wosten. 1979. Flow patterns during extended saturated flow in two undisturbed swelling clay soils with different macrostructures. Soil Sci. Soc. Am. J. 43:16-22.

Bouma, J. L.W. Dekker and J.H.M. Wosten. 1978. A case study on infiltration into dry clay soil. II. Physical measurements. Geoderma 20:41-51.

Bouma, J., A. Jongerius and D. Schoonderbeek. 1979. Calculation of hydraulic conductivity of some saturated clay soils using micromorphometric data. Soil Sci. Soc. Am. J. 43:261-265.

Bouma, J., R.F. Paetzhold and R.B. Grossman. 1980a. Application of hydraulic conductivity measurements in soil survey. Soil Cons. Serv. USDA. Soil Survey Investigation Rep., Washington, DC, 12 pp.

Bouma, J., P.J.M. De Laat, R.C.H.M. Awater, A.F. Van Holst, H.C. Van Heesen and Th.J.M. Van de Nes. 1980b. Use of soil survey data in a simulation model for predicting regional soil moisture regimes. Soil Sci. Soc. Am. J. 44:808-814.

Bouma, J., L.W. Dekker and J.C.F.M. Haans. 1980c. Measurement of depth to water table in a heavy clay soil. Soil Sci. 130:264-270.

Brewer, R. 1964. Fabric and Mineral Analysis of Soils. John Wiley, NY.

Buckingham, E. 1907. USDA Bur. of Soils Bull. 38, U.S. Government Printing Office, Washington, DC.

Bullock, P. and A.J. Thomasson. 1979. Rothamsted studies of soil structure. 2. Measurement and characterization of macroporosity by image analysis and comparison with data from water retention measurements. J. Soil Sci. 30:391-414.

Darcy, H. 1856. Les fountaines publique de Ville de Dijon. Dalmont, Paris.

Edwards, W.M., R.R. Van der Ploeg and W. Ehlers. 1979. A numerical study of the effects of noncapillary-sized pores upon infiltration. Soil Sci. Soc. Am. J. 43:851-856.

Gaiser, R.N. 1952. Root channels and roots in forest soils. Soil Sci. Soc. Am. Proc. 16:62-65.

Gardner, W. 1920. The capillary potential and its relation to soil moisture constants. Soil Sci. 10:357-359.

Germann, P. 1989. Macropores and hydrologic hillslope processes. *In* Anderson and Burt (eds.) Surface and Subsurface Processes in Hydrology. John Wiley and Sons, NY.

Germann, P. and K.J. Beven. 1981. Water flow in soil macropores. 1. An experimental approach, J. Soil Sci. 32:1-13.

Green, R.D. and G.P. Askew. 1965. Observations on the biological development of macropores in soils of Romney Marsh. J. Soil Sci. 16:342.

Hole, F.D. 1981. Effects of animals on soils. Geoderma 25:75-112.

Hoogmoed, W.B. and J. Bouma. 1980. A simulation model for predicting infiltration into cracked clay soil. Soil Sci. Soc. Am. J. 44:458-462.

Johnson, W.M., J.E. McClelland and S.A. McCaleb. 1960. Classification and description of soil pores. Soil Sci. 89:319-321.

Jury, W.A. 1983. Chemical transport modelling: Current approaches and unresolved problems. pp. 49-62. *In* D.W. Nelson, D.E. Elrick and K.K. Tanji (eds.) Chemical mobility and reactivity in soil systems. SSSA Spec. Publ. No. 11.

Jury, W.A., L.H. Stolzy and P. Shouse. 1982. A field test of the transfer function model for predicting solute transport. Water Resour. Res. 18:369-375.

Lauren, J.G., R.J. Wagenet, J. Bouma and H. Wosten. 1987. Variability of saturated hydraulic conductivity in a glossaquic hapludalf with macropores (in press). Soil Sci.

Lawes, J.B., J.H. Gilbert and R. Warington. 1882. On the amount and composition of the rain and drainage water collected at Rothamsted, Williams, Glowes and Sons Ltd., London.

Lewis, D.T. 1977. Subgroup designation of three udolls in southeastern Nebraska. Soil Sci. Soc. Am. J. 41:940-945.

Luxmoore, R.J. 1981. Micro-, meso- and macroporosity of soil. Soil Sci. Soc. Am. J. 45:671.

Marchall, T.J. 1959. Relation between water and soil. Tech. Comm. 50, Commonwealth Bur. of Soils. Harpenden, UK.

McDonald, P.M. 1967. Disposition of soil moisture held in temporary storage in large pores. Soil Sci. 103:139-143.

Mosley, M.P. 1982. Subsurface flow velocities through selected forest soils, South Island. J. Hydrol. 55:65-92.

Nelson, W.R. and L.D. Baver. 1970. Movement of water through soils in relation to the nature of the pores. Soil Sci. Soc. Am. Proc. 5:69-76.

Nielsen, D.R., J. Metthey and J.W. Biggar. 1981. Soil hydraulic properties, spatial variability and soil water movement. pp. 47-68. *In* I.K. Iskander (ed.) Modeling wastewater renovation. John Wiley and Sons, NY.

Nielsen, D.R., M.Th. van Genuchten and J.W. Biggar. 1986. Water flow and solute transport processes in the unsaturated zone. Water Resour. Res. 22:895-1085.

Omoti, U. and A. Wild. 1979. Use of fluorescent dyes to mark the pathways of solute movement through soils under leaching conditions. 2. Field experiments. Soil Sci. 128:98-104.

Parlange, J.-Y. and D.E. Hill. 1976. Theoretical analysis of wetting front instability in soils. Soil Sci. 122:236-239.

Penman, H.L. and R.K. Schofield. 1941. Drainage and evaporation from fallow soil at Rothamsted. J. Agric. Sci. 31:74-109.

Quisenberry, V.L. and R.E. Phillips. 1976. Percolation of surface-applied water in the field. Soil Sci. Soc. Am. J. 40:484-489.

Quisenberry, V.L. and R.E. Phillips. 1978. Displacement of soil water by simulated rainfall. Soil Sci. Soc. Am. J. 2:675-679.

Reeves, M.J., D.G.M. Hall and P. Bullock. 1980. The effect of soil composition and environmental factors on the shrinkage of some clayey British soils. J. Soil Sci. 31:429-442.

Reeves, M.J. 1980. Recharge of the English chalk. A possible mechanism. Eng. Geol. 14:231-240.

Richards, L.A. 1931. Capillary conduction of liquids in porous mediums. Physics 1:318-333.

Ritchie, J.T., D.E. Kisel and E. Burnet. 1972. Water movement in undisturbed swelling clay soils. Soil Sci. Soc. Am. Proc. 36:874-879.

Schumacher, W. 1864. Die Physik des Bodens, Berlin.

Shaffer, K.A., D.D. Fritton and D.E. Baker. 1979.

Drainage water sampling in a wet, dual-pore soil system. J. Env. Qual. 8:241-246.

Shuford, J.W., D.D. Fritton and D.E. Baker. 1977. Nitrate-nitrogen and chloride movement through undisturbed field soil. J. Env. Qual. 6:736-739.

Starr, J.L., H.D. De Roo, C.R. Frink and J.-Y. Parlange. 1978. Leaching characteristics of a layered field soil. Soil Sci. Soc. Am. J. 42:386-391.

Starr, J.L., J.-Y. Parlange and C.R. Frink. 1986. Water and chloride movement through a layered field soil. Soil Sci. Soc. Am. J. 50:1384-1390.

Thomas, G.W. and R.E. Phillips. 1979. Consequences of water movement in macropores. J. Env. Qual. 8:149-152.

Topp, G.C. and J.L. Davis. 1981. Detecting infiltration of water through soil cracks by time-domain reflectometry. Geoderma 26:13-23.

U.S. Geological Survey. 1971. Water resources data for Kentucky. U.S. Geological Survey, Reston, VA, 75 pp.

van Genuchten, M.Th. and W.J. Alves. 1982. Analytical solutions of the convective-dispersive solute transport equation. USDA Technical Bulletin No. 1601, U.S. Salinity Laboratory, Riverside, CA.

van Genuchten, M.Th. and P.J. Wierenga. 1976. Mass transfer studies in sorbing porous media. I. Analytical solutions. Soil Sci. Soc. Am. J. 40:473-480.

White, R.E. 1985a. A model for nitrate leaching in undisturbed structured clay soil during unsteady flow. J. Hydrol. 79:37-51.

White, R.E. 1985b. The influence of macropores on the transport of dissolved and suspended matter through soil. pp. 95-120 (Vol. 3) *In* B.A. Stewart (ed.) Adv. in Soil Science, Springer-Verlag.

White, R.E., G.W. Thomas and M.S. Smith. 1984. Modeling water flow through undisturbed soil cores using a transfer function model derived from ^3HOH and Cl transport. J. Soil Sci. 35:159-168.

Estimating Parameters Controlling Saturated and Unsaturated Flow in Soils

A.S. Rogowski

Soil Scientist
Northeast Watershed Research Center
Agricultural Research Service, USDA
University Park, PA 16802
and
Adjunct Professor of Agronomy
The Pennsylvania State University
University Park, PA 16802

Contribution from the U.S. Department of Agriculture, Agricultural Research Service, in cooperation with the Pennsylvania Agricultural Experiment Station, The Pennsylvania State University, University Park, Pennsylvania

CONTENTS

INTRODUCTION

Preventing contamination of ground water from agriculture requires an accurate assessment of the pollution potential of the manure, fertilizer and pesticides that are regularly applied to the land. To aid such an assessment, pertinent parameters need to be identified and appropriate sampling and modeling techniques developed. The potential for ground water pollution should also be related to measurable chemical properties of soil. These properties can be expressed as quantity, capacity, and intensity factors (Baker and Amacher 1981). The quantity factor (Q) is the absolute amount of a chemical that can be leached by infiltrating water or taken up by plant roots. The intensity factor (I) describes the availability of the chemical to plant roots or for transport. The capacity factor (dI/dQ) indicates the change in intensity or quantity with a corresponding change in the other. If the availability of one chemical is influenced by the presence of another, the relative intensity factor must also be considered. The distribution of these factors in the zone between soil surface and ground water and their interaction with saturated or unsaturated water flow will determine the amount of ground water pollution. It should be remembered that water flow itself will also vary spatially and over time, subject to the external influences of weather and the internal geometry of the soil.

Quantitative knowledge of soil properties is required for prediction of hydrologic behavior (Sharma and Rogowski 1985). The actual approach is dictated by the magnitude of spatial variability and the distribution of hydrological properties. Which properties should be measured, what sample volume of soil should be taken at what location and at what frequency are some of the aspects that must be determined. Of prime importance are the objectives, the desired accuracy of the predictions, and the availability of funds, all of which will influence the level of sophistication and detail at which a watershed is sampled and data are analyzed.

Soil Classification

Soil classification is based on the premise that soil properties vary in space and soil surveys are used to identify and delineate the soil boundaries. But in most classification schemes, these boundaries are imprecisely defined and the properties themselves are assigned or approximated. The soils of a watershed may then be grouped into soil series with emphasis on profile similarities. The soil survey classification is based on the broad morphological features of the landscape and correlated with sampled profile properties such as color, horizon, depth, structure, and texture. However, the extent and nature of variability and mapping purity within a soil unit are not always recorded. As a first approximation, properties of these soil series may be used in hydrologic modeling, keeping in mind that they are generally known as "soft data."

However, the criteria used in classifying soils may not coincide with those affecting hydrologic response of an area. Furthermore, appreciable spatial variability in soil hydrological properties has been observed within a soil series (Rogowski 1972; Nielsen et al. 1973; Sharma et al. 1980; Vieira et al. 1982; Webster 1985; Philip 1986), which may affect the areal response (Sharma and Luxmoore 1979). Thus, under many conditions, field characterization of watersheds is important. Methods for quantifying soil hydrological properties in the field should be simple, rapid, and reliable so that a large number of measurements can be made. In general, grid- or transect-sampling schemes are preferred.

Desired Approach

At this point we pause to describe our approach. Our objective is to develop a methodology for evaluating the impact of agriculture on ground water quality, particularly in the northeastern United States. We may proceed in one of three ways. First, we can simplify the problems, solve them analytically, and examine practical implications. This approach is favored by Philip (1987).

Second, we could solve the problems numerically using limited data, simplified assumptions, and deterministic or stochastic modeling. This approach is now generally favored in soil science and is widely used by regulators and consultants. Third, we could infer ground water pollution potential based on a modest soil and aquifer sampling campaign, available "soft" data, and primary transport pathway identification.

This last approach will be discussed here. It is similar to the Transfer Function Model (TFM) of Jury (1982) and the geographical information systems approach of Evans and Myers (1986). But it operates on a larger scale, incorporating possible landform effects, delineation of preferential-flow pathways, probability kriging, and geographical information systems with readily available or obtainable data. The approach may lack the elegance of analytical and numerical solutions or the scope and flexibility of modeling efforts. However, its purpose is to describe the real world as accurately as possible using methodologies readily understandable and available.

SCALE OF OBSERVATIONS

Any practical evaluation of ground water pollution potential must simultaneously consider a series of events that occur at different time and space scales. While quantity, intensity, and capacity factors may describe the chemical pollution potential at a point, water and solute transport mechanisms are active over a volume. Transport of solutes occurs on a pore-sized scale within the soil matrix and through larger cracks, fissures, and macropores which may often short-circuit the bulk of the profile. Environmental impacts generated on a local scale by discrete points within individual profiles may, when taken together, affect aquifer quality on a substantially larger regional scale. But fluxes readily observable on a plot or in the laboratory may be indiscernible in the field. Conversely, short-circuiting, although characteristic of many field locations, can seldom be observed in the laboratory and is not easily modeled in the field. So by the time contamination of the aquifer is detected, extensive and expensive clean-up procedures are usually required.

It would therefore be preferable to develop acceptability criteria that can be applied to individual sites based on general water flow distribution and the quantity (Q), intensity (I), and capacity (dI/dQ) factors. Unfortunately, this approach requires extensive data and computational facilities as well as a conceptual framework that is not yet entirely in place. The extensive heterogeneity of natural systems and our inability to decide what information is needed are roadblocks to developing these criteria.

Measures of Variability

Observed variability may be expressed on a characteristic length scale (Dagan 1986) that approximates an average distance over which particular properties are correlated. Alternately, variogram analyses (Journel 1986) can be used to assess the extent and structure of variability for each scale. Suppose that a field is sampled for some property on a regular grid. A plot of the average sum of differences squared at each lag distance that separates respective sample values is known as a *semivariogram* [γ(h)],

$$\gamma(h) = \frac{1}{2N} \sum_{i=1}^{N} \left[Z(x_i) - Z(x_i + h) \right]^2 \qquad (1)$$

where N is the number of pairs [$Z(x_i)$, $Z(x_i+h)$] for a particular distance (or time) increment h. A semivariogram may be thought of as a variance of differences given as a function of separation distance.

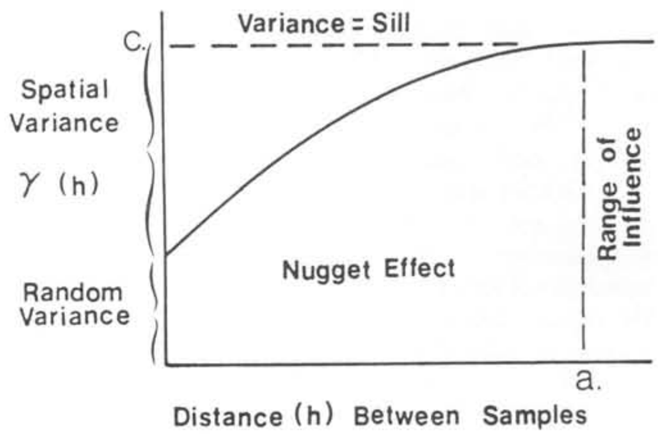

Figure 1. Idealized semivariogram for a soil property showing sill (c), range (a) and nugget effect.

Semivariograms can be simple or complex. Simple semivariograms with sills generally operate on a single variability scale (Fig. 1) and will show a distance (range a) that denotes extent of continuity and an *a priori* variance value (sill C) as a function of sample size. Complex or nested semivariograms can span several variability scales and display (Fig. 2) specific structures that may denote extent of continuity (range a) and magnitude of variability (sill C) at each scale.

Often γ (h) does not pass through the origin at h = 0 and has some positive value called a *nugget effect* (C_0). A nugget effect usually suggests spatial structure on a scale smaller than the sampling interval. This spatial structure can be delineated by sampling at shorter intervals.

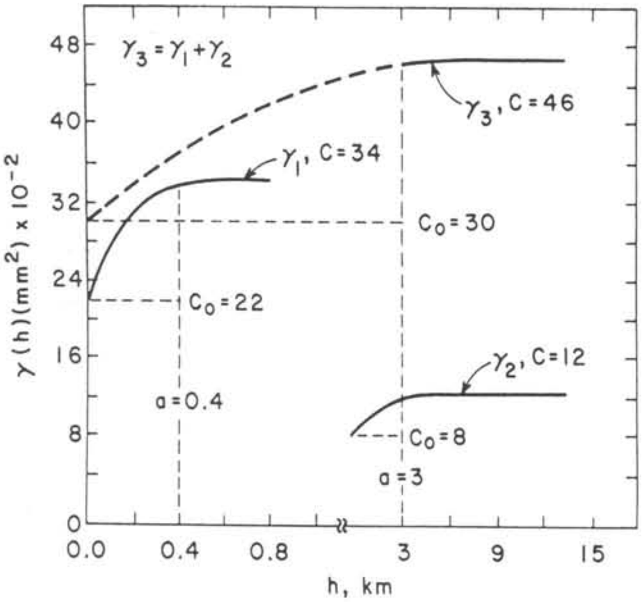

Figure 2. Complex semivariograms of soil loss, for 0.1-km (γ_1), and 1.5-km spacings (γ_2), and for a combined nested model (γ_3); values of sill (C) and nugget effect (C_0) are in $mm^2 \times 10^{-2}$, values of range (a) are in km (from Rogowski et al., 1985a).

Occasionally, for a pure nugget effect variogram, a different sampling scheme on a larger scale may identify the underlying structure of the variability.

Meaningful structural analysis of such data involves combining concepts from soil science and geostatistics. In soil classification, for example, we may expect a certain amount of variability on a plot or field scale. These are referred to as *series' variants* or *inclusions*. For an entire farm or township, a different level of variability (γ_1) may become apparent that is related to differences in landforms. On a regional scale differences (γ_2) due to gross changes in topography and geology come into play. All these sources of variability act together and can appear as nested structures γ_3 (Fig. 2) if sampling has been sufficiently extensive. Insights gained by variogram analysis can then be used to specify and measure only those variables that are pertinent to the scale of interest.

With extensive sampling, preferably on a regular grid, primary sources of variability on each scale can be identified. But what happens if we analyze a given attribute based on a larger volume? Then we must consider the geostatistical notion of support, which is similar to a representative elementary volume (REV) and is illustrated by the following example.

Figure 3 shows three semivariograms of bulk density measured on different volumes of soil. The topmost variogram is based on a relatively small volumeter sampler, and the second and third variograms are for

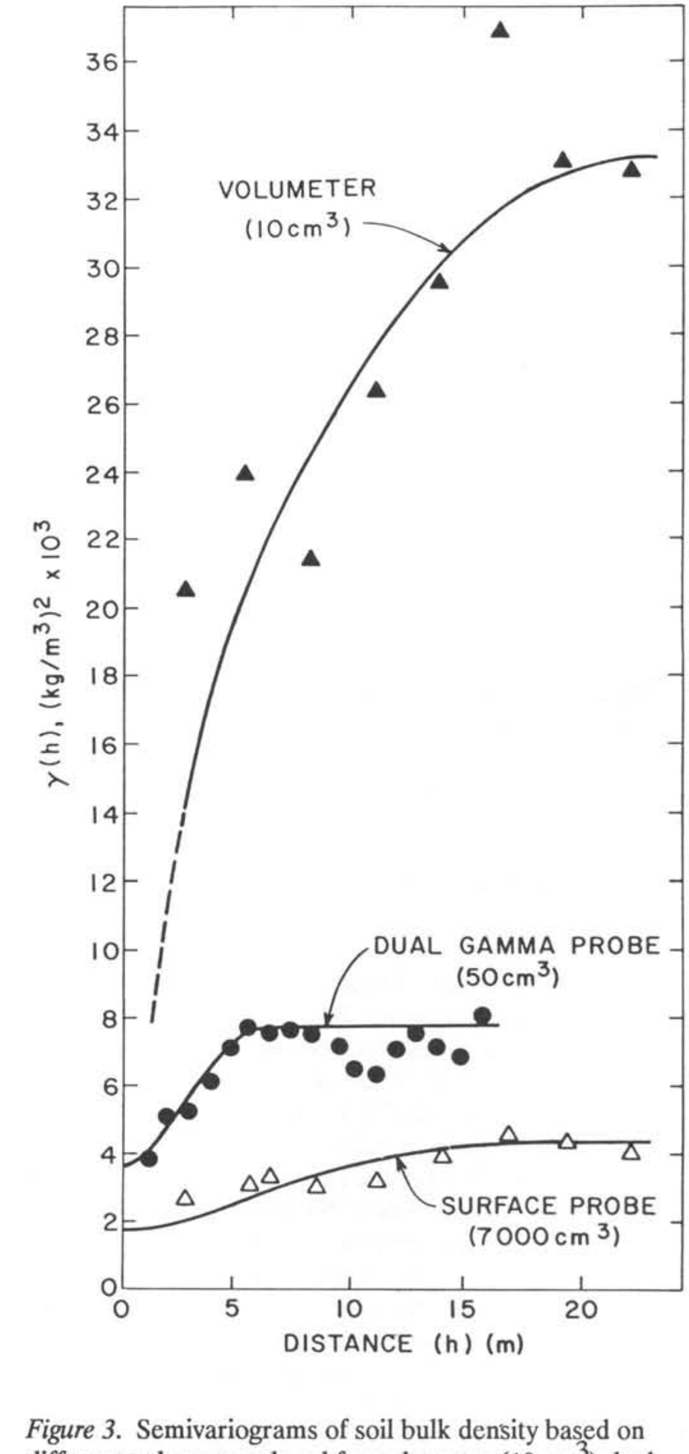

Figure 3. Semivariograms of soil bulk density based on different volumes analyzed for volumeter (10 cm^3) dual gamma probe (50 cm^3), and surface density probe (7000 cm^3) (from Rogowski et al., 1985a).

dual gamma and surface-density probes. It appears that samples of volumeter size represent point values, while dual gamma and surface probe samples reflect decreasing variability because of the larger volumes over which the density is averaged. The intrinsic variability of a property at a point should remain about the same, but, when averaging over larger and larger volumes, apparent variability decreases. This point needs to be kept in mind when using the REV approach.

Thus it is important to ascertain the sample size at which the variability of the data manifests itself, keeping in mind that comparisons between data sets can only be made at a point. Then one can decide on what scale the problem at hand requires a solution and consider only those aspects of variability that pertain.

Our inability to conceptually formulate the risks of nonpoint source pollution stems from averaging measured or simulated values over larger and larger areas without a good rationale. It may therefore be worthwhile to first separate the potential impacts into the respective scales at which they operate. For example, local impacts from highly continuous zones should be considered apart from impacts at a landform level or regional scale. Furthermore, variability at a point should not be confused with apparent decline in variability as larger volumes are averaged. Practical implications of this approach have recently been illustrated by Parkin (1987), who found that denitrification "hot spots" in soil are associated with particulate organic C, rather than being a function of volume alone.

WATER FLOW REGIME

The vadose zone (Nielsen et al. 1986), also known as the unsaturated or partially saturated zone, may be thought of as that portion of the soil profile, extending laterally, that is bounded above by the soil surface and below by ground water. Water in that zone generally occupies less than the total available porosity and is considered to be under tension (or negative potential). The vadose zone includes the capillary fringe above the water table as well as areas of intermittent saturation due to surface ponding and temporary, perched water tables. Consequently, any cracks, channels, or macropores are an integral part of the vadose zone.

The forces acting on the water in this zone (pressure, solute, electrochemical, and gravitational potentials), coupled with proportionality coefficients in the form of hydraulic conductivity, are generally used to describe flow and transport. Following Nielsen et al. (1986), Darcy's Law may be extended to unsaturated conditions.

$$q = \sum_i K_i \frac{\delta \psi_i}{\delta z}$$

Or if (2)

$$h = \psi_p / g$$

where g is the acceleration of gravity, and ψ_p is the pressure potential,

$$q = K(h) \frac{\delta h}{\delta z} \quad \text{(see Fig. 4).} \tag{3}$$

Under the assumptions of continuity, the change of water content (θ) with time (t) is equal to the change of flux density (q) with depth plus contributions from any potential sources or sinks (ϕ) of water in the system.

$$\frac{\delta \theta}{\delta t} = \frac{\delta q}{\delta z} + \phi \tag{4}$$

Subsequent to the application of the continuity assumption, Darcy's Law becomes Richards' equation (Richards 1931),

$$\frac{\delta \theta}{\delta h}\frac{\delta h}{\delta t} = \frac{\delta}{\delta z}\left[K(h)\frac{\delta h}{\delta z}\right] - \frac{\delta K(h)}{\delta z} + \phi \tag{5}$$

In Richards' equation the proportionality coefficients of Darcy's Law are in a form of hydraulic conductivity given as a function of total potential. Compressibilities of solute and matrix are generally ignored and solute density is assumed to be independent of concentration and location. While Richards' equation is a convenient conceptualization of a homogeneous small-scale flow system, it is nonlinear, which makes it difficult to solve. It may not hold for low-flow rates in fine-textured soils or in structured materials and soils with channels, macropores, or high coarse-fragment content (Nielsen et al. 1986). To apply Darcy's Law or Richards' equation, we must assume

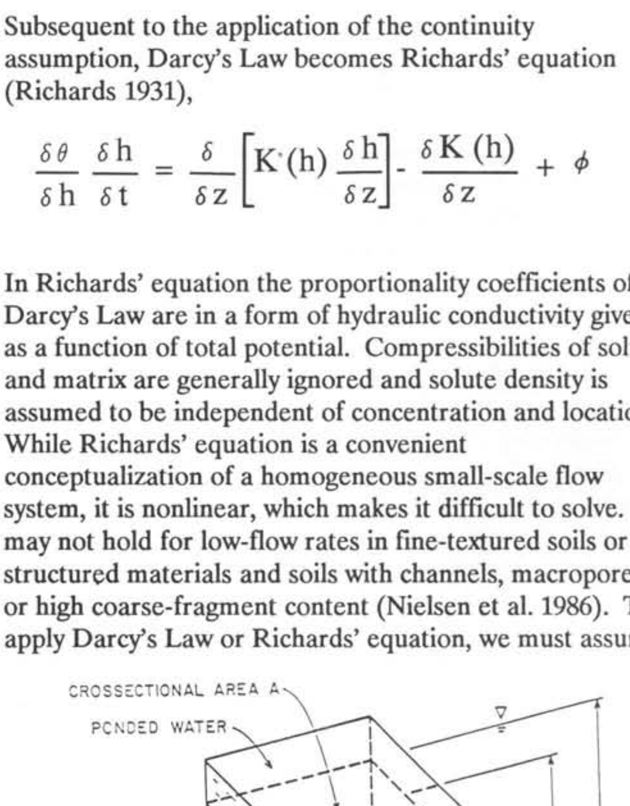

Figure 4. Schematic illustration of Darcy's Law (Darcy, 1856) applied to an imaginary block of soil.

that time and space averaging is permissible and representative of a location or volume of soil, since values of hydraulic conductivity and total and component potentials vary in space and time. If values differ from point to point, we also have to choose an arbitrary dividing line between one location and the adjoining one. Such choices and assumptions are liable to be in error. If they are not in error, the computed solute transport to a water table will at least be relatively slow.

Slow transport does not always correspond to reality, since rapid transfer of water and solutes is frequently observed even under unsaturated conditions. Such rapid flows are usually ascribed to the presence of preferential flow pathways or macropores (White 1985), which may comprise only 0.1 to 5% of the cross-sectional area but may account for nearly all of the observed flow and transport.

MACROPORE FLOW

Seepage into Cavities

The general background relating to preferential pathway or macropore flow is aptly presented by Wagenet and Germann (1989). Here I will attempt to introduce some additional findings and discuss some alternative aspects of this problem.

Conventional thinking on preferential flow pathways during unsaturated flow in the vadose zone is that, to take part in flow, macropores must extend to the soil surface where there must be a source of free water. Recent theoretical studies in Australia (Philip 1987) suggest that under steady downward seepage in uniform soil there may be a buildup of water pressure at the walls of a buried hole or cavity which may not necessarily extend to the soil surface. The condition for seepage into the cavity is that pressure at some point of the cavity wall reaches atmospheric. Similar problems may arise whenever cavities are replaced by stones or rocks. It appears that flat-topped cavities are not particularly efficient at excluding flux and will admit water from unsaturated seepage quite readily. We may also expect water to pond and flow around flat stones and rocks, fertilizer granules, and organic debris. Such selective flows, which could be particularly significant in soils in the northeastern United States because of their high coarse-fragment content, may concentrate leachate into preferential flow pathways even

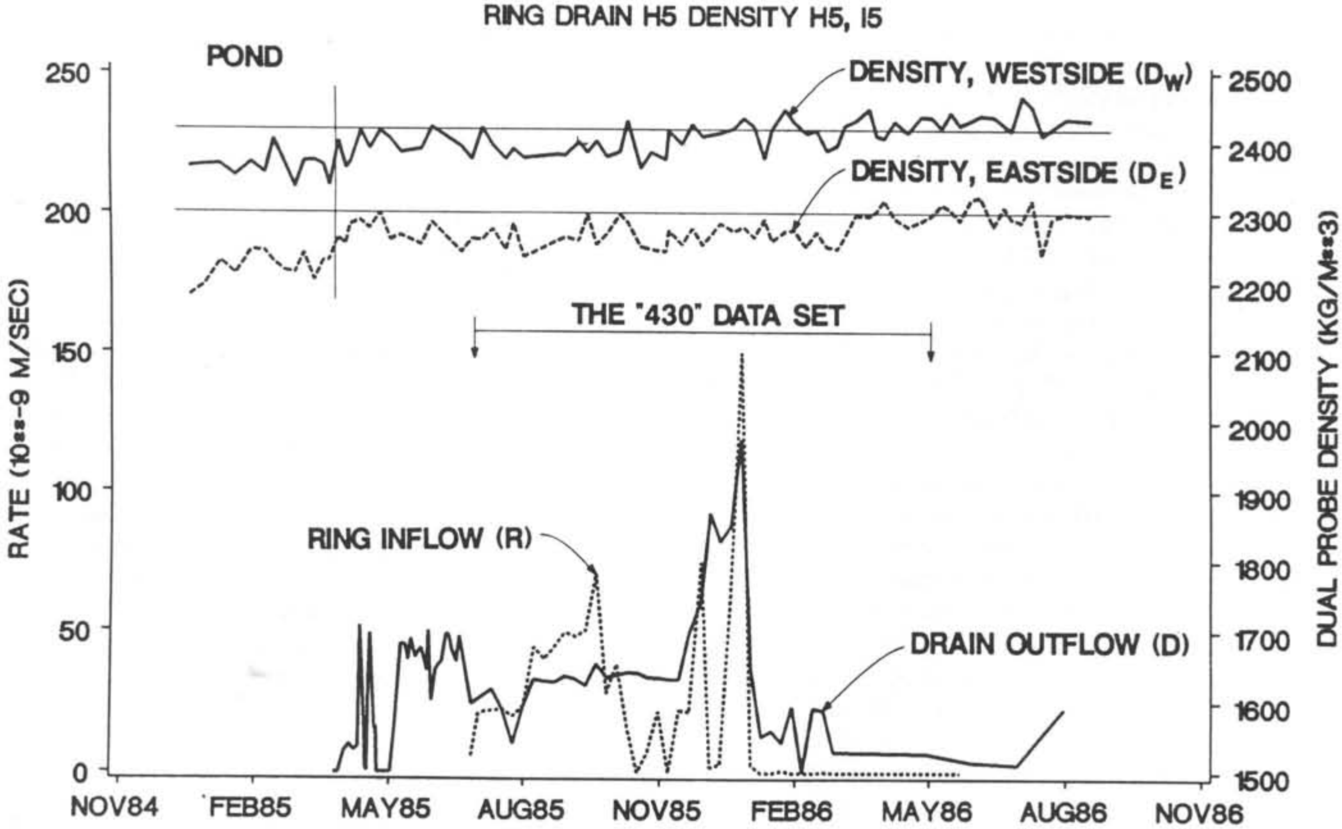

Figure 5. Distribution in time of ring inflow (R), and drain outflow (D) at a monitoring site H5, and of bulk density at adjoining sites to the west (D_W) and east (D_E) of the ring/drain site (from Rogowski et al., 1985a).

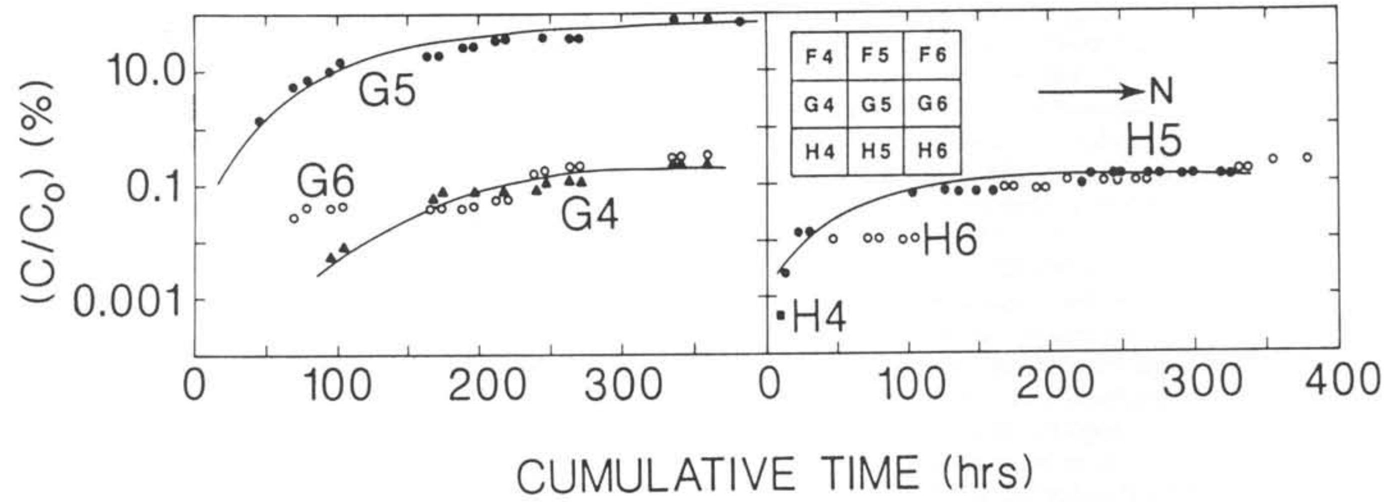

Figure 6. Relative concentration of Br^- in leachate from site G5 and surrounding drains (from Rogowski, 1987).

under unsaturated conditions. Accumulated solutes could then be flushed down at high rates following widespread rainfall.

Tracer Studies

Tracer and dye studies are useful for evaluating distribution of preferential flow pathways. For example, the movement and distribution of tracer Br^- have been found to approximate the movement and distribution of the pesticide Aldicarb (Brasino 1986). Examples used here are discussed in detail in Rogowski (1986a,b) and Rogowski and Simmons (1988). They are based on a percolation study in which a 0.3-m thick clay layer B horizon of a Typic Hapludult was compacted on an elevated 9 x 23 m platform and ponded with water for one year. Despite ponding, the flow of water occurred under unsaturated conditions. Inflow, outflow, and bulk density measurements were taken continuously on a 0.9 x 0.9 m square grid at 184 locations.

Figure 5 illustrates flow and bulk density (Rogowski 1988) over time for selected locations where rapid breakthrough of water or tracer have been either observed or postulated. These locations are in zones of high flow; therefore, the observed response could reflect a history of a preferential flow path at a point. Figure 5 illustrates typical flow (lower) and bulk density (upper) patterns observed. The ring (R) inflow and drain (D) outflow rates (left-hand scale) are in 10^{-9} m/sec, while bulk density values, measured east (D_E) and west (D_W) of the respective ring and drain locations, are in kg/m^3 (right-hand scale). Thin horizontal lines near bulk density curves are included to highlight any trend in density data, while a thin vertical line indicates time of ponding. The "430 data set" refers to a time interval of ten months over which highly variable flows were averaged to obtain a single representative value.

At this site (H5) rapid breakthrough of Br^- tracer applied to the central (G5) ring (Fig. 6) was observed after 22 hours with a very high apparent flow rate of 10,000 x 10^{-9} m/sec. Since the observed outflow was no more than 10 x 10^{-9} m/sec, the effective porosity through which tracer passed could be as little as 0.1% of the local cross-sectional area.

These results are illustrated in Fig. 6, which shows the relative concentration of Br^- tracer in drain leachate as a function of time after the tracer was applied to a central ring. Density and flow records for the site and breakthrough history at the site, as well as order of magnitude differences between lab and field values (Rogowski 1988), suggest a highly variable "neighborhood" with an effective porosity ranging from as little as 0.1% to more than 5%. If potential impacts on ground water from these sites were to be considered, breakthrough times and concentrations delivered should be carefully evaluated.

Such calculations may explain the absence of any major changes in bulk density adjacent to the site in Fig. 5. Although the bulk density of clay on the east side (D_E) of the ring-drain combination was about 100 kg/m^3 less than on the west side (D_W), little change at either location occurred during the one-year study. Comparison with the thin horizontal lines suggests a very small increase in bulk density at ponding for both sites and a very small overall increase during the one-year study. Such small increases in density may be accounted for by water movement into the cracks and channels at ponding and into the clay matrix following ponding. In terms of environmental impact, the flow rate of concern would be the 10,000 x 10^{-9} m/sec breakthrough time for tracer

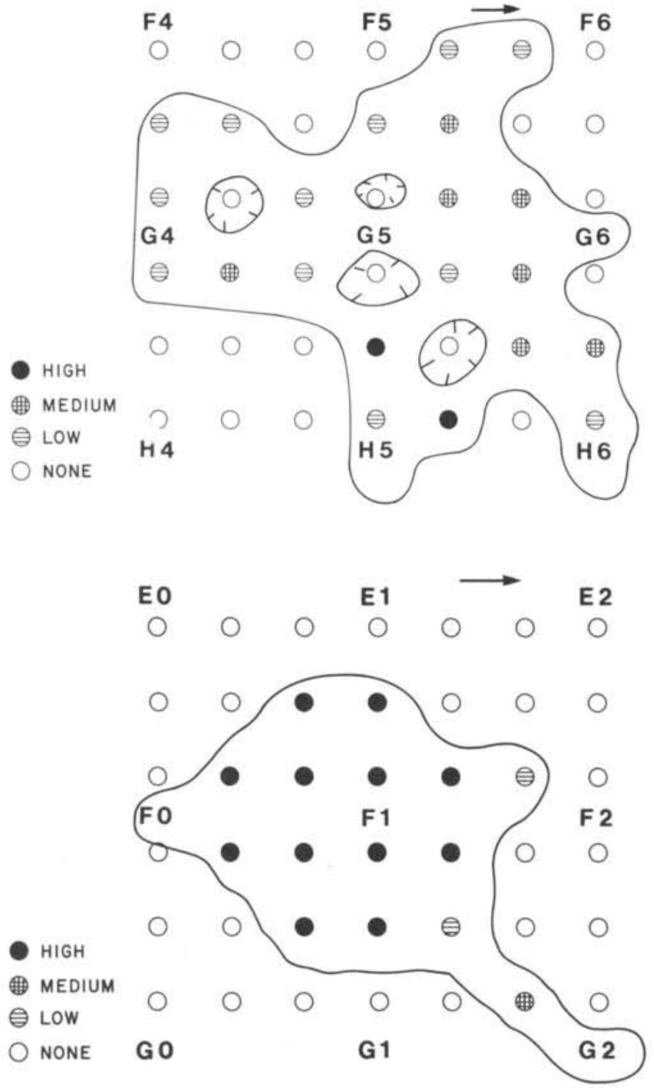

Figure 7. Relative distribution of Br^- tracer in compacted soil around fast leaching application sites G5 (a-top) and a slow leaching application site F1 (b-bottom) (from Rogowski, 1988)

rather than the very slow ($\sim 1 \times 10^{-9}$ m/sec) matrix flow component. However, to evaluate the potential impact on ground water, this high flow rate needs to be considered in terms of cross-sectional area contributing to flow ($\leq 1\%$) and the observed concentrations of tracer in the recharge.

Subsequent application of fluorescent dyes, coring of the tracer application area and surrounding sites on the 0.30 m grid with a Veihmeier tube, inspection of holes with a borescope, and qualitative tests for Br^- and dye corroborated preliminary observations. The Veihmeier sampling tube (Veihmeier, 1929) is 2.46 cm in diameter and is driven into the ground with a pointed weight that also doubles as an extractor. Relatively intact cores 0.30 m in length were removed for analysis. Based on such cores our results showed that on sites where much tracer was lost in leachate, the Br^- distribution "plume" was quite extensive but rather diffuse throughout the area of nine drains (Fig. 7a). But on other sites where little tracer was lost as leachate, strong evidence of a tight, well-defined plume was found in corings (Fig. 7b) surrounding the infiltration ring to which tracer had been added. Sites with more rapid breakthrough appeared to have a higher incidence of macropores, as evidenced by inspection of the Veihmeier holes with a borescope. These results suggest that highly mobile tracers or contaminants may be more difficult to detect close to the source of application, even though their impact could be greatest further away. Additional corroboration of these findings may be found in data on the quality of chemical leachate from the compacted-clay study (Rogowski 1986b). Leachate from sites with relatively low observed flow rates (and few potential macropores) consistently had a higher electrical conductivity and cation content than did the leachate from sites with relatively high flow rates (and many potential macropores), where residual salts appear to have been leached out prior to sampling.

Extrapolating these concepts to the field, the impact of a field site on ground water quality may be dictated by its potential to raise a contaminant concentration above some predetermined threshold level. The impact may also need to be evaluated from a standpoint of how quickly a given field site can deliver the contaminant to ground water. Consequently, sites exhibiting high probability of outflow rates that exceed a predetermined threshold value have a potential for greater impact.

ROLE OF LANDFORMS

Analytical concepts of unsaturated flow have been studied extensively in the laboratory, in the field (for homogeneous materials), and in small plots. Shifting emphasis to macropore flow (Wagenet and Germann 1989) as a pathway of solute transport in the vadose zone, coupled with recent work by Philip (1987) and coworkers on seepage into cavities and their effect on transport, still involves microscale considerations. The parameters of interest are usually water content and values of potential measured at a point and extensive use of tracers, fluorescent dyes, and stains to mark the pathways of flow.

Even though agriculture impacts ground water quality at a point, impact zones may be predetermined and dominated to a large extent by the geometry and spatial configuration of landforms, associated soils, and the nature of parent material. The scale of consideration is much larger but not so large as to make it a regional concern. It is against this background that we need to review possible effects of anisotropy in flow parameters and associated effects of slope and its changes.

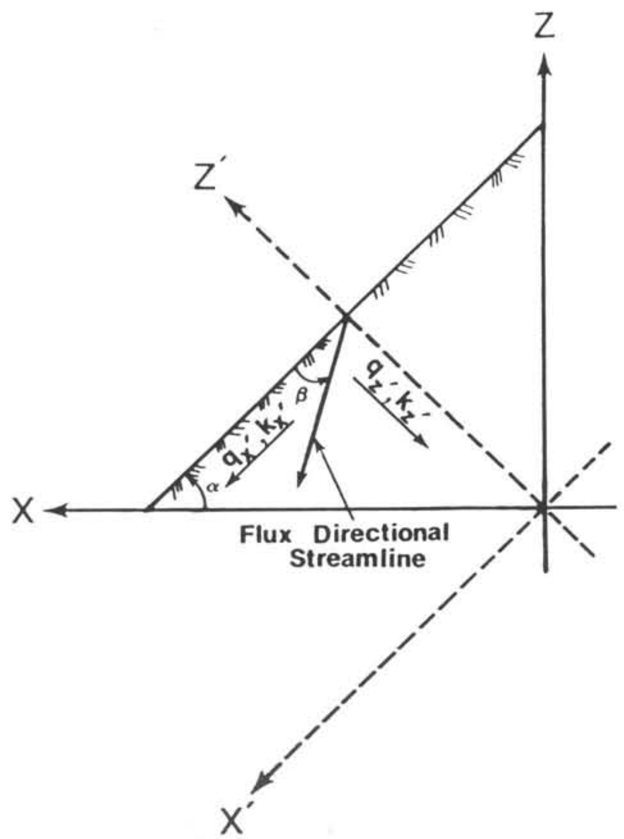

Figure 8. Schematic representation of flow components (q'_x) and (q'_z) and the hydraulic conductivities K_V, K_H normal and parallel to the soil layers; soil surface slopes at an angle α (from Zaslavsky and Rogowski, 1969).

In a two-layered system there is usually a tendency for infiltration flux to deviate from the vertical (Zaslavsky and Rogowski 1969). When the top layer is more permeable than the bottom one, the infiltrating water will slant downhill. The actual direction of flow is highly variable. For saturated conditions, the combined hydraulic conductivity value K'_V normal to the layers is always less than the conductivity parallel (K'_H) to the layers (Bear et al. 1968), and the profile as a whole behaves anisotropically. A similar phenomenon should be expected for unsaturated flow. Furthermore, in a sloping soil the combined force of gravity and potential gradient will usually deviate away from the normal in the downhill direction. Under these conditions (Fig. 9) we may expect, in addition to the flow component (q'_z) normal to the soil layers, a downhill lateral flow component (q'_x) parallel to the soil layers (Fig. 8). In general we may write,

$$q'_x / q'_z = \eta \tan \alpha \qquad (6)$$

where η is the degree of anisotropy given by the ratio of horizontal and vertical hydraulic conductivities (K'_H/K'_V) for a soil sloping at an angle α. The degree of anisotropy (η) will either remain constant throughout the entire range of saturation or change for different values of moisture content. The q'_x component cannot vanish unless the slope is zero. Thus, even with a slight rain, there exists a flow component parallel to the soil surface. The angle ß (Fig. 9) between the soil surface and the resultant flow direction can then be expressed as

$$\tan \beta = q'_z / q'_x = 1/\eta \tan \alpha \qquad (7)$$

Thus angle ß will decrease as tangent α and η increase, i.e., for larger slope and larger degree of anisotropy, the resultant flow direction will approach the direction parallel to the soil surface. Figure 9 illustrates these results for a hypothetical landform with anisotropy $\eta = 5$. The cross section is a transition from a concave to a convex shape, a common landform configuration in the Northeast. Figure 10 illustrates the convergency of hypothetical flows at various depths and distances along a concave slope. Where flows converge to a point, we could have localized zones of saturated flow. From Figs. 9 and 10 it is apparent that flowlines generally will converge below a concave slope. Moreover, as gradients cannot increase beyond a certain value, positive pressure towards the soil surface will develop, water content will build up, water will seep out, and rain will become overland flow. Thus, combinations of larger slopes and concave surfaces are more likely to produce seepage zones and runoff. The illustrations included here are in two dimensions to

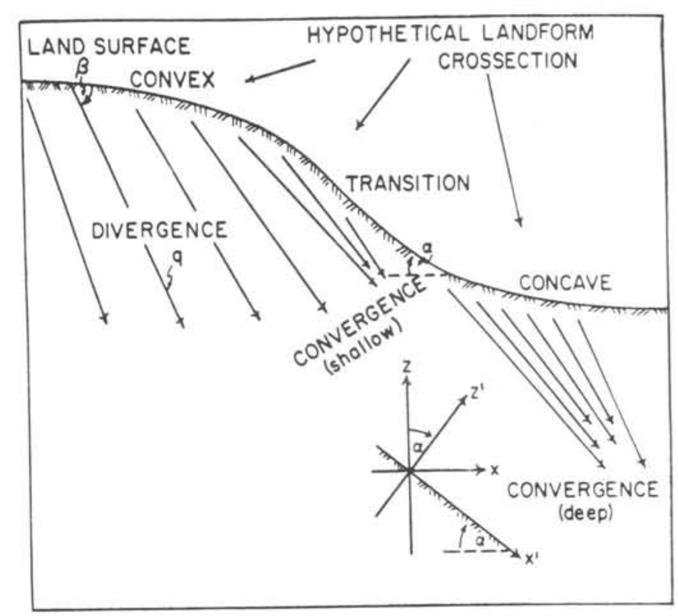

Figure 9. Hypothetical slope, and the direction of streamline (adapted from Zaslavsky and Rogowski, 1969)

emphasize the principles. Under natural conditions we should direct our attention to three-dimensional concave hollows. Seepage and runoff zones can occur when there are no highly impermeable layers, no perched ground water in the rest of the field, and no high intensity rain to produce incipient ponding.

This has numerous implications for solute transport. From the standpoint of mass balance, we cannot assume that pollutants leaving the root zone are transported directly to the underlying ground water. They may reemerge as seepage at the toe of a concave landform. Such three-dimensional landforms, where flowlines converge, are likely to act as zones of concentration for the solutes originating further upslope. Consequently, measurements taken at such locations should be interpreted with caution.

Although some concave landforms typically give rise to streams and seepage zones, others are but a part of the undulations of a mature landscape. Under these circumstances, zones of flowline convergence become an input to lower-lying areas, which by nature of position may be closer to the ground water. Thus discrete zones of flowline convergence that are distributed throughout a watershed may constitute primary locations where chemical transport is partitioned between surface water and ground water. Potential occurrence of such partitioning should be carefully considered in the northeastern United States, where hilly terrain, concave landform segments, and differences in the horizontal and vertical hydraulic conductivity are common.

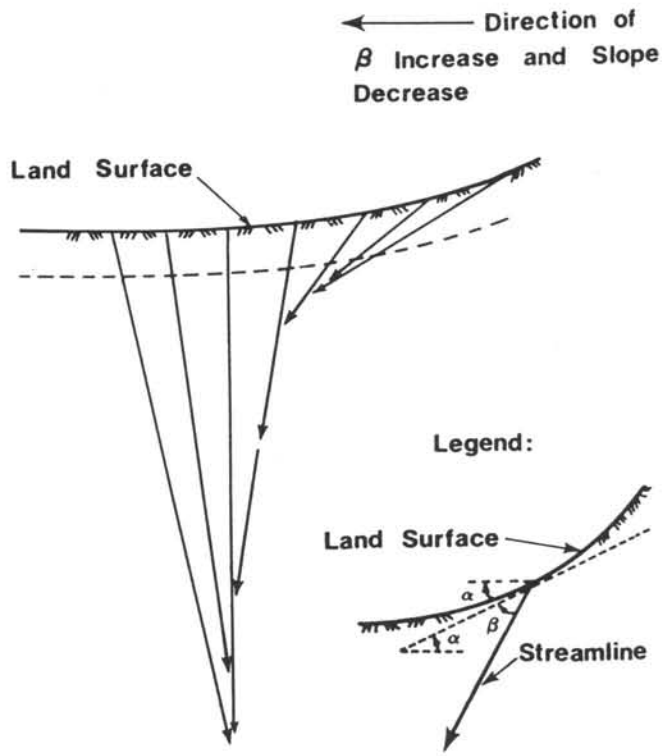

Figure 10. A schematic representation of streamline convergence. Dashed line is drawn parallel to the soil surface (adapted from Zaslavsky and Rogowski, 1969)

REPRESENTING HYDROLOGICAL VARIABILITY BY SCALING

Another approach to soil heterogeneity (Sharma and Rogowski 1985) is the use of scaling theory based on the concepts of similar media (Miller and Miller 1956), according to which soil hydrologic variability can be expressed by a single parameter (Warrick et al. 1977; Sharma et al. 1980). Laboratory studies (Reichardt et al. 1972; Youngs and Price 1981) suggest that this theory could be empirically extended for hydrologically dissimilar media, thus simplifying the computation of hydrological properties.

For many practical purposes, soil variability can be expressed in easily measured infiltration parameters such as sorptivity (S) and saturated hydraulic conductivity (K_s) (Sharma et al. 1980). Approximate conductivity [$K(\theta)$] and potential [$\psi(\theta)$] as functions of soil water content may be computed by using scaling theory (Peck et al. 1977; Sharma and Luxmoore 1979). It is better to characterize variability adequately over an area or in space by a procedure that allows rapid and reliable measurement of a property such as K, rather than to expend the same energy for a few measurements of the exact functions at a limited number of points.

The theory of scaling was developed from surface tension-flow considerations based on the concepts of similar porous media (Miller and Miller 1956). It aims to reduce the number of variables; thus, it provides a framework for simplifying the heterogeneity observed in the field. According to the theory, similar media are scale models of one another and differ only in the magnitude of the characteristic pore dimension, called the characteristic length λ. Soils satisfying similar-media criteria must have identical porosity and the same relative pore-size distribution; this relativity should not change with the degree of saturation. These requirements may be easier to satisfy on a field scale, when we usually are dealing with a single soil series, than on a watershed where many soil series make up the landscape.

A dimensionless scaling factor α can be defined as a ratio of characteristic lengths λ,

$$\alpha_i = \lambda_i / \lambda_r \tag{8}$$

where subscripts i and r refer to ith and reference soil respectively. Based on similar-media criteria, water potential ψ and hydraulic conductivity K are scaled as

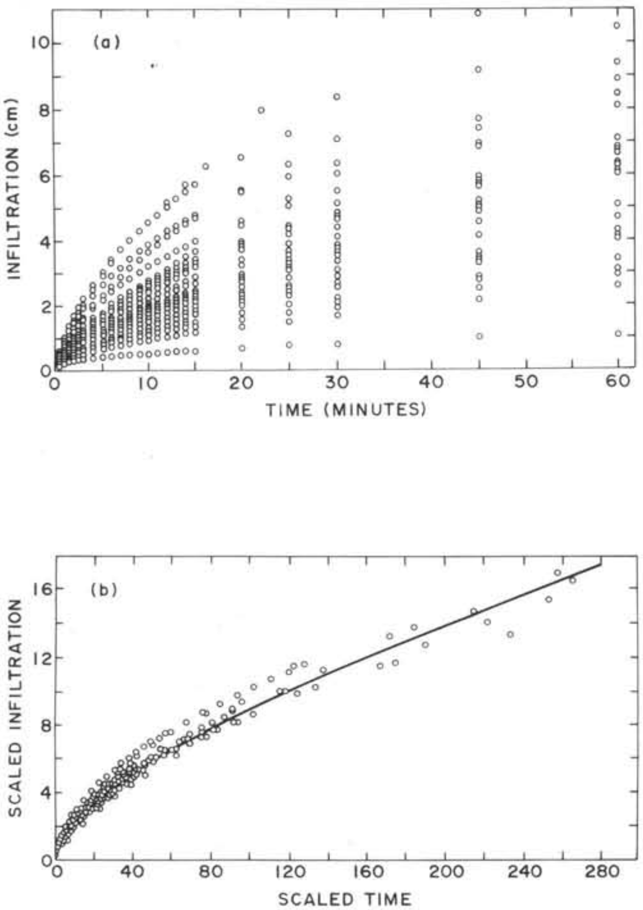

Figure 11. Plots of field measured cumulative infiltration (I) as a function of time (t) for a watershed (a) in the original form and (b) scaled form (from Sharma et al., 1980)

$$\psi_i = \psi_r \alpha_i \quad (9)$$

$$K_i = K_r \alpha_i^2 \quad (10)$$

It has been shown that with some approximations, such as the Philip (1969) infiltration equation, the theory can be used to scale field-measured hydrological properties (Warrick et al. 1977; Sharma et al. 1980), and soil hydrological variability can be expressed by a single parameter.

For scaling field-measured potential $\psi(\theta)$ and hydraulic conductivity $K(\theta)$ as functions of soil water content, the above expressions apply to conditions of identical water content. Since field soils do not have identical porosity and, therefore water content (θ), Warrick et al. (1977) used degree of saturation ($s = \theta / \theta_s$) in place of θ. As a result the variability of data points in the scaled form was considerably reduced, and potential (ψ) and conductivity (K) as functions of relative saturation coalesced into relatively narrow bands.

Sharma et al. (1980) demonstrated that the scaling factors can be obtained from infiltration parameters, which can be measured easily and quickly in the field using double-ring infiltrometers. At times, however, the use of ring infiltrometers is questionable, because the method suffers from a large wall-to-area ratio, often overestimates infiltration flux, and disturbs soil when rings are installed. Nevertheless, use of infiltration rings and scaling theory shows promise.

The utility of scaling theory is demonstrated in Fig. 11. In this case, the cumulative infiltration, I, as a function of time, t, at each of 26 locations in a watershed was approximated by Philip's (1957) two-parameter equation, $I = St^{1/2} + At$, where S is sorptivity and A is related to $K_s (A = 1/3\, K_s)$. Scaling factors based on S and A were calculated as follows:

$$\alpha S_i = (S_i / S_t)^2 \quad (11)$$

$$\alpha A_i = (A_i / A_r)^{\frac{1}{2}} \quad (12)$$

Studies to date suggest that scaling theory provides a framework to deal with voluminous data, thus quantifying the areal variability of hydrological properties in terms of a single, physically based parameter. The effect of varying this parameter on the hydrologic response of a watershed can then be examined. Such an approach has been used in studying the effect of variability on areal infiltration and drainage (Warrick and Amoozegar-Fard 1979), on water and solute transport (Bresler et al. 1979), and on water balance components of watersheds (Peck et al. 1977; Sharma and Luxmoore 1979; Luxmoore and Sharma 1980). It has recently been extended (Warrick et al. 1985) to include a generalized solution to infiltration.

Most studies indicate that it is necessary to account for spatial variability of hydrological properties and that the areal hydrologic response cannot be predicted from simple average properties. In these studies spatial dependence of the scaling factor (hydrological properties) has not been considered. Incorporation of spatial dependence and evaluation of its effect on the areal hydrologic response would be a fruitful research effort.

AGRICULTURAL WATERSHEDS

Bounded aggregations of landforms, where surface and ground water flow can be accounted for, make up a watershed. The boundary is usually an elevational boundary, and the ground water is the primary shallow aquifer that collects inputs from agricultural operations, supplies much of the potable water for livestock and humans, and gives rise to surface streams and ponds in predominantly rural areas. Consequently, the scale of

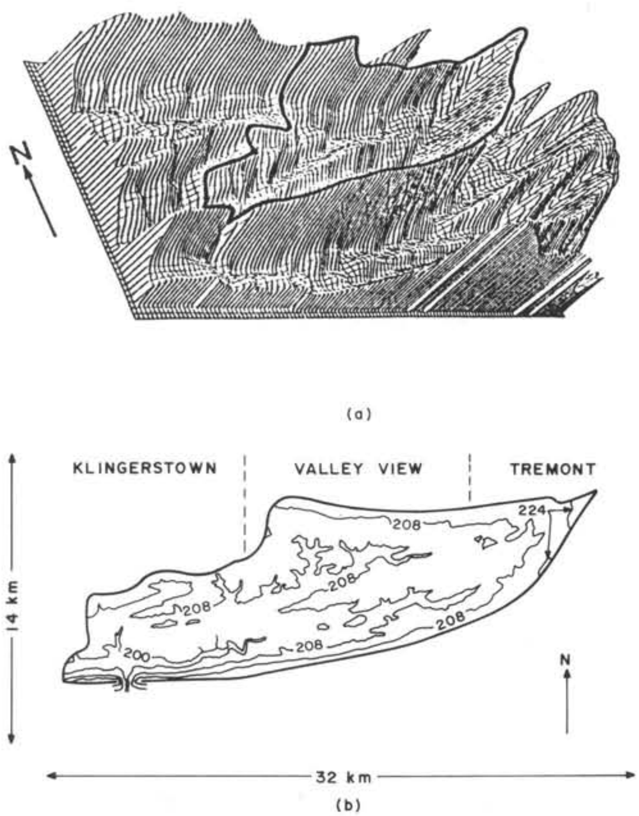

Figure 12. Digital elevation model (DEM) values for Klingerstown, Valley View and Tremont 7-1/2 minute topographic quadrangles comprising a 109 km^2 agricultural watershed shown: (a) perspective (b) gray scale.

concern needs to be sufficiently large to evaluate potential impacts from distributions of landforms, geologic, cover, land use, and soil effects. At this scale we are considering rather gross approximations and must attempt to evaluate possible trends in time and space. The problems that arise are the type of data that are available, the resolution that is possible, and the computer manipulations that must be used.

Type of Data

The type of data available generally consists of digitized 7-1/2 minute (1:24,000) topographic quad sheets referred to as United States Geological Survey Digital Elevation Model (DEM). Each elevation is for a pixel-sized (30 x 30 m) chunk of land, accurate to within 15 m. On the map, each pixel or cell represents a 1 mm^2 area. But elevations averaged on a 30 x 30 m grid could have considerable local deviations. The use of such data for analytical or numerical models may be limited. The data are generally coprocessed with other pertinent information, and digitized on a similar size grid, using geographical information system modules, available data, and models such as DRASTIC (EPA, 1985).

Modeling Potential Ground Water Pollution

DRASTIC is a model developed by the National Water Well Association under a grant from the United States EPA. It was designed to evaluate the pollution potential of any hydrogeologic setting within the country. The model consists of designated mappable units (usually on a 30 x 30 m grid or larger) and a rating system for readily obtainable factors digitized on the same grid. The factors are: depth to the ground water, net recharge, type of aquifer and soil media, site topography (slope), potential impact of the vadose zone, and hydraulic conductivity of the aquifer. A numerical ranking system is used to assess ground water pollution potential. The system consists of the weights, ranges, and ratings assigned by a panel of experts. Most significant factors have a weight of 5, the least significant a weight of 1. Factor values are divided into suitable ranges that reflect relative severity of potential impacts, and each range is assigned a rating from 1 to 10. The pollution potential for a location is then computed as the sum of weight and rating products of each factor.

In contrast to DRASTIC, the TFM of Jury (1982) attempts to describe average solute concentrations at any depth within a profile, based on measured distribution of solute travel times between the soil surface and some reference depth. While this has considerable merit on a field scale, application to large agricultural watersheds may be questionable, since in reality we seldom can tell how large an area is represented by one set of measured travel times.

Working with Large Data Sets

The usefulness of any predictive methodology depends on the extent to which the distribution of true values is matched both in space and time by their predicted or simulated counterparts. Our own problems at this scale involve primarily the extensive size of data sets and realistic computer costs. For example, attempts to construct variograms of elevations based on 7-1/2 minute quads are prohibitively expensive and time consuming, while selected elevation transects represent only ~500 data points or less than 0.5% of the population.

Computer manipulation of such data sets using geographical information systems (GIS) offers promise. Figure 12 represents elevations on a 109 km^2 agricultural watershed shown as a three dimensional perspective (a) and as contours based on gray scale (b). Figure 12a is based on subsampling of the original data set by taking

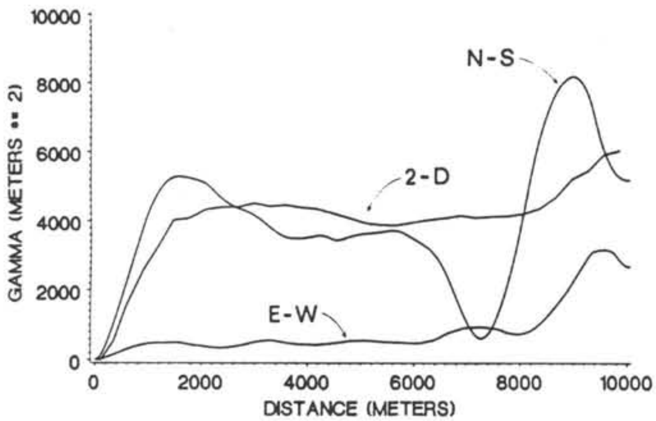

Figure 13. Semi variograms based on N-S (actually S-N), E-W elevation transects and 2-D subsampling (every 160 values) of Valley View Quad sheet.

every tenth (300 m) value. Figure 12b represents all values. Figure 13 shows N-S, E-W transects and a two-dimensional subsampling (every 160 points) of Valley View Quadrangle in Fig. 12. Of particular interest is the considerable variability evident in the N-S transect contrasted with the good structure of the E-W transect and the two-dimensional semivariograms. Figure 14 illustrates the mapping of respective slope (a) and slope aspects (b) on the Valley View Quadrangle and Table 1 gives a breakdown of the watershed by slope and aspect. Similar mappings and listings of soils, depth to ground water, cover, incidence of concave landforms, and other attributes are possible.

Once assembled, such data sets must be processed. It is not realistic to assume they can be processed using models such as TFM unless suitable averaging and sampling techniques are developed where each cell can truthfully represent $\sim 1\ km^2$. It appears that geostatistical programs can handle structure based on transect or discrete samplings. Thus total area and volume analysis may not be needed. A novel system of probability kriging and conditional simulations (Rogowski and Simmons 1988) may perhaps provide a methodology where ground truth and simulated reality, based on a diagnostic or even exploratory sampling campaign, can be combined with available soft data to provide realistic predictions of potential ground water pollution.

PROBABILITY KRIGING

To control ground water contamination, a critical action level needs to be established and maintained, subject to an objective and readily understood risk assessment procedure. The selected level should be

Table 1. Gray scale listings for slope and slope aspect DEM analysis for Valley Quadrangle.

Class	L-Limit	U-Limit	Pixels/ Class	Acres/Class (for 7.5 DEMS)	%
			Slope		
1	0.0000	0.1000	50752	11286.9766	32
2	0.1000	0.2000	64502	14344.9062	41
3	0.2000	0.3000	24749	5504.0469	16
4	0.3000	0.4000	11634	2587.3413	7
5	0.4000	0.5000	3568	793.5044	2
6	0.5000	0.6000	745	165.6842	<1
7	0.6000	0.7000	243	54.0419	<1
8	0.7000	0.8000	124	27.5770	<1
9	0.8000	0.9000	66	14.6781	<1
10	0.9000	1.0000	43	9.6530	<1
Unclassified Pixel Elements		94			
			Slope Aspect		
1	0 North	90.0000	27385	6090.2812	18
2	90.0000 East	180.0000	47694	10606.8945	30
3	180.0000 South	270.0000	32508	7229.6055	21
4	270.0000 West	361.0000	48933	10882.4414	31
Unclassified Pixel Elements		0			

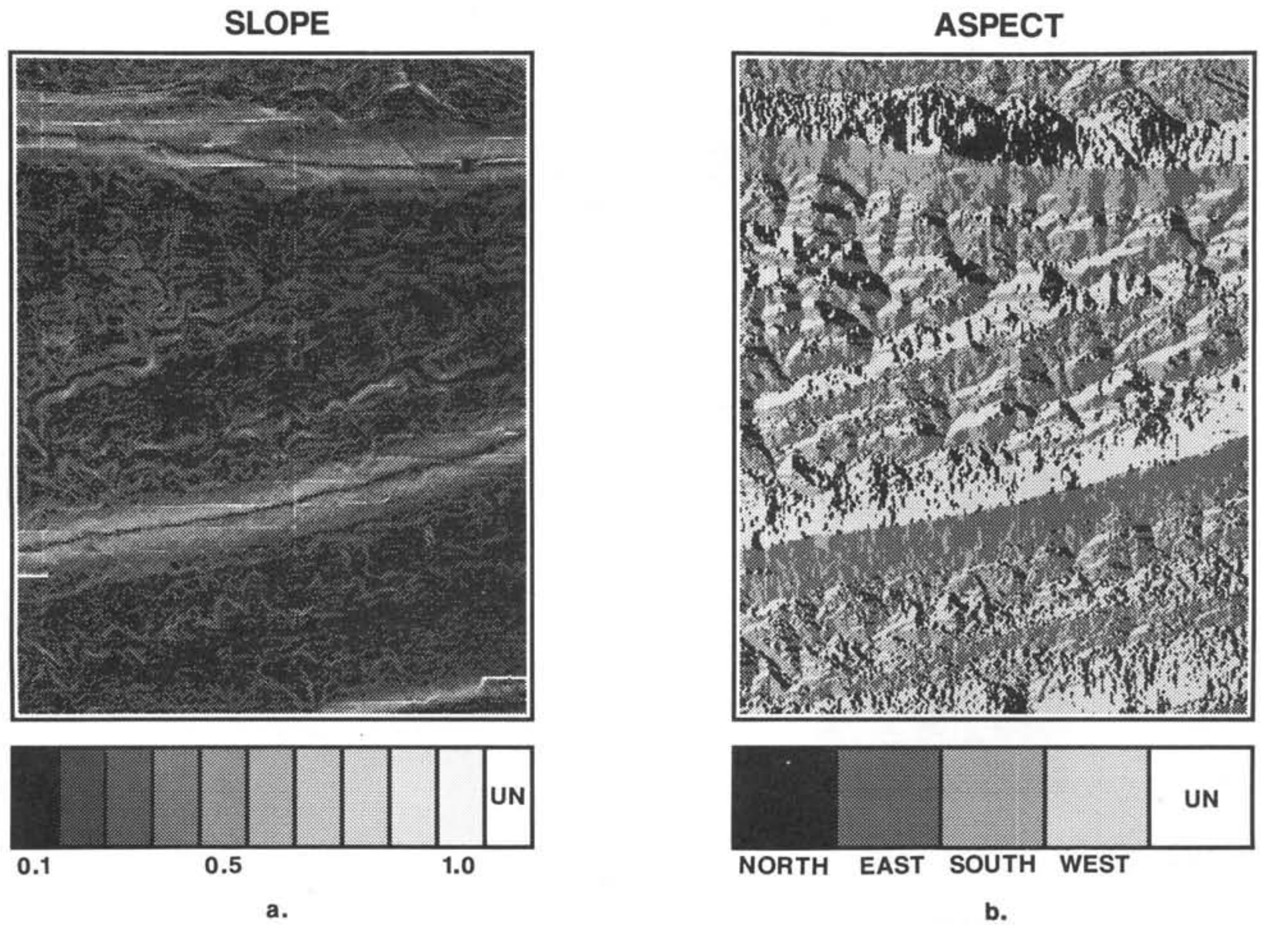

Figure 14. Distribution of slope (a) and slope aspect (b) based on DEM data values for Valley View Quad sheet; UN stands for unclassified elements.

justifiable in terms of potential impact on people and environment. There are multiple strategies available to tackle this problem. We will briefly describe one approach--probability kriging--illustrating it with examples from a study of water flow in compacted clay. Probability kriging was developed at Stanford University by Journel (1983, 1988) and his students (Isaaks 1984 and Sullivan 1984a,b) and has been applied to distribution of lead in soil surrounding a lead smelter in Dallas, Texas by Flatman et al. (1985). It is related to the nonparametric, risk-qualified indicator approach to data analysis (Journel 1983) and has recently been used by us to evaluate the conditional probability distribution of hydraulic conductivity in compacted clay (Rogowski and Simmons 1988).

To institute meaningful corrective action, we first need to know the impact area involved and a critical action level associated with a contaminant. While action levels for some agricultural chemicals are available (i.e., Canter 1986), impact areas are more difficult to define. The difficulty is that specific impacts originate on individual farm fields, while impacted areas such as seepage faces, streams, or portions of an aquifer are usually located "downstream" from the actual point of impact.

Exploratory sampling and analysis are not always a good indicator of potential problems. Although such sampling is generally conducted where experts believe pollutants should accumulate (Flatman 1984), fast-moving chemical species could have already been carried away from the sampled site, either through volatilization, runoff, or erosion, or by selective rapid percolation through macropores. A better exploratory approach is the use of models that utilize as input information derived from moderate amounts of field sampling, and detailed soil survey and topographic maps of the area. Unfortunately modeling often seems to be an end in itself. Seldom if ever are such exploratory analyses followed by a thorough diagnostic sampling of the potential impact areas (Flatman et al. 1985).

Spatial variables are often correlated within a certain distance of each other, i.e., samples taken close together are more similar than samples taken further apart. The spatial dependence of neighboring observations is usually expressed as a semivariogram [$\gamma(h)$], which describes the

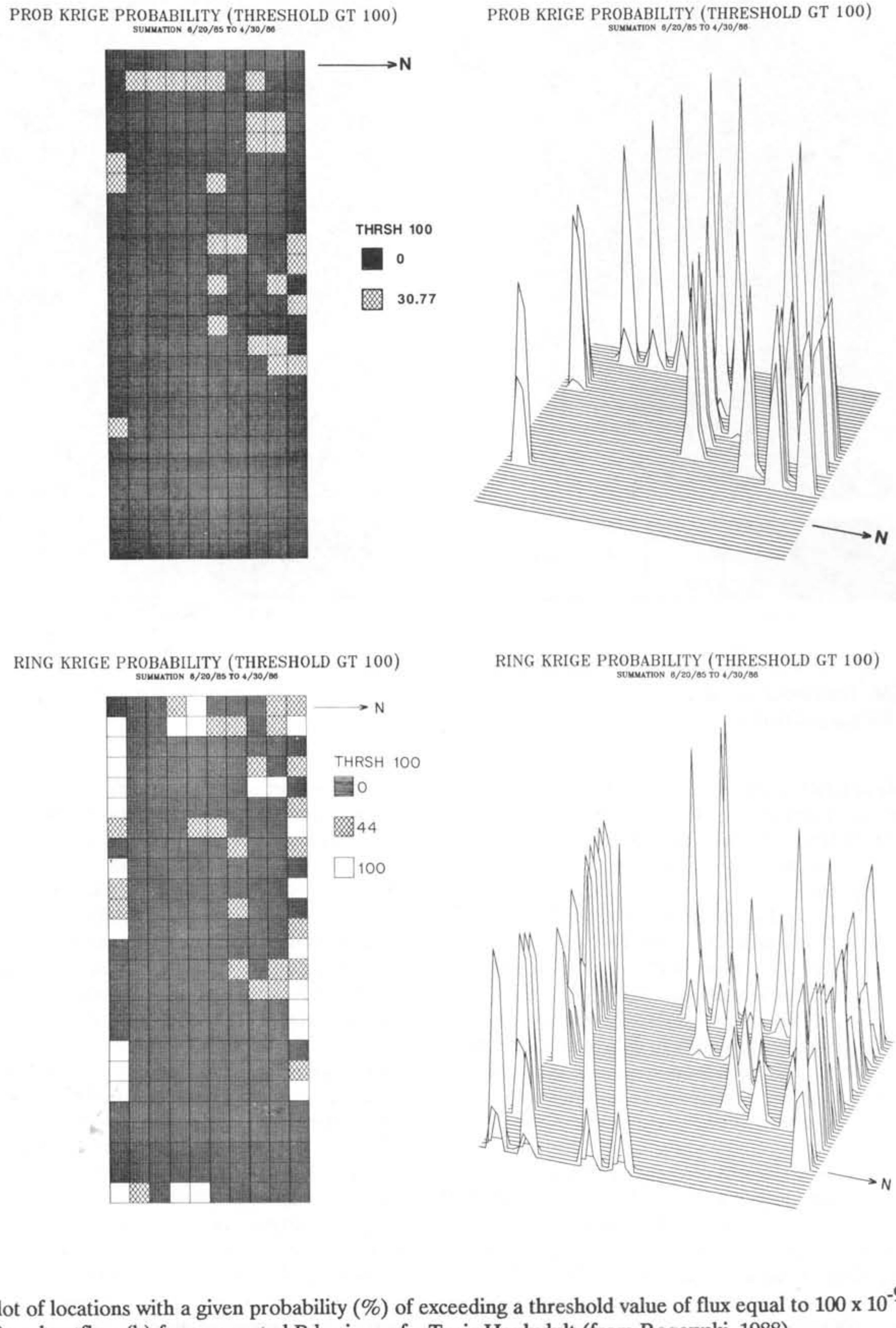

Figure 15. Plot of locations with a given probability (%) of exceeding a threshold value of flux equal to 100×10^{-9} m/sec for inflow (a) and outflow (b) for compacted B horizon of a Typic Hapludult (from Rogowski, 1988).

average rate of change of $\gamma(h)$ with distance and shows the variance structure of observations that may or may not follow any type of probability density function. Once the spatial structure is defined by a semivariogram, a square sampling grid equal to about 2/3 of the range is chosen and superimposed over the entire exhaustively sampled area. Sampling points can then be kriged to describe the distribution of a property over an area, in space, or in time. Using information from the semivariogram, ordinary kriging then produces linear estimates of a property at a point and an associated value of variance.

In contrast, probability kriging takes measured values sampled over an area and transforms them into an empirical frequency distribution, providing probability estimates based on a frequency of sample values in a neighborhood, rather than on their magnitudes. Means, medians, or any quantile of the frequency distribution can then be displayed as spatial contours for any area of interest.

To construct the model of uncertainty, we observe that the probability of an unknown value of flux, which is less than or equal to some critical level, corresponds to the fraction of measured values that are less than or equal to that level. Thus for each grid location and chosen critical level, an indicator variable is defined as zero (if the observed value is less than the selected level) and one (if the observed value is greater than or equal to the selected level). Since there is likely to be a dependence between measured values that are close together, the model of uncertainty can also be adjusted using appropriate weights and supplemented by additional data in the form of uniform rank order transformations (Journel 1988). Uncertainty about the estimated value of flux at a point, which can be thought of as potential recharge to the ground water, can be expressed using probability kriging isopleths of expected values of outflow flux. Uncertainty can also be expressed by quantiles of the distribution. For example, the 0.8 quantile map gives the flux distribution that has an 80% probability of being less (and 20% of being more) than a given contour value. If the expected value and the 0.8 quantile distribution are similar, there is, on the average, only a 20% chance of exceeding the expected value anywhere within the study area.

Perhaps the most versatile and easily understood tools of probability kriging are isopleths of maximum probability of exceeding a given threshold or tolerance value of flow. For example, Fig. 15 shows the location of sites that have a 44% or 100% probability of exceeding a 100×10^{-9} m/sec threshold for inflow (a) and 30% probability of exceeding the same threshold outflow (b). These "outlier" locations could be prime sites for preferential flow. Because it is not always possible to sample on as dense a grid as used in this study (Rogowski 1988), the questions of how representative the sampling is and of how to identify outliers naturally arise. Figure 16 shows four different outflow sampling schemes and their semivariograms with and without outliers. For every sampling scheme used, the presence of outliers led to highly variable semivariograms, while the removal of outliers resulted in similar variograms for all schemes. These results suggest that spatial distribution of outflow (and perhaps of inflow) may consist of a bimodal field. One part of the field is composed of continuous, highly correlated values of matrix flux. The other is essentially either a random distribution or a distribution that is continuous on a much larger scale of preferential-flow pathway clusters, superimposed on the clay matrix. The notion is similar to the concepts of flow in "structured soils" discussed by Nielsen et al. (1986).

Uncertainty about exceeding some preselected critical value can also be expressed as a risk of false positive (i.e., the probability of wrongly declaring the flux to be greater than a given level) or false negative (i.e., the probability of wrongly declaring the flux to be less than or equal to a given level). The selected critical level may in turn be related to some impact strategy chosen to minimize monetary loss or environmental impact. The risk may be defined statistically as the expectation of loss, or defined deterministically as the product of the probability of loss times the observed loss (Flatman et al. 1985). The false positive or false negative risk may thus have to be evaluated against the background of acceptable probability of loss for each particular situation.

SPATIAL VARIABILITY

Although analytical solutions can be applied only to highly simplified systems, introduction of increasingly powerful and sophisticated numerical models (Nielsen et al., 1986) requires realistic evaluation of field parameters. To be meaningful on a watershed scale, such parameters must involve some type of an averaging or approximation over an area or volume of soil and a decision regarding time interval to be considered. Averaging over time is often advisable (Fig. 5 "430 data"), particularly for cases when small volumes (or concentration) are involved on a daily or weekly basis (Rogowski 1988). Gelhar (1986) suggests that heterogeneity of aquifers favors a probabilistic rather than a deterministic approach for assessment of variability in water flow and solute transport. Much the same can be said about soils.

What complicates the matter further is confusion about the nature of the variability of hydraulic parameters. Particular difficulties exist with regard to hydraulic conductivity and total potential. Neither can be considered a field property (Cushman 1986) that can be measured directly, but each is a constituent property that is derived from field measurements such as flow, tension,

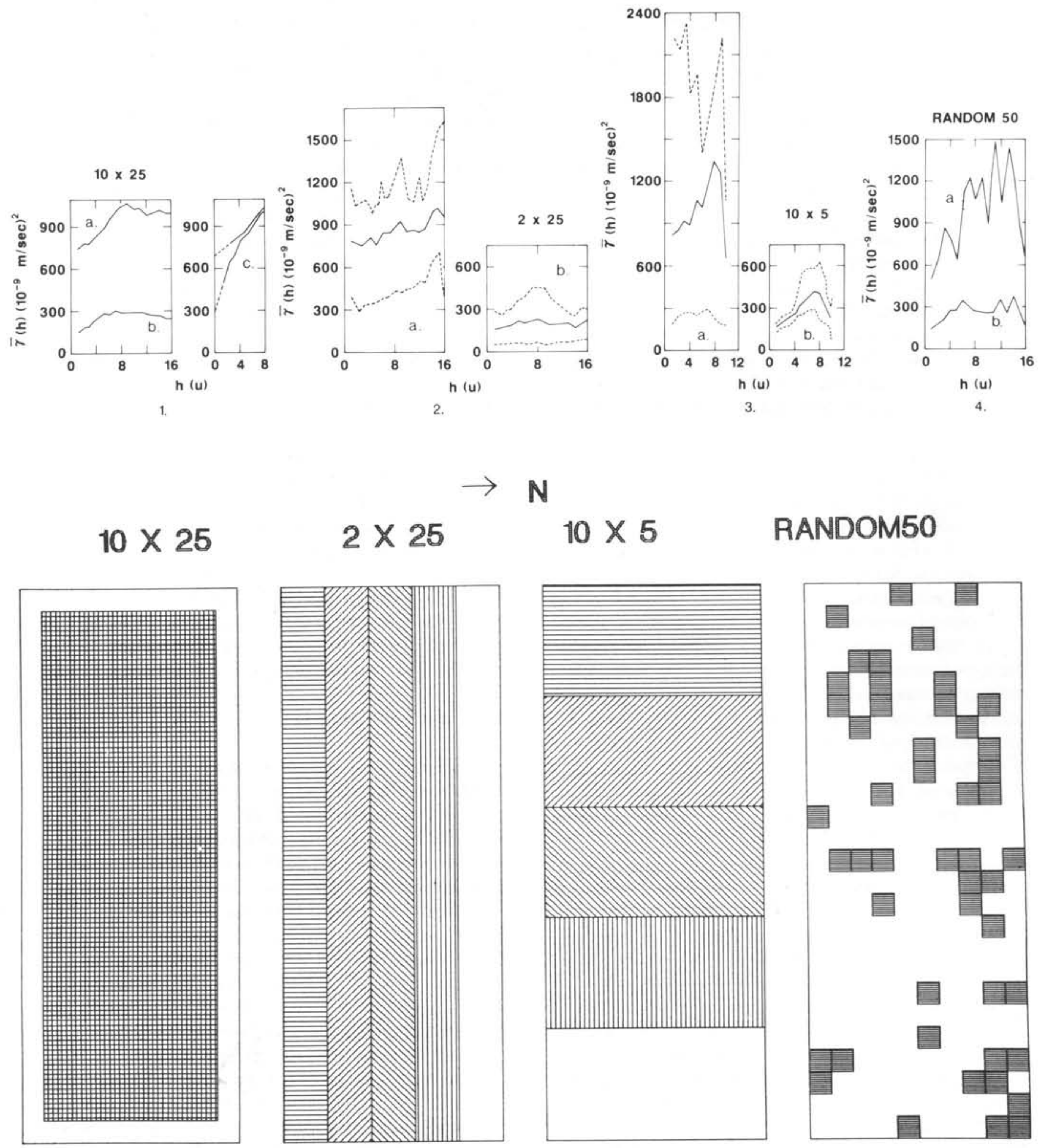

Figure 16. Four different sampling schemes (inset) evaluated using raw semivariograms of outflow flux density with (a) and without (b) outlier values shown in Fig. 15b; insert (c) in #1 denotes differences in apparent nugget effect with (top) and without (bottom curve) high flow parameter values; dashed lines indicate extent of variability, solid lines are the mean values (from Rogowski and Simmons, 1988).

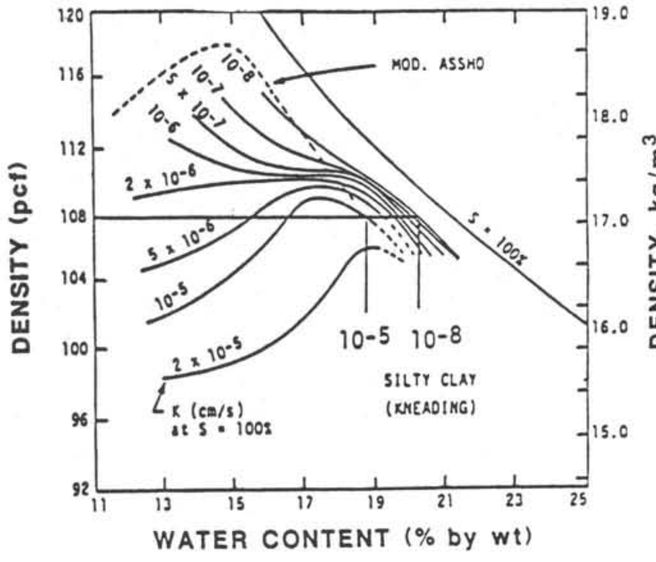

Figure 17. Laboratory measured permeability as a function of density and moulding water content; dashed line indicates modified ASSHO compaction curve (adapted from Mitchell et al., 1965, their Fig. 13).

changes in water content, and bulk density. One difficulty is that, while tension is usually measured at a point, solute and water fluxes, moisture contents, and bulk densities often represent averages over indeterminate volumes. For example, measurements with surface moisture/density probes (Fig. 3) can involve different volumes of soil, depending not only on the water content but also on the distribution of water with relation to the probe. In subsequent computations of hydraulic conductivity, measurements of water content and potential, which may be based on different volumes of soil, are tacitly assumed to apply to the same volume.

Darcy's Law and its extension to unsaturated systems are based on a form of space- and time-averaging over a volume, although the amounts of volume and time are not made clear. Mitchell et al. (1965) studied small soil cores compacted at different values of moulding water content and different levels of dry density, illustrating the combined effects of water content and density on hydraulic conductivity at a point (Fig. 17). Within a very narrow range of water content and density, hydraulic conductivity varied by three orders of magnitude.

Cushman (1986) suggests scaling measuring devices in order to describe true distribution of measured field properties. He maintains that laboratory apparatus needs to be modified and scaled up to evaluate field values. The problem reverts to the prior discussion on the geostatistical concepts of scale and support as well as to Bear's (1972) concept of REV.

SUMMARY AND CONCLUSIONS

What then do we need to know? As we attempt to evaluate the potential for ground water pollution in the Northeast, it appears that some very basic questions still require answers.

Different processes such as water flow, solute transport, adsorption, desorption, diffusion and dispersion all operate at different scales of time and space. We need to know over what volume of soil and what interval of time hydraulic properties such as conductivity and potential and chemical attributes of leachability and transportability should be averaged to describe water flow and solute transport at a field, landform, or watershed scale. Moreover, we need to know what types of measuring instruments and methodology are optimal.

Similar questions pertain to the transfer function model of Jury (1982) and its field applications (i.e., Jury and Stolzy 1982) and the conditional probability distribution extrapolations of Rogowski and Simmons (1988) and Rogowski et al. (1988). Suction lysimeters can sample only point volumes for tracer breakthrough even though the tracer is applied as a pulse over the whole field.

Since macropores, cavities, rocks, fertilizer granules, and organic debris may concentrate seepage flows in discrete portions of an otherwise unsaturated profile, how much do we need to know about their distributions in the vadose zone? The answers are needed for meaningful interpretation of flow and transport as well as for identification of possible sites of chemical and biological transformations.

The role of landforms, particularly the spatial distribution of concave hollows on watersheds, needs to be evaluated from the standpoint of concentrating seepage flows on scales larger than farm fields.

Because extensive heterogeneity and uncertainty is characteristic of natural systems, the use of both well-tried and novel statistical methods needs to be strongly encouraged. Applications involving hierarchical analysis of variance and geostatistical tools of structural analysis, kriging, probability kriging, and risk analysis should be emphasized, along with methods exploring applicability of fractals, time series, and spectral processes. Geostatistical methods are now widely employed in various fields of soil science (Webster 1985; Gutjahr 1985; and Nielsen and Bouma 1985). The problems that pose difficulties center primarily around the concept of stationarity (homogeneity) and the need for "adequate" numbers of samples (Philip 1987; Journel 1982). Nevertheless, approaches such as probability kriging (Journel 1988) may offer a positive alternative to models such as DRASTIC (EPA 1985).

Variogram analysis, employed to ascertain the structure of distributions on a field or watershed scale, holds much promise. Additional effort is needed to apply predictive and extrapolative methods of geostatistics to extremely large data sets, such as those of DEM and associated digitized soils, cover, and geologic information. GIS and geostatistical approaches must also be integrated on a large watershed scale.

Sposito (1986) made a strong recommendation to identify groups of similarity transformations of the Richards' equation to classify behavior of water in soils. Still, I am somewhat skeptical about our ability to represent spatially variable hydraulic properties on a watershed in a dimensionless, reduced form. However, it is an area that should be investigated further, particularly on a large scale and incorporating notions of spatial variability.

Conceptually, the processes associated with the production of food and fiber at a field scale on individual farms are potentially the source of ground water pollution on a watershed scale. Consequently, an integrated research approach is needed that simultaneously examines flow and transport at several different scales in order to evaluate distribution, linkages, impacts and risks of agricultural pollution. Up to now the criteria for chemical and organic constituents of soils as well as for fertilizer and pesticide needs have been based on availability of the element to plants and increased efficiency of production. Field soil properties, soil testing methods, and survey and classification procedures need to be thoroughly reexamined. They should possibly be changed and redirected towards the assessment of leachability and transportability of soil components and agricultural chemicals.

In conclusion I offer a brief quotation from Burges (1986):

> "Progress is made by those who take as broad a view as possible and recognize the totality of a problem, rather than a fragment."

LITERATURE CITED

Baker, D.E. and M.C. Amacher. 1981. Development and interpretation of a diagnostic soil testing program. Pa. Agric. Exp. Stn. Bul. No. 826, University Park, PA.

Bear, J. 1972. Dynamics of fluids in porous media. America Elsevier, NY.

Bear, J., D. Zaslavsky and S. Irmay. 1968. Physical principles of water percolation and seepage. UNESCO, Paris, France. 425 pp.

Brasino, J.S. 1986. A simple model predicting conservative mass-transport through the unsaturated zone into ground water. University Microfilm International, Ann Arbor, MI 48106.

Bresler, E., H. Bielorai and A. Laufer. 1979. Field test of solution flow models in a heterogeneous irrigated cropped soil. Water Resour. Res. 15:645-652.

Burges, S.J. 1986. Trends and directions in hydrology. Water Resour. Res. 122(9):1S-5S.

Canter, L.W. 1986. Environmental impacts of agricultural production activities. Lewis Publishers, Inc., Chelsea, MI. 382 pp.

Cushman, J.H. 1986. On measurement, scale, and scaling. Water Resour. Res. 22(2):129-134.

Dagan, G. 1986. Statistical theory of ground water flow and transport: Pore to laboratory, laboratory to formation, and formation to regional scale. Water Resour. Res. 22(9):120S-134S.

Darcy, H. 1856. Les Fontaines Publiques de La Ville de Dijon. Paris, Dalmont. 647 pp.

Evans, B.M. and W.L. Myers. 1986. Computer analysis of multiple layers of spatially oriented environmental data for evaluating potential groundwater impacts. Institute for Research on Land and Water, The Pennsylvania State University, University Park, PA. 67 pp.

Flatman, G.T. 1984. Using geostatistics in assessing lead contamination near smelters. *In* Environmental Sampling for Hazardous Waste Sites, Schweitzer Publ. ACS, Washington, DC, pp. 43-52.

Flatman, G.T., K.W. Brown and J.W. Mullins. 1985. Probabilistic spatial contouring of the plume around a lead smelter. *In* The 6th National Conference on Management of Uncontrolled Hazardous Waste Sites, Nov. 4-6, Washington, DC, pp. 442-448.

Gelhar, L.W. 1986. Subsurface hydrology from theory to application. Water Resour. Res. 22(9):135S-145S.

Gutjahr, A. 1985. Spatial variability: geostatistical methods. *In* D.R. Nielsen and J. Bouma (eds.) Soil Spatial Variability, Proc. ISSS and SSSA Workshop, pp. 9-34.

Isaaks, E.H. 1984. Risk qualified mappings for hazardous wastes, a case study in nonparametric geostatistics. MSc Thesis, Branner Earth Sciences Library, Stanford, CA. 85 pp.

Journel, A.G. 1983. Nonparametric estimation of spatial distributions. Mathematical Geology 15(3):445-468.

Journel, A.G. 1986. Geostatistics: Models and tools for the earth sciences. Mathematical Geology 18:119-140.

Journel, A.G. 1988. Nonparametric geostatistics for risk and additional sampling assessment. *In* Principles of Environmental Sampling, American Chemical Society, L.H. Keith (ed.) pp. 45-72.

Jury, W.A. 1982. Simulation of solute transport using a transfer function model. Water Resour. Res. 18:363-368.

Jury, W.A. and L.H. Stolzy. 1982. A field test of the transfer function model for predicting solute transport. Water Resour. Res. 18:369-375.

Luxmoore, R.J. and M.L. Sharma. 1980. Runoff responses to soil heterogeneity: Experimental and simulation comparisons for two contrasting watersheds. Water Resour. Res. 16:675-684.

Miller, E.E. and E.D. Miller. 1956. Physical theory for capillary flow phenomena. J. Appl. Phys. 27:324-332.

Mitchell, J.K., D.R. Hager, and R.G. Campanella. 1965. Permeability of compacted clay. J. Soil Mech. and Foundations Div. ASCE 91:41-65.

Nielsen, D.R., J.W. Biggar and K.T. Ehr. 1973. Spatial variability of field measured soil-water properties. Hilgardia 42:215-259.

Nielsen, D.R. and J. Bouma (eds.). 1985. Soil spatial variability. Proc. of a Workshop of the ISSS and The SSSA, Las Vegas, USA, 30 Nov.-1 Dec. 1984. Pudoc, Wageningen. 243 pp.

Nielsen, D.R., M. Th. van Genuchten and J.W. Biggar. 1986. Water flow and solute transport processes in the unsaturated zone. Water Resour. Res. 22:89S-108S.

Parkin, T.B. 1987. Soil microsites as a source of denitrification variability. Soil Sci. Soc. Am. J. 51(5):1194-1199.

Peck, A.J., R.J. Luxmoore and J.L. Stolzy. 1977. Effects of spatial variability of soil hydraulic properties in water budget modeling. Water Resour. Res. 13:348-354.

Philip, J.R. 1957. The theory of infiltration. 4. Sorptivity and algebraic infiltration equations. Soil Sci. 84:257-264.

Philip, J.R. 1986. Issues in flow and transport in heterogeneous porous media. Transport in Porous Media 1:319-338.

Philip, J.R. 1987. The quasilinear analysis, the scattering analog, and other aspects of infiltration and seepage. *In* Proceedings, International Conference on Infiltration Development and Application, Jan. 6-9, Yu-Si Fok (ed.) Univ. of Hawaii, Honolulu, Hawaii, pp. 1-27.

Reichardt, K., D.R. Nielsen and J.W. Biggar. 1972. Scaling of horizontal infiltration into homogeneous soils. Soil Sci. Soc. Am. Proc. 36:241-245.

Richards, L.A. 1931. Capillary conduction of liquids in porous mediums. Physics 1:18-333.

Rogowski, A.S. 1972. Watershed physics: Soil variability criteria. Water Resour. Res. 8:1015-1023.

Rogowski, A.S., R.M. Khanbilvardi and R.J. DeAngelis. 1985. Estimating erosion on plot, field, and watershed scales. *In* S.A. El-Swaify, W.C. Moldenhauer, and Andrew Lo (eds.) Soil Erosion and Conservation. Soil Conservation Society of America, Ankeny, Iowa, pp. 149-166.

Rogowski, A.S., B.E. Weinrich and D.E. Simmons. 1985. Permeability assessment in a compacted clay liner. *In* Eighth Annual Madison Waste Conference, Municipal and Industrial Waste. University of Wisconsin, Madison, WI, pp. 315-335.

Rogowski, A.S. 1986a. Hydraulic conductivity of compacted clay soils. *In* Proceedings 12th Annual Research Symposium, U.S. Environmental Protection Agency, EPA-600/9-86-022, Cincinnati, OH, pp. 29-39.

Rogowski, A.S. 1986b. Degree of saturation, hydraulic conductivity, and leachate quality in a compacted clay liner. *In* R.M. Khanbilvardi and J. Fillos (eds.) Ground Water Hydrology, Contamination, and Remediation, Scientific Publications Co., Washington, DC, pp. 339-353.

Rogowski, A.S. 1987. Distribution of flow rates and tracer breakthrough times in field soil. *In* Second Int. Conf. on New Frontiers in Hazardous Waste Management, Proceedings. EPA/600/9-87/018T, pp. 219-230.

Rogowski, A.S. and D.E. Simmons. 1988. Geostatistical estimates of field hydraulic conductivity. Mathematical Geology. 20(4):423-446.

Rogowski, A.S. 1988. Flux density and breakthrough times for water and tracer in a compacted clay soil. J. Contaminant Hydrol. 3:327-348.

Rogowski, A.S., J.K. Wolf and D.E. Simmons. 1988. Conditional simulation of flow and transport through a spatially variable compacted clay soil. *In* Proceedings Validation of Flow and Transport Models for the Unsaturated Zone, Peter J. Wierenga (ed.) May 23-26, Ruidoso, NM.

Sharma, M.L. and R.J. Luxmoore. 1979. Soil spatial variability and the consequences on simulated water balance. Water Resour. Res. 15:1567-2573.

Sharma, M.L., G.A. Gander and C.G. Hunt. 1980. Spatial variability of infiltration in a watershed. J. Hydrol. 45:101-122.

Sharma, M.L. and A.S. Rogowski. 1985. Hydrological characterization of watersheds. Proceedings, Natural Resources Modeling Symposium, Donn G. DeCoursey (ed.) Oct. 16-21, 1983, Pingree Park, Colorado, USDA-ARS, ARS 30, pp. 291-295.

Sposito, G. 1986. The "Physics" of soil water physics. Water Resour. Res. 27:(9)835-885.

Sullivan, J.A. 1984a. Nonparametric estimation of spatial distributions. Ph.D. Thesis, University Microfilms, Ann Arbor, MI. 368 pp.

Sullivan, J.A. 1984b. Conditional recovery estimation through probability kriging - theory and practice. *In* G. Verly et al. (eds.) Geostatistics for Natural Resources Characterization, Part 1, D. Reidel Publ. Co., Boston, MA, pp. 365-384.

U.S. Environmental Protection Agency. 1985. DRASTIC: A standardized system for evaluating groundwater pollution potential using hydrogeologic settings. Robert S. Kerr Environmental Research Laboratory, EPA/600/2-85/018. 163 pp.

Veihmeier, F.J. 1929. An improved soil sampling tube. Soil Sci. 27:147-152.

Vieira, S.R., D.R. Nielsen and J.W. Biggar. 1982. Spatial variability of field-measured infiltration rate. Soil Sci. Soc. Am. J. 45:1040-1048.

Wagenet, R.J. and R.F Germann 1989. Concepts and models of water flow in macropore soils. *In* Frink, C.R. Ground Water in the Northeast. The Connecticut Agricultural Experiment Station. Bulletin 876. 75pp.

Warrick, A.W., G.J. Mullen and D.R. Nielsen. 1977. Scaling field-measured soil hydraulic properties using similar media concept. Water Resour. Res. 13:355-362.

Warrick, A.W. and Amoozegan-Fard. 1979. Infiltration and drainage calculation using spatially scaled hydraulic properties. Water Resour. Res. 15:1116-1120.

Warrick, A.W., D.O. Lomen and S.R. Yates. 1985. A generalized solution to infiltration. Soil Sci. Soc. Am. J. 49:34-38.

Webster, R. 1985. Quantitative spatial analysis of soil in the field. *In* B.A. Stewart (ed.) Advances in Soil Science, 3:1-70.

White, R.E. 1985. The influence of macropores on the transport of dissolved and suspended matter through soil. In B.A. Stewart (ed.) Advances in Soil Science, 3:95-120.

Youngs, E.C. and R.I. Price. 1981. Scaling of filtration behavior in dissimilar porous materials. Water Resour. Res. 17:1065-1070.

Zaslavsky, D. and A.S. Rogowski. 1969. Hydrologic and morphologic implications of anisotropy and infiltration in soil profile development. Soil Sci. Soc. Am. Proc. 33:(4)594-599.

LIST OF SYMBOLS

$\gamma(h)$ - semivariogram variance measure, $(\ )^2$
h - lag distance, (m)
N - number of pairs
$Z(x_i)$ - value of a variable at location x_i
C - sill, $(\ )^2$
C_o - nugget, $(\ )^2$
q - flux density, m/s
K - hydraulic conductivity, m/s
ψ - potential, kPa
h - head, m
g - acceleration of gravity, m/s^2
z - distance, vertical, m
ϕ - source or sink, 1/s
θ - water content, m^3/m^3
j - gradient, m/m
Q - outflow volume, m^3/s
L - block length, m
t - time, s
K'_H - hydraulic conductivity parallel to soil layers, m/s
K'_V - hydraulic conductivity normal to soil layers, m/s
q'_x - flow component parallel to soil layers, m/s
q'_z - flow component normal to soil layers, m/s
α - soil slope, degrees
ß - angle between soil surface and resultant flow, degrees
η - K'_H/K'_V degree of anisotropy
α_i - scaling factor
λ_i - characteristic length
K_S - saturated hydraulic conductivity, m/s
S - sorptivity, $m/s^{1/2}$
A - (1/3)KS, m/s

Ground Water Flow

in the Northeast

William J. Gburek
Hydrologist
Northeast Watershed Research Center
Agricultural Research Service, USDA
University Park, PA 16802
and
Adjunct Associate Professor of Civil Engineering
The Pennsylvania State University
University Park, PA 16802

Contribution from the U.S. Department of Agriculture, Agricultural Research Service, in cooperation with the Pennsylvania Agricultural Experiment Station, The Pennsylvania State University, University Park, Pennsylvania.

CONTENTS

INTRODUCTION

Ground water is an important water source in nearly all inhabited places on earth. It is the dominant source of water for rural populations, it is the largest source of water for irrigation in arid or semiarid regions, and it is an important source for urban and industrial uses in humid climates. Traditionally, ground water has also been an economical resource. It is commonly available at the point of need at relatively low cost, not requiring construction of reservoirs and pipelines. Finally, it is usually thought to be of good quality, free of suspended sediment and bacterial contamination, and without the quality problems often associated with surface sources.

BASIC TERMINOLOGY AND GROUND WATER FLOW SYSTEMS

Before examining ground water in the northeastern setting, we must first establish some basic terminology and examine generalized subsurface flow. Ground water refers to subsurface water in fully saturated soils and geologic formations. It exists within the openings of the soil and rock, with the volume and transmissive characteristics of these openings dependent on the mineral composition and structure of the rock. Thus, to understand ground water, we must also be familiar with the rocks in which it exists.

Rocks exposed at the earth's surface are divided into three types. Igneous rocks are formed from molten or partially molten material, sedimentary rocks are formed by deposition of sediment by water, ice, or air, and metamorphic rocks are igneous or sedimentary rocks subjected to heat and pressure great enough to alter their structural characteristics and chemical composition. The nature of water-bearing openings in all these rock types depends to a great extent on the geologic age of the rocks as well as on the processes that formed them.

The youngest rocks are unconsolidated sedimentary deposits, and their transmissive openings are simply the pores between mineral grains. These rocks tend to have pore space greater than that of older rocks of the same type which have been subjected to consolidation and partial or complete filling of the openings by deposition of minerals and sediment. Intrusive igneous (rocks which have solidified at great depths below the land surface), and intrusive metamorphic rocks are among the oldest water-bearing rocks. When they were formed they contained no appreciable openings due to the great pressure of overlying rock strata. These rocks were originally subjected to pressures acting nearly parallel to the land surface. As the rocks were gradually exposed to erosion of the overlying rocks, the compressive forces acting on them were relieved, and they broke along sets of vertical and horizontal fractures. These fractures then serve as water-bearing openings. Similar fractures form in sedimentary rocks that had been buried deeply, consolidated, and then exposed by erosion of the overlying rock.

The surface of the earth is also underlain in many places by unconsolidated material: relatively young sedimentary deposits of rock fragments, alluvium, glacial drift, or material derived from mechanical or chemical weathering of the underlying bedrock. This material collectively is termed regolith, and can be an important consideration in analyses of ground water flow.

Figure 1 is a schematic of the subsurface in one dimension. The subsurface can be divided into the unsaturated zone where pore space contains both water and air, and the saturated zone which contains water alone. The capillary fringe, saturated but at a pressure less than atmospheric, is considered part of the saturated zone. The unsaturated or vadose zone, discussed in the other papers in this series, contains the root zone and that zone below extending to the top of the capillary fringe. The latter is sometimes termed the intermediate zone, and is generally looked on as the no-man's-land of hydrology. The water table, near the upper part of the saturated zone, is the level at which water is at atmospheric pressure; it is the level at which water in a

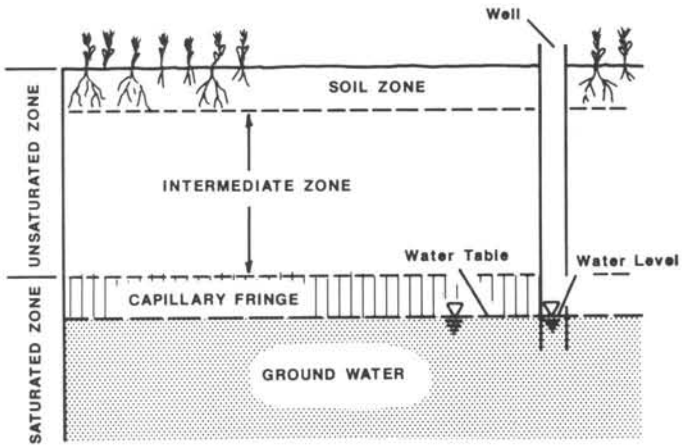

Figure 1. Water in the subsurface.

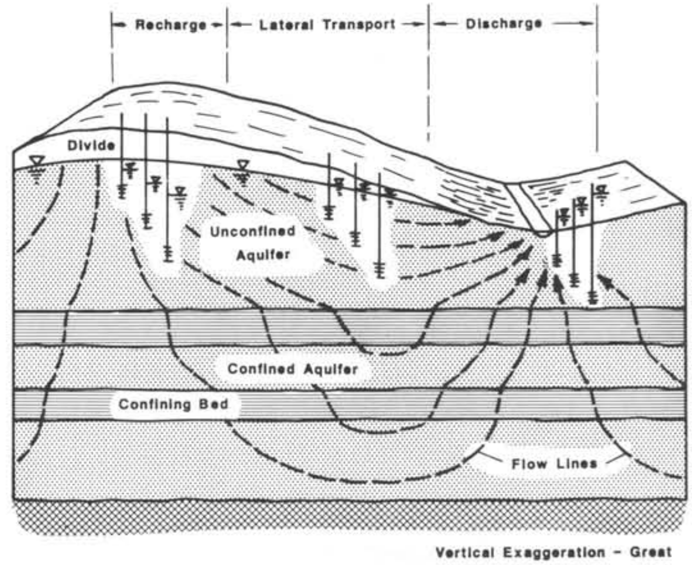

Figure 3. The generalized subsurface flow system.

shallow well will stand. Below the water table is ground water that will discharge through springs, seeps, and free-flowing or pumping wells. There can be more than one water table in a vertical section due to ground water perching on clay layers, fragipans or other discontinuities in hydraulic conductivity. Also, confined aquifers, discussed subsequently, will each have their own water table.

Figure 2 indicates terminology and processes of importance within the root zone. Precipitation is the source of water to the subsurface flow system. When

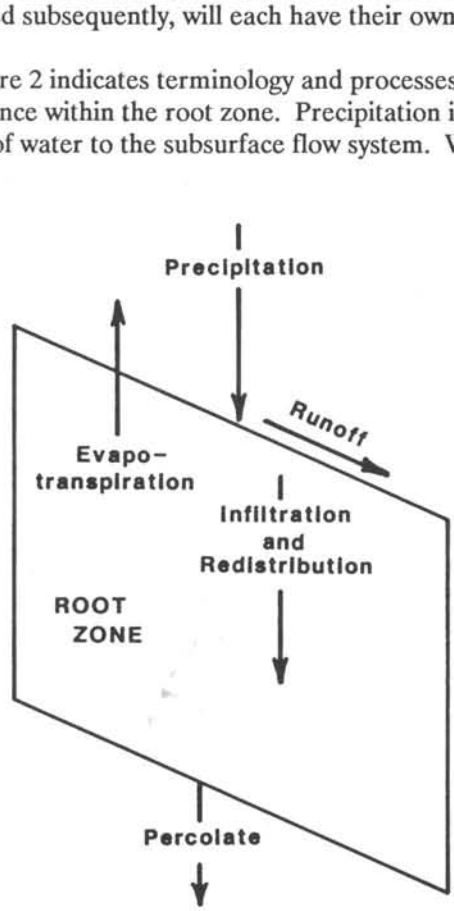

Figure 2. Flow components and processes in the root zone.

precipitation falls, some may become surface runoff and not penetrate the soil. The remainder, which enters the soil, is subject to redistribution depending on soil hydraulic characteristics and the initial moisture distribution within the soil profile. The resultant soil moisture distribution is then subject to the separate effects of evaporation from the land surface, and transpiration from plants dependent on the rooting depth at the site. These separate losses are commonly lumped as evapotranspiration (ET). In general, precipitation which does not run off or is not lost to ET continues moving downward past the bottom of the root zone as percolate.

That portion of precipitation which eventually moves to the water table is called ground water recharge. Recharge generally occurs where unsaturated conditions exist, and these areas are termed recharge zones as shown on Fig. 3. After reaching the saturated zone, water moves downward, laterally, and upward as the result of hydraulic gradients determined by flow system boundaries, material properties, and amount of recharge. Eventually, the water arrives at a ground water discharge area. Saturated units with hydraulic conductivity and porosity great enough to supply usable amounts of water are termed aquifers, while less conductive layers are termed confining or retarding layers (also aquicludes or aquitards). Where bedding is nearly horizontal, confining layers impede the vertical movement of water, and the more active movement of ground water tends to be in the shallower layers. In this paper, the term aquifer applies to any saturated unit with active subsurface flow.

In most areas, the land surface is underlain by relatively permeable layers, which are in turn underlain by less permeable layers. The surficial layers range in thickness from a few meters to over a hundred meters. Ground water moving through these layers forms an unconfined or water-table aquifer, i.e., an aquifer in which the water table forms the upper boundary. The water table in unconfined aquifers varies during the course of the year, depending on the rate of recharge. If water percolates through a confining bed to a more permeable zone below, or the permeable zone below receives water from distant areas where it is exposed at the land surface due to tilted beds, that zone becomes a confined aquifer. The water level in a confined aquifer is controlled by the overlying confining bed, but its energy head may be substantially above this elevation. As shown in Fig. 3, energy levels in confined aquifers will vary depending on their position in the flow system. In zones of recharge, heads will tend to decrease in the downward direction, while in zones of discharge, heads will increase downward. Water moves between confined and unconfined aquifers depending on the energy heads and hydraulic conductivities of the intervening confining layers.

GROUND WATER USE IN THE NORTHEAST

In 1980, withdrawals from ground water represented about 40% of all fresh water use in the U.S., discounting hydropower generation and power plant cooling. While not as great a part of total water use in the Northeast (NE), ground water is nonetheless important. For purposes of this paper, the Northeast will be considered to include Maine, New Hampshire, Vermont, Massachusetts, Rhode Island, Connecticut, New York, New Jersey, Pennsylvania, Maryland, Delaware, Virginia, West Virginia and the District of Columbia. Table 1 details selected aspects of water use within these states, particularly as related to ground water. Ground water withdrawals range from a low of 4% of total water use in West Virginia to a high of 59% in Delaware. The average over the entire region is approximately 10%. The population served by ground water ranges from 24% in Rhode Island to about 60% in both Delaware and New Hampshire, with an average over the region of 39%. The most striking feature of these data is the rural population

Table 1. Water use in the Northeast (from USGS 1984).

	CT	DE	ME	MD/DC	MA	NH	NJ
Area sq km	12,600	5,000	80,300	25,600	20,300	23,300	19,300
Pop. in 1,000's	3,100	590	1,100	4,300	5,800	920	7,600
Fresh Water Withdrawn (million liters/day)							
Total	4,900	530	3,200	6,400	9,500	1,400	11,000
Ground Water (GW)	570	310	300	670	1,200	250	2,800
GW % of Total	11	59	9	10	13	17	25
Millimeters of GW	15	23	1	10	23	5	53
Ground water							
% Pop. Served	32	60	57	30	33	60	45
% Rural Pop. Served	98	100	100	98	100	98	100

	NY	PA	RI	VT	VA	WV
Area sq km	123,000	116,000	2,750	24,000	103,000	62,400
Pop. in 1,000's	17,600	11,900	940	510	5,340	1,960
Fresh Water Withdrawn (million liters/day)						
Total	30,000	61,000	640	1,300	21,000	21,000
Ground Water (GW)	3,700	3,800	140	170	1,400	830
GW % Total	12	6	22	13	7	4
Millimeters of GW	10	13	18	3	5	5
Ground water						
% Pop. Served	35	44	24	54	41	53
% Rural Pop. Served	100	85	100	95	100	95

served by ground water: it is very nearly 100% for the entire NE. These figures represent almost 30% of the U.S. population but only 8% of its land area (excluding Alaska). In the NE, over 24 million people rely on ground water, and 11 million of these are in rural areas.

An interesting figure is derived if annual ground water withdrawals are examined as though evenly abstracted from the area of each state. Withdrawals average approximately one mm uniformly over the area of Maine, to greater than 50 mm over New Jersey. On average over the NE, the withdrawal is about 10 mm, showing that ground water is valuable to the well-being of the region.

Tables 2 and 3 detail types and sources of ground water contamination in states in the NE as perceived by EPA (1985). The information was garnered from a variety of sources by a variety of people, so there is some inconsistency. However, the general characteristics of ground water contamination in the NE are apparent. Table 2 shows that organic chemicals, nitrates, and pesticides are the dominant ground water contaminants. Table 3 shows the dominant sources of ground water contamination to be septic tanks, municipal landfills, industrial landfills, underground storage tanks, and agricultural activities. Not surprising, the more northern states also report ground water contaminated by highway deicing.

Thus, while ground water is important to the region, it is also impacted by man's activities. The impacts range from small-scale local incidences of contamination by underground storage tanks and landfills, to larger-scale areal contamination by agricultural activities. The contaminants are both organic and inorganic. Analysis and management of ground water problems in the NE must be able to respond to these different scales and chemicals.

Selected facts specific to states are detailed in the following section to further illustrate the importance of ground water (abstracted from USGS 1984). While the narrative is specific for each state, many of the statements apply to more than one state or to the entire NE.

Connecticut

Ground water is an increasingly important resource because of several factors: land for additional surface reservoirs is scarce, cost of developing and operating surface water supplies is large, and state policy favors

Table 2. Major types of ground water contamination in the Northeast (from EPA 1985).

	CT	DE	ME	MD/ DC	MA	NJ	NY	RI	VT	WV
Organic Chemicals	X		X	X	X	X	X	X	X	X
Nitrates		X	X	X	X	X	X	X	X	
Brine/Salt						X			X	
Metals			X		X	X		X	X	
Pesticides	X		X		X	X	X	X	X	

NH, PA and VA - no report.

Table 3. Major sources of ground water contamination in the Northeast (from EPA 1985).

	CT	DE	ME	MD/ DC	MA	NH	NJ	NY	PA	RI	VT	VA	WV
Septic Tanks			X	X	X	X	X	X	X	X	X	X	
Landfills	X	X	X	X	X	X	X	X	X	X	X		X
UG Storage Tanks	X	X	X	X	X	X	X	X		X	X	X	
Haz. Waste Sites			X	X	X		X	X		X	X		
Salt Water Intrus.		X	X	X			X						
Agriculture	X	X	X	X	X		X	X		X	X		
Highway Deicing	X		X	X	X	X	X			X	X		

development of aquifers for future supplies. Principal aquifers are susceptible to contamination because of their shallow depths and thin or permeable overburden. Although recharge rates are variable, the long-term average ranges from 180 to 500 mm annually, and occurs mainly during the dormant season. Because of the shallow aquifer depths and the moderate topographic relief, ground water circulation is generally localized within each basin drained by a perennial stream; larger regional flow systems are present within the sedimentary rock of the larger valley lowlands. Stratified drift aquifers are particularly susceptible to contamination, and ground water quality has been affected locally in almost all regions of the state (Rolston et al. 1979). Major sources of ground water contamination are detailed in Handman et al. (1979). Because of the widespread dependence on induced recharge to sustain withdrawals from stratified drift aquifers, the most significant impact of development is depletion of streamflow. Withdrawal rates are generally greatest in summer when ground water and streamflow levels are relatively low, and consequently, some smaller streams may dry up completely.

Delaware

Ground water is the primary source of public, rural, and industrial supply in approximately 95% of the state by area. Only the northmost 5% is supplied predominantly by surface water, but this area contains about 40% of the state's population. Most of the 1,090 mm of precipitation evaporates, is transpired, or runs off to streams. Johnson (1973) estimated that only about 360 mm of precipitation recharges the ground water system annually. Water levels near major pumping centers are generally declining.

Maine

The greatest area of pumping is in southwestern and coastal Maine. The quality of ground water is suitable for most uses, but local contamination has occurred from point sources such as storage tanks, salt-storage sites, and landfills, and nonpoint sources such as agriculture and deicing. Precipitation is sufficient to replenish the water pumped from Maine's aquifers, and while annual fluctuations in water levels are large, long-term depths to the water table are relatively stable. In the southwestern coastal region, some water supply shortages are occurring due to the increasing demands of a large summer tourist population and a steadily increasing year-round population. Water supply shortages are predicted for nearly 60% of the towns in coastal Maine by the year 1990 (Caswell and Ludwig 1978).

Maryland

Annual precipitation ranges from about 940 to 1,190 mm, and recharge rates vary, with about 1/4 to 1/3 of the precipitation reaching the water table. A small part of the recharge moves to deeper aquifers; most discharges to nearby streams and provides 50 to 70% of their long-term flow. Ground water levels in the confined coastal aquifers have declined with major pumping, and there has been induced movement of brackish water into some coastal aquifers. Water levels inland are generally stable.

Massachusetts

Contamination and drought have affected the ground water in the past. Degradation of ground water quality by wastes and chemicals have caused water shortages. Between 1978 and 1981, 25 public-supply wells were taken out of service because of contamination. Drought caused by deficient precipitation resulted in 38 communities declaring water shortages in 1981. Ground water recharge is directly from precipitation of about 1,120 mm. Recharge is estimated to be about 500 mm through soils developed on sand and gravel, and about 150 mm on soils developed on glacial till. Most ground water is pumped from wells less than 100 m deep. Long-term decline has not been observed in any of the state's aquifers, but seasonal water level variations may be large. Generally, little recharge occurs during the 180-day summer growing season.

New Hampshire

The quality of ground water is generally suitable for human consumption; 80% of ground water withdrawals in 1980 supplied drinking water systems. Locally, the chemical quality of ground water may reflect overlying land-use practices. Degradation of quality occurs near unsewered residential and village areas, near underground storage tanks and waste-disposal sites, and near agricultural lands and highways. Average precipitation is about 1,090 mm, and recharge rates range from 360 to 510 mm. Primary pumpage is in the more populous southern half of the state, but progressive long-term declines in water levels have not been documented.

New Jersey

From both the quantity and quality view, the ground water in New Jersey is under the most stress in the NE. An average precipitation of 1,120 mm results in recharge varying from 380 to 990 mm across the state. Since the 1900's, ground water withdrawals have increased and resulted in regional declines in water levels in several aquifers in the state. Many water-supply problems associated with increased withdrawal rates have been documented, including declining levels in confined aquifers, and resultant movement of brackish or saline water from surface water bodies or adjacent aquifers. An observation well in Camden County in southwest NJ near a major water supply and industrial pumping center has declined about 15 m from 1965 to 1983, a head loss of over one m/yr. Ground water level declines of 6 m between 1965 to 1983 were common around the state, while a public supply well in east-central Monmouth County declined about 30 m over the same period.

New York

Of the more than six million residents relying on ground water, more than half live on Long Island. Precipitation on Long Island is about 1,090 mm annually, and about 50% of this reaches the water table. In the remainder of the state, precipitation ranges from 810 to 1,270 mm, and recharge rates vary from none to 50% of precipitation. Most major pumping in Upstate New York is from valley fill aquifers, so water level decline is minimal. However, on Long Island, withdrawals have steadily increased since the late 1930's. Declines were such that remedial measures involving recharge via wells and pits were instituted to counteract salt-water intrusion into the supply aquifers.

Pennsylvania

Precipitation averages 1,120 mm, and about 55 to 60% of this falls during the warm half of the year, most during intense rainstorms. Although ground water use is greatest in the population centers of the southeast and southwest, the percentage of users dependent on ground water is greatest in the rural areas, almost 100%. Ground water levels over the state are generally stable.

Rhode Island

Ground water reserves appear adequate to meet a substantial part of the state's projected water needs. However, because of high aquifer permeability and general shallow depths to the water table, ground water is susceptible to contamination. Local contamination has resulted from leaking gasoline tanks, leaching from landfills and salt storage piles, and agricultural activities. Spills of organic solvents have contaminated both public-supply wells and domestic wells. Precipitation ranges from 1,070 to 1,220 mm, and recharge is estimated to be 200 to 230 mm in areas of till and 530 to 630 mm in areas of stratified drift. Most major withdrawals are from pumping centers within hundreds of meters of a stream, and withdrawals are generally replenished with recharge induced from the streams. Thus, overall ground water levels are relatively stable.

Vermont

The quality of ground water is generally suitable for most purposes, but locally the chemical quality may reflect overlying land uses. Contamination has been observed near unsewered villages and ski areas, underground storage tanks, and waste disposal sites, agricultural lands, and highways. Annual precipitation is about 1,040 mm, and recharge is probably in the range of 300 to 510 mm. Annual ground water levels respond to seasonal differences in recharge, natural discharge, and pumping rates, but monitored wells reflect a dynamic equilibrium with no long-term progressive declines in water levels observed.

Virginia

Ground water is an important source of industrial and public supply for both the Eastern Shore Peninsula and many rural areas. Most major metropolitan areas rely on surface water supplies. Precipitation ranges from 910 to 1,270 mm, and recharge varies from about 200 mm in the west to about 250 mm in the coastal plain. Ground water discharges to local streams and maintains flow during periods of little or no precipitation. Continued withdrawal of ground water has caused a steady decline of water levels and expansion of the cones of depression around major pumping centers in the eastern part of the state. In contrast, little change has occurred in a confined aquifer under the Eastern Shore peninsula, but water has moved downward from the overlying unconfined aquifer to counteract the pumpage.

West Virginia

In the alluvial and limestone areas of the east, ground water is plentiful and potential for further development is good. In the western two-thirds of the state however, ground water is less plentiful, and generally, only small quantities are obtainable. Precipitation ranges from about 760 to 1,520 mm. Recharge ranges from 50 to 150 mm in areas underlain by shale to about 150 to 300 mm in areas underlain by sandstone and limestone (Hobba 1985). A major percentage of the amount that recharges ground water discharges to nearby streams; very little moves into deeper aquifers. Overall, ground water levels in the state are relatively stable, showing no long-term declines due to pumping.

GROUND WATER REGIONS OF THE NORTHEAST

While some aspects of ground water related to demographics are readily examined in terms of political boundaries, the hydrogeologic aspects of ground water in the NE are more conveniently examined in the context of regions where features affecting the occurrence and availability of ground water are similar. These features are:

1) components of the flow system and their arrangement

2) nature of the dominant water-bearing openings of the aquifer

3) mineral composition of the rock matrix
4) water storage and transmission characteristics of the aquifer
5) nature and location of recharge and discharge areas

A number of such classifications have been made (e.g., Meinzer 1923; Thomas 1952), but that proposed by Heath (1984) will be employed here. The NE contains parts of Heath's regions termed: Nonglaciated Central Region, Glaciated Central Region, Piedmont and Blue Ridge, Northeast (and Superior) Uplands, Atlantic (and Gulf) Coastal Plain, and Alluvial Valleys. The location of these regions (excepting Alluvial Valleys) in the NE is presented as Fig. 4. Alluvial Valleys are not a contiguous region geographically; rather, they are segments of perennial stream valleys underlain by sand and gravel thick enough to be hydrologically significant (thicknesses generally greater than eight m). Alluvial Valleys are shown on Fig. 5. The nature and extent of the dominant aquifers and their relations to other units of the ground water system are the primary criteria used in delineating the regions. Consequently, boundaries of the regions generally coincide with major geomorphic units, not with surface drainage divides. This lack of coincidence emphasizes that the hydrologically important physical characteristics of the subsurface and surface flow systems are controlled by different factors, but it does not imply that the systems are not interrelated. Ground water and stream systems are intimately related in the NE, as will be seen throughout the remainder of this paper. A brief discussion of each region follows, while common ranges of their hydraulic characteristics are given in Table 4.

Nonglaciated Central Region

The region is characterized by thin regolith underlain by fractured sedimentary bedrock ranging in age from Paleozoic to Tertiary, and consisting largely of sandstone,

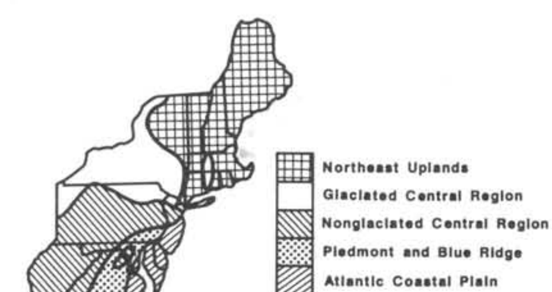

Figure 4. Ground water regions of the Northeast (after Heath 1984).

Figure 5. Alluvial valleys of the Northeast (after Heath 1984).

shale, carbonate rocks (limestone and dolomite), and conglomerate. It is a topographically and geologically complex area, and with the principal exception of the Ridge and Valley Province, the rock layers are horizontal or gently dipping. In most of the region the regolith is formed by mechanical or chemical breakdown of the bedrock.

The principal water-bearing openings in the bedrock are fractures which generally occur in three sets. The set of most importance to ground water flow was developed along the contact between different layers, i.e., bedding planes. Where the layers are horizontal, these fractures are roughly parallel to the land surface. The two remaining fracture sets are essentially vertical and cross the bedding-plane fractures at a steep angle. The vertical fractures enable water to move across the rock layers and serve as the connection between the bedding plane fractures.

Where the bedrock has been folded or bent (i.e., the Ridge and Valley Province), the occurrence and orientation of the fractures is more complex. The dip of the rock layers and associated bedding plane fractures ranges from horizontal to vertical, and fractures parallel to the land surface are probably less numerous and of a more limited extent than where the rock layers are horizontal. Openings developed along the fractures are usually less than one millimeter wide. The principal exception to this is in the limestones and dolomites, where water moving through the original fractures may enlarge them to form, at the extreme, extensive cavernous systems capable of transmitting large amounts of subsurface flow.

Recharge of ground water in this region generally occurs in the outcrop areas of the bedrock aquifers in the

Table 4. Common ranges of hydraulic characteristics of the dominant aquifers in the Northeast (from Heath 1984).

Region	Hydraulic Conductivity m/day	Recharge Rate mm/yr	Well Yield cu m/min
Nonglaciated Central	3-300	5-500	0.4-20
Glaciated Central	2-300	5-300	0.2-2
Piedmont and Blue Ridge	0.001-1	30-300	0.2-2
Northeast Uplands	2-300	30-300	0.1-1
Atlantic Coastal Plain	3-100	50-500	0.4-20
Alluvial Valleys	30-2,000	50-500	0.4-20

uplands between streams. Discharge from the ground water system is by springs, seepage areas, and direct inflow to the stream bed, and by evaporation and transpiration in near-stream areas where the water table is near the land surface. Ground water circulation is limited at depths greater than about 100 m due to the decrease in frequency and size of fractures.

Glaciated Central Region

The region is characterized by glacial deposits over fractured sedimentary rock. The relatively flat-lying consolidated sedimentary bedrock ranges in age from Paleozoic to Tertiary, and consists mainly of sandstone, shale, limestone and dolomite. Bedrock is overlain by glacial deposits, which, in most of the area, consists of till, an unsorted mixture of rock particles, sand, and clay deposited directly by the ice. The till may be interbedded with, and overlain by, sand and gravel deposited by meltwater streams, or by silt and clay deposits from glacial lakes.

Ground water occurs in both the glacial deposits and in the bedrock. In the glacial deposits, water is in the pore spaces between grains, while in the bedrock, water occurs primarily in the fracture system. The fracture system is similar to that of the Nonglaciated Central Region. The glacial deposits are recharged by precipitation on the interstream areas, and serve as a source of water to shallow wells and to the bedrock below. Recharge occurs primarily in the fall when plant growth stops and evaporation is reduced, and again during the spring thaw, usually the most dominant recharge event. Small amounts of recharge may occur during midwinter thaws and during large summer storms. Ground water discharge is by seeps and springs into nearby perennial streams, and active circulation of the ground water flow systems is at depths similar to those of the Nonglaciated Central Region.

Piedmont and Blue Ridge Mountain Region

The region is characterized by thick regolith over fractured crystalline and metamorphosed sedimentary bedrock of Precambrian and Paleozoic age, including granite, gneiss, schist, quartzite, slate, marble, and phyllite. The regolith is a clay-rich unconsolidated material called saprolite which was derived from in situ weathering of the underlying bedrock. It averages 8 to 15 m in thickness, and may be as much as 100 m thick on some ridges. The regolith contains water in pore spaces between grains. In contrast, the underlying bedrock does not have significant intergranular porosity, but contains water in sheetlike openings formed along fractures.

All ground water systems function analogously to reservoirs which store water, and to pipelines which transmit water from recharge to discharge areas. The Piedmont and Blue Ridge Region can be viewed as having a ground water system in which the reservoir and pipelines are separated. The regolith, with its larger porosity, stores water and feeds it downward into the fracture system of the bedrock. The fractures act as a pipeline to transmit water laterally through the system. Recharge occurs in the upland areas removed from the flood plains of the streams, and discharge occurs as seeps and springs near the bases of slopes and along streams. Valleys and other more minor surface depressions generally indicate zones of bedrock more intensively fractured and more susceptible to weathering and erosion than those of the uplands. These zones act as ground water collectors and discharge points.

Northeast and Superior Uplands

The region is typified by glacial deposits over fractured crystalline rock ranging in age from Precambrian to Paleozoic, and consisting primarily of granite, syenite, anorthosite, and other intrusive igneous rocks, and metamorphosed sedimentary rocks such as gneiss, schist, quartzite, slate, and marble. Most of the igneous and metamorphosed sedimentary rocks have been folded and cut by numerous faults. Bedrock is overlain by unconsolidated deposits laid down by ice sheets which covered the area one or more times during the Pleistocene, and by sand, gravel, silt, and clay laid down by meltwater streams and lakes. The most extensive deposit is till. In most of the valleys, the till is covered by glacial outwash of interbedded sands and gravels. In other areas, including parts of the Champlain Valley, the deposits consist of clay and silt deposited in lakes formed during the melt periods.

Ground water occurs in both the glacial deposits and in the bedrock. The glacial outwash deposits, when large enough, are considered under the Alluvial Valley section, but even smaller outwash deposits are significant sources of water locally. Water in the bedrock occurs in a fracture system similar to that of the Piedmont and Blue Ridge Region in frequency, occurrence, and origin. Recharge begins after the cessation of plant growth in the fall, continues intermittently during winter thaws, and is maximum between the start of the spring and the start of the growing season. As in the Piedmont and Blue Ridge region, the glacial deposits serve as a ground water reservoir, and the bedrock fracture system serves as a pipeline from recharge to discharge zone. Because they are nonresidual, the glacial deposits tend to obscure the more fractured zones in the bedrock.

Atlantic and Gulf Coastal Plain

The region is characterized by complexly interbedded sands, silts, and clays transported by streams from the adjoining uplands. The sediments range in age from Jurassic to the present, and range from a few meters thick near the most inland edge to thousands of meters thick off the coast. The sediments were deposited on flood plains and as deltas where inland streams reached the coast, and were reworked by waves and ocean currents during different invasions of the onshore areas by the ocean; thus, the complex interbedding. The formations dip toward the coast, so that outcrops form a series of bands roughly parallel to the coast. Within any formation, the more coarse-grained materials tend to be most abundant near the source areas, while the clay and silt layers become thicker and more numerous downdip.

Recharge to the system occurs in the interstream upland areas where sand layers outcrop, and by downward flux through the interbedded clay and silt layers. Discharge occurs by seepage to streams, estuaries, and the ocean floor. Movement of water between the recharge and discharge zones is controlled by hydraulic gradients as in all ground water systems, but here, the pattern is altered by the downgradient thickening of clay which limits upward discharge. This condition results in confined aquifers at depth near and under the ocean, and major convergence of flowlines to the land surface where major streams cross the downdip portion of outcrop areas.

Withdrawals by pumping from ground water near the outcrop areas are quickly counteracted by increased recharge or reductions in natural discharge. Withdrawals at significant distances downdip have no effect on conditions in the recharge areas; rather, they result in lowering of head and removal of water from storage in the aquifer and confining beds.

Alluvial Valleys

These consist of relatively thick sand and gravel deposits beneath the floodplains and terraces of streams. In the preceding descriptions, streams and other surface water bodies were points of ground water discharge. Throughout the NE, ground water and stream systems are usually so interconnected that a change in one results in a change in the other. If ground water is withdrawn near a channel, gradients normally toward the stream may be reversed, causing water to move into the aquifer from the stream. In other regions of the NE, the amount of water entering the aquifer because of this gradient reversal would be small due to relatively low hydraulic conductivities of the aquifers and poor interconnection. In the Alluvial Valleys, this is not so. By definition, alluvial valleys have: 1) sand and gravel deposits thick enough to supply water to wells at moderate to large rates, 2) these deposits are in hydraulic contact with a perennial stream which can serve as a source of recharge and which has a flow normally much greater than typical well field demand, and 3) the deposits occur in a well-defined band that normally does not extend beyond the floodplain or adjacent terraces. Under natural conditions, alluvial deposits are recharged by precipitation directly on the valleys, by ground water discharging from adjacent and underlying aquifers, and by overbank flooding of the stream. The alluvial valleys then discharge to the stream. However, when these aquifers are pumped, discharge to the stream is reduced, and if the withdrawals are large enough, water moves from the streams into the deposits. Because of the contrast in conductivities, excess pumping from the alluvial valleys usually has little effect on water levels in adjacent aquifers.

REGIONAL COMMONALITIES

Although the ground water regions of the NE are different in specifics, they contain commonalities in both structure and hydrologic performance which allows them to be grouped for analysis.

The Glaciated Central region resembles both the Nonglaciated Central and Northeast Uplands regions in certain aspects. Bedrock in the Glaciated and Nonglaciated regions is similar in both composition and structure; the primary difference between the regions is composition of the overlying unconsolidated material. Hydrologically, the unconsolidated material in the Nonglaciated region is of minor importance, both as a source of water and as a reservoir for water storage. In contrast, the glacial deposits of the Glaciated region can serve both as a source of water and as a storage reservoir for flow to the bedrock.

The Glaciated and Northeast Uplands regions are similar in that the unconsolidated material of both regions consists of glacial deposits, but the bedrock is dissimilar. The Piedmont and Blue Ridge Region and the Northeast Uplands are similar in bedrock, but dissimilar in composition of material overlying the bedrock.

Yet, all four regions (Glaciated, Nonglaciated, Piedmont and Northeast Uplands) exhibit a common generalized structure, i.e., a layer of regolith (limited in the Nonglaciated Region) overlying fractured bedrock. The regolith acts as a storage reservoir providing flow to the fracture system of the underlying bedrock, and in some cases, it serves as a source of water supply as well.

The Coastal Plain and Alluvial Valley regions do not fall into this classification, but they do share some common characteristics. While of different scales and composed of differing materials, each consists of relatively self-contained flow systems in which the aquifer material is as classically envisioned, i.e., relatively high conductivity porous media bounded by "confining" layers of clay in the case of the Coastal Plain, or bounded by other regions in the case of the Alluvial Valleys.

The above commonalities result from the geologic structure, but a greater commonality emerges when the geologic settings are combined with the humid temperate climate of the NE, and the significant precipitation occurring relatively uniformly throughout the year. As a result of this combination, ground water flow systems of the NE generally share the following features:

1) the unconfined aquifers at the top of the geologic profile are relatively shallow and have relatively high water tables

2) distances between recharge and discharge zones tend to be small

3) the shallow unconfined aquifers are refilled completely during the dormant season, especially during spring snowmelt

4) springs and seepage zones play a major part in discharging ground water to the stream channel systems and result in water tables near the land surface in near-stream zones

5) a relatively high percentage of streamflow is derived from subsurface sources

6) perhaps most importantly, there is an intimate and relatively immediate connection between soil water, percolate to shallow ground water, the ground water flow system, and ground water discharge to the stream; i.e., response to precipitation and travel times through most unconfined subsurface flow systems tend to be relatively quick.

Figure 6 illustrates these latter features on a seasonal basis using data from a 7.2-km^2 watershed in central Pennsylvania. Precipitation, the input to the system, is almost uniformly distributed throughout the year. The other driving mechanism and major loss from the system, ET, varies between minimum values during the dormant season and maximums during the growing season. Soil moisture, the prime source of water for ET and the major control on ground water recharge, is the virtual inverse of ET. Ground water levels respond directly to the reduction in recharge resulting from the reduction of soil moisture by ET, reaching their minimum at the end of the growing season. The decrease in ET and corresponding increase in soil moisture in early fall results in an immediate increase in ground water level. Lastly, the almost one-to-one correspondence of monthly streamflow and ground water level shows the direct and immediate influence of the ground water flow system on streamflow leaving the watershed.

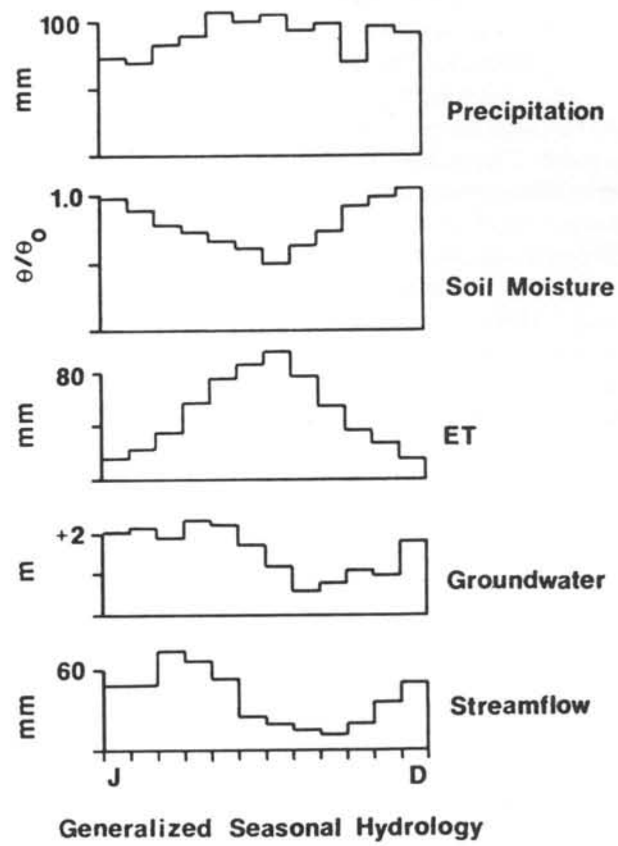

Figure 6. Seasonal relationships between components of the hydrologic cycle in the Northeast.

Another characteristic of NE subsurface flow systems is that they are dominated by heterogeneities, both at the small and large scale. At the small scale, variabilities in hydraulic properties are exhibited by the interbedded deposits of the coastal plains and the alluvial valleys, and by the fracture system within fractured bedrock. At the large scale, the NE has entire flow systems dominated by the full range of porous media: clays (saprolites, glacial, and confining beds in coastal conditions), sands and other "classical" porous media (coastal, glacial, and alluvial valleys), fractures (most bedrock), and solution channels (certain limestones and dolomites).

Man's activities in the NE also result in heterogeneities. If we superimpose these on the characteristic subsurface flow system, we get a sense of the scale or scales at which we must be able to characterize, analyze, sample, and model ground water flow. There are small- and large-scale heterogeneities in population distribution within the NE. The Atlantic Coast "megalopolis" is a dominant large-scale variation. At the smaller scale of the urban fringe, we find agricultural land use abutting unsewered suburban development, which, in turn may adjoin a manufacturing plant with on-site waste disposal landfills or lagoons.

Tables 2 and 3 showed that man's influence in the NE has manifested itself in both point source contamination of ground water, such as leaking storage tanks and landfills, and nonpoint source contamination associated with agricultural activities and highway deicing. For purposes of ground water management and protection, we must be able to analyze and separate the effects of these various factors on the variety of NE subsurface flow systems. We must develop the capability to describe ground water systems at both the local and the areal scale. Further, because of the intimate connection between ground water and stream systems, and the fact that stream quality is critical to both water supply and ecologic considerations as in the Chesapeake Bay, we must be able to adequately describe the contact between ground water and streamflow. Lastly, contaminants of concern are generally introduced to the system within the soil zone or relatively near the land surface. Thus, we must also have the capability to predict the movement of water and chemicals from the point of introduction of the contaminant down to and through the ground water body. Since increasing emphasis is being placed on land disposal of wastes, this latter need becomes critical.

MODELING NORTHEASTERN GROUND WATER FLOW SYSTEMS

As a result of the need for multiscale considerations, analyses of entire ground water flow systems are necessary, including their interactions with both recharge zones and discharge sites. Conversely, selected portions of the systems must sometimes be isolated for detailed analysis. The fact that contaminant transport within these systems is also important requires that we be concerned not just with flow volumes, but rather with flow pathways, times of travel along these pathways, and interactions between the contaminants being transported and that portion of the media through which they are moving.

The tool for these analyses is a model. A model can range from simple analytic descriptions of pumping tests to complex three-dimensional numerical solutions of large-scale heterogeneous subsurface flow systems. To model any subsurface flow system, the following must be determined:

1) equations of flow and transport applicable to the flow system and the problem
2) size and shape of the region of flow important to the problem
3) boundary conditions around the region
4) spatial distribution of parameters that control flow and contaminant transport
5) initial conditions within the region (for transient solutions)
6) a mathematical model of solution

Factors 2) and 3) involve defining geometry of the flow system, while 4) and 5) involve flow system testing, monitoring, and sampling. Further, the subsurface flow system may again have to be sampled for model calibration, verification, or validation during and after the modeling process. A brief overview of the above factors in the context of NE ground water systems follows.

Equations of Subsurface Flow and Contaminant Transport

Based on characterizations of NE ground water systems, we must analyze subsurface flow and transport at the large (watershed) and the small (contaminant plume) scale. Additionally, we must analyze flow and transport in layered systems consisting of conventional (granular) porous media and a medium of the general type discussed earlier by Wagenet and Germann (1989) and Rogowski (1989), i.e., one having preferential flow pathways.

All basic ground water texts adequately treat flow through granular media. But recently, the importance of fracture systems in subsurface flow and their need to be considered separately has been recognized. Therefore, the balance of this section will emphasize analysis of fractured systems.

Rock unit properties resulting from a fracture system are termed "secondary" properties. The increase in permeability associated with a fracture system is secondary permeability, in contrast to intergranular, or matrix permeability. Porosity of the fracture system is secondary porosity, as compared to primary porosity, that

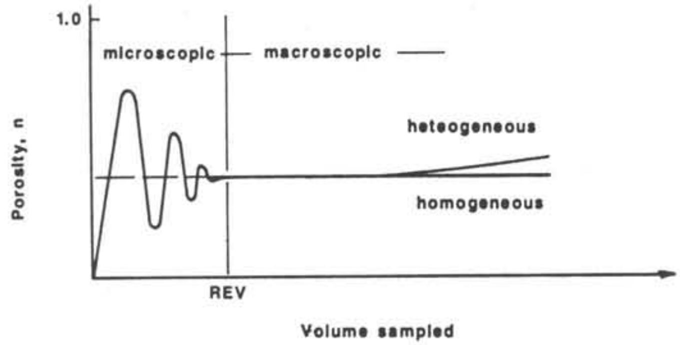

Figure 7. The representative elementary volume (REV).

of the matrix. Secondary permeability is usually large compared to that of the rock matrix, while secondary porosity may be significantly smaller than matrix porosity. When concerned only with quantities of subsurface flow (i.e., amount of water moving to a well or seeping under a dam), the structure of hydraulic conductivity and porosity within the subsurface medium is irrelevant. However, when concerned with contaminant transport, the structure of these properties as controlled by geometry of the pore and/or fracture system becomes important. Knowledge of flow pathways and velocities along the pathways is necessary to define transport and the environment to which the transported chemical is exposed.

The scale at which a system is being viewed, and the concept of the representative elemental volume (REV) is critical to understanding and analyzing flow in granular and fractured media. Darcy's law describing flow through porous media is a macroscopic law, only applicable to water movement at the macroscopic scale. There is a lower limit to the volume of a porous media to which the law applies. Hubbert (1940) addressed this problem by defining "macroscopic" with the aid of Fig. 7. When the volume of porous medium sampled reaches a size where further increases in volume result in the same porosity, sampling is at the scale of the REV. The REV is a volume large enough to permit a meaningful statistical average of the property of concern, and is usually significantly larger than a single pore within even the most simple porous medium. By observing, sampling, and characterizing a porous medium at a scale at or above its REV, the medium can be analyzed using the continuum approach, whereby the agglomeration of individual grains and pores is replaced by a representative continuum with bulk properties amenable to analysis by macroscopic laws, such as Darcy's. It is of interest to note that increasing the sample size significantly beyond that of the REV introduces the possibility of encountering large-scale heterogeneities which alter the property of the sample. When this happens, there is a need to treat the system as spatially variable.

Flow in most granular media, even relatively heterogeneous till, can usually be analyzed using the continuum approach, while flow in fractured rock can be analyzed using either the continuum or the noncontinuum approach. Using the continuum approach, the fractured medium is replaced by an equivalent representative continuum which can have large-scale spatial distributions of conductivity and porosity similar to those of the fractured aquifer but does not recognize flow within individual fractures. Use of this approach is justified when fracture density is large with respect to the scale of the problem being analyzed; i.e., effects of individual fractures are not critical to analysis of the overall flow field. This approach does not preclude representation of anisotropies or heterogeneities caused by irregularities in fracture spacing or density. Snow (1968, 1969) has shown that both such systems can be analyzed using the continuum approach by representing the flow field with standard porous-media flow analyses and an anisotropic hydraulic conductivity tensor.

If an aquifer has a low fracture density though, and/or the scale of a problem is so small that only a few fractures occur within the area of concern, the noncontinuum approach may be apropos. Here, flow in individual fractures must be considered, and analysis is commonly based on principles of fluid mechanics and the Navier-Stokes equations. For instance, the noncontinuum approach would have to be taken in the karst-type environment, where individual fissures, caves, etc., exert a dominant control on flow. Observing this system at a scale large enough to lump parameters in an overall media property is virtually impossible. Fortunately, problems of flow and contaminant transport at the field scale in the non-karst environment can usually be treated using the continuum approach.

There are two additional factors which must be considered in a continuum analysis of flow in fractured rock. Based on the rate of fluid flow and the width of the fractures, fracture flow may be either Darcian or non-Darcian, each analyzed by different equations. Sharp et al. (1972) present laboratory-derived data supporting a nonlinear flow law for fractured rock, while Wittke (1973) proposes separate flow laws for the linear laminar (Darcy) range, the nonlinear laminar range, and the turbulent range of flow. Hickey (1984) describes a method of field testing the Darcy flow hypothesis in carbonate rock in Florida. Darcy's law is generally considered valid up to Reynolds numbers of 1 to 10, while beyond that, a turbulent flow formulation is usually employed.

A continuum-based approach allows use of the generalized equation of ground water movement applicable to transient flow through a saturated anisotropic heterogeneous porous medium

$$\frac{\partial}{\partial x}\left[K_x \frac{\partial h}{\partial x}\right]+\frac{\partial}{\partial y}\left[K_y \frac{\partial h}{\partial y}\right]+\frac{\partial}{\partial z}\left[K_z \frac{\partial h}{\partial z}\right]=S_s\frac{\partial h}{\partial t} \quad (1)$$

where x, y and z are coordinate directions (L), K is hydraulic conductivity (L/T), h is hydraulic head (L), S_s is specific storage (1/L), and t is time (T). There are a number of analytic and semi-analytic solutions to (1) for simple geometries and boundary conditions, but for complex flow situations, numerical solutions are usually necessary.

The processes governing contaminant transport through any porous media, granular or fractured, are advection, molecular diffusion, mechanical dispersion, and a variety of chemical and biological reactions such as adsorption and degradation. However, transport rates through the two systems differ substantially. Assuming that a porous medium can be characterized using bulk properties to represent its small-scale internal nonhomogeneities, Darcy's equation governing flow through the medium is

$$Q = -KiA \quad (2)$$

where Q is flow (L^3/T), K is bulk hydraulic conductivity (L/T), i is the driving hydraulic gradient (L/L), and A is the cross-sectional area of flow (L^2).

Dividing (2) through by area gives

$$V = -Ki \quad (3)$$

where V is a fluid velocity (L/T) which is termed the Darcy velocity or specific discharge. The point must be made here that (3) gives the flux of water through the entire cross-section, i.e., the volume of water per unit time passing through a unit area consisting of both solids and voids. To calculate water velocity within the pores, porosity is incorporated as

$$V_p = (K/n)\, i \quad (4)$$

where V_p is average linear velocity of water within the pores (L/T), and n is effective bulk porosity of the media (L^3/L^3). Two additional points must be made here. One is that V_p is not velocity within individual pores; rather, it is average within-pores velocity component in the direction of flow (perpendicular to the cross-section). Velocities within individual pores can be higher due to the tortuosity of the pore system. Secondly, the porosity of concern is not necessarily total porosity of the medium; it is the "effective" porosity through which the fluid and contaminant move. For instance, closed pores, and to some extent, dead-end pores, are not part of "effective" porosity, even though they contribute to total porosity of the medium. This latter point leads directly to the disproportionate effect of fractures on contaminant transport. To examine a fractured medium using (4), the continuum assumption must first be acceptable; i.e., parameters in (4) must be derived from a fractured medium volume sufficiently large to be described by bulk properties. Individual fractures contributing to K and n must be small and closely spaced as compared to the volume of the medium over which K and n are measured. As stated previously, fractured rock usually has a relatively high hydraulic conductivity, but its effective porosity is usually quite small. Values of effective porosity on the order of 0.001% to 1.0% are typical of fracture systems, whereas matrix (total) porosity may be an order of magnitude greater. These small effective porosities coupled with high K-values result in high velocities of contaminant transport.

As an illustration, consider two hypothetical media, one a sand with a hydraulic conductivity of 0.3 m/day and a porosity of 30%, and the other a fractured media with the same K but an effective porosity of 0.1%. These typify a fractured sandstone reported by Urban and Gburek (1986). Using a hydraulic gradient of 0.02, a value commonly observed in the field, application of (4) to the sand gives an average linear pore velocity of approximately 0.02 m/day. In the fractured sandstone though, (4) gives an average linear in-fracture velocity of 6.0 m/day, over two orders of magnitude greater. In each case, the Darcy velocity, or specific discharge, is the same, 0.006 m/day.

If the entire range of typical fracture porosities is considered, the potential for fracture systems to "accelerate" chemical transport through an aquifer is indeed large. Further, (4) gives only average in-fracture velocity. To complicate the issue, velocities in individual fractures may be orders of magnitudes above and below this average value, depending on fracture size and roughness variations.

Analysis of contaminant transport in fractured media at the field scale commonly resorts to use of the advection-dispersion equation used for analysis of transport in granular media. The two-dimensional equation describing transport of a nonreactive contaminant through a heterogeneous isotropic saturated porous media is

$$\frac{\partial}{\partial s_l}\left[D_l \frac{\partial C}{\partial s_l}\right]+\frac{\partial}{\partial s_t}\left[D_t \frac{\partial C}{\partial s_t}\right]-\frac{\partial}{\partial s_l}(\bar{v}_l C)=\frac{\partial C}{\partial t} \quad (5)$$

where s_l and s_t are the directions of the ground water flow lines and the normals to these lines respectively, D is the dispersion coefficient (L/T), and C is concentration (M/L^3). For homogeneous media, steady-state flow, and simple boundary conditions, this equation can be solved analytically, as can (1). For more complex flow situations though, numerical solution techniques must be used. The reader is referred to Ogata (1970) and Bear (1972) for derivations and simple solutions of the advection-dispersion equation.

Granular media are usually assumed isotropic with respect to dispersivity; i.e., the longitudinal dispersion coefficient has a single value regardless of the direction of the velocity vector. Transverse (lateral) dispersion coefficients then have a single smaller value, usually related to the longitudinal value. Differences between the two values are a function of the mechanisms of dispersion rather than properties of the media. Fractured media, on the other hand, are usually anisotropic with respect to the frequency and orientation of the fracture pattern. Thus, the assumption of isotropic dispersivity may not properly describe dispersion within a fractured medium. While this is recognized qualitatively, quantitative treatment of dispersion in a fractured media is difficult because of its dependency on the fracture pattern. The end result is that fractured media are commonly treated as isotropic (see Schwartz et al. 1983; Rasmuson 1985; Ross 1986), or dispersion coefficients are determined directly by calibration or two-hole pumping tests (e.g., Claassen and Cordes 1975).

Increasingly complex versions of (5) are available to handle more complex contaminants, such as those which are reactive with the media, or those which undergo degradation (e.g., pesticides, radioactivity). Laboratory or theoretical model studies have been used to develop distribution coefficients which describe the migration of such contaminants through fractured systems (see Burkholder 1976). Under very limiting simplifying assumptions usually involving smooth-sided, uniform-aperature, planar fracture systems, the coefficients can be extended to the field scale, but the complexities of natural fracture system geometries make such extensions unrealistic. So field studies in fractured media commonly use bulk reaction rates or degradation rates developed from calibration or observations at other sites.

Discussion to this point has assumed a fracture system in which matrix porosity and conductivity are insignificant. Yet, in some of the more common fractured media of the NE, such as fractured sandstones, there is an added complexity. While the effects of the fracture system may override those of the matrix in the short time frame, matrix K and S_s become important at longer time scales and larger space scales. This topic was treated in depth by Grisak and Pickens (1980) and Grisak et al. (1980).

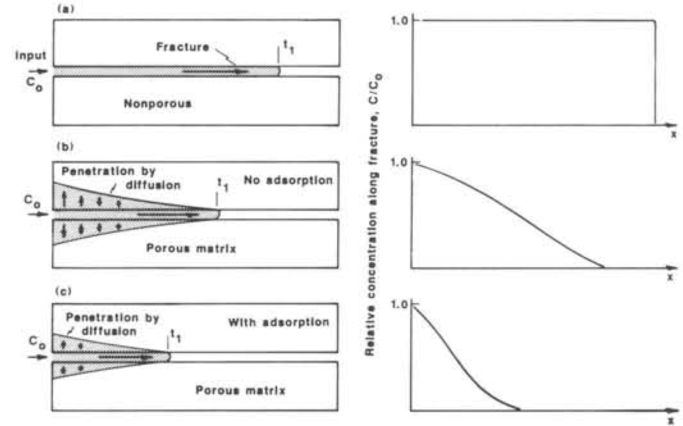

Figure 8. Effect of matrix diffusion on contaminant migration in porous fractured media (after Freeze and Cherry 1979).

Interactions between fracture and matrix flow are summarized in Fig. 8 (adapted from Freeze and Cherry 1979). Assuming that dispersion within the fracture is insignificant, comparison of Fig. 8 (a) and (b) shows that diffusion of the solute into the matrix causes the concentration within the fracture to gradually diminish between the constant concentration input and the front of the contaminated zone. Contaminant transport within the fracture appears to be retarded because part of the input is transferred into the matrix; the general shape of the longitudinal concentration profile is similar to that resulting from the process of longitudinal dispersion. If the contaminant also undergoes adsorption on the matrix as in Fig. 8 (c), diffusion into the matrix causes adsorption to occur over a much larger surface area than if all flow were entirely within the fracture, so advance of the front is delayed even more (see Maloszewski and Zuber 1985).

The effect of fracture-matrix interaction on contaminant transport in the field is shown schematically

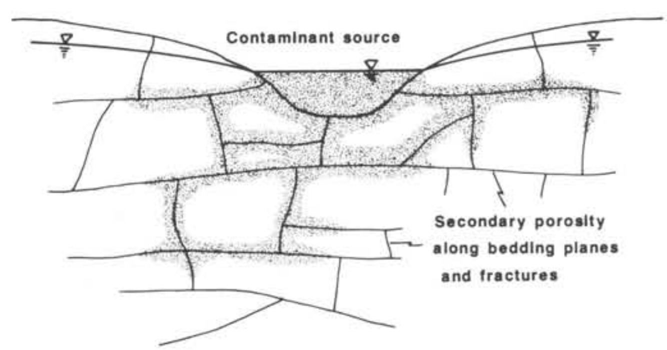

Figure 9. Contaminant migration through porous fractured media.

in Fig. 9. While the contaminant moves relatively rapidly through the fracture system, it will also be diffusing into the matrix as shown by the shaded areas. So long as the contaminant source remains, contamination will continue to propagate rapidly through the fracture network and diffuse slowly into the matrix. If the contaminant source is stopped, the fracture system will be rapidly flushed of its major contamination, but that contaminant previously diffused into the matrix will now diffuse slowly back into the fractures, resulting in a long-term low-level contamination of ground water flow through the fracture system. Some of the recently developed models discussed later can be applied to this complex situation.

Region of Flow

Characterization of flow system geometry can employ the tools commonly used in aquifer exploration, surface and subsurface geological and geophysical techniques.

Surface Geological Techniques. As in any research, initial steps of flow region characterization should involve a "literature review." There are numerous types of topographic, soils, surficial geology, and hydrogeologic maps which are published for most parts of the country by Federal agencies such as the Geologic Survey, Bureau of Reclamation, Soil Conservation Service, and Agricultural Research Service, or their state equivalents. These maps, along with air photo interpretations, may yield information on generalized bedrock formations and stratigraphy, distribution and origin of surficial deposits, water resource and geochemical interpretations, and hydraulic properties in the area of concern. All these factors can be used to develop initial indications of flow system geometry. These same information sources can also be used to indicate the extent and location of bedrock fracturing (see Way 1973). In carbonate terrain in Pennsylvania, Lattman and Parizek (1964) and Parizek and Drew (1966) demonstrated that well yields could be substantially increased by siting wells on fracture traces (major fracture concentrations) located by air photo interpretation. Jammallo (1984) demonstrated similar results in fractured crystalline bedrock in Vermont.

Subsurface Geological Techniques. The surficially oriented investigations discussed previously will usually have to be augmented by investigations oriented toward the subsurface. Again, existing information should first be sought out. Many state and local governmental agencies are now requiring that geologic logs of all wells drilled be maintained in centralized files. These field-based "observations" can be used to augment the interpretations derived from the surface geological investigations. In conjunction with the surface investigations, they can also guide the installation of test drilling or coring to expand the information base. In concert, these information sources can provide detailed local stratigraphy and geologic structure of both the overburden and bedrock. The test holes may also provide access to the aquifer for water level observation, aquifer testing, and water quality sampling (all discussed in later sections). Cores taken can be analyzed for hydraulic properties in addition to their providing lithologic information.

Further, well logs used in conjunction with test drilling and coring are imperative to investigations of fractured media, because this is one of the few ways to actually "see" the fracture system governing water movement and resulting chemical transport. Other techniques discussed later may characterize the bulk hydraulic properties of the aquifer, but subsurface geologic investigations are necessary to determine fracture frequency, sizes, orientation, and fracture zone boundaries, all of which have significant impact on interpretation and application of results from other tests. Information deduced from well logs and cores also provide point measurements helpful in characterizing areal distributions of aquifer properties (e.g., Urban and Gburek 1986).

Surface Geophysical Techniques. Surface geophysical techniques generally allow deduction of subsurface configuration from measurements made at the land surface (e.g., Dobrin 1960; Lennox and Carlson 1967). Three techniques which may be of some help in characterizing flow system geometry and the extent and location of fracturing are: seismic refraction, electrical resistivity, and ground penetrating radar.

Seismic refraction exploration is most reliable where different layers exhibit very different seismic velocities, and the layering of the system is limited and simple. Seismic investigations are commonly used to delineate depth to bedrock, areal aquifer extent, and presence of buried channels (e.g., Tibbetts et al. 1966; Watkins and Spieker 1971), but they have also been used to detect the depth of shallow fracture zones in sedimentary and crystalline rock. Using conventional seismic techniques, Urban and Gburek (1988) were successful in defining the depth of fracturing (2-15 m) in fractured shales and sandstones of east-central Pennsylvania, while Cybriwski et al. (1984) used vertical seismic profiling to define fracture zones and estimate their hydraulic conductivity in crystalline bedrock in northeastern Vermont.

Electrical resistivity interpretations are made by comparing apparent resistivities measured in the field to theoretical values for simple idealized systems (see USGS 1980). This technique is commonly used to determine thickness of sand and gravel aquifers, locate salt water-fresh water interfaces, and determine the extent of contaminant plumes where the contaminant has an electrical resistivity different than water. But it may also have potential to characterize fractured media. Huntley and Mishler (1984) report some success in relating hydraulic conductivity to resistivity in fractured crystalline rock in southern California. Taylor (1984) and Leonard-Mayer (1984) report measurement of vertical joint orientations and fracture porosity using azimuthal resistance techniques applied to carbonate bedrock in Wisconsin and Minnesota. Tselentis (1986) has combined conventional resistivity techniques with on-site microcomputer processing to investigate rock discontinuities.

Ground penetrating radar is a relatively new technique which relies on differences in dielectric properties of subsurface material to provide maps of subsurface layering. There are a number of reports of its use in characterizing rock contacts, faults, bedding planes, and jointing patterns (e.g., Morey 1974; Shih and Dolittle 1984). However, Urban and Gburek (1988) had no success in delineating the depth of fracturing at the same site where they experienced success with seismic techniques. They attribute the lack of success to interference problems resulting from numerous bedding planes within the shales.

While all three techniques are potentially useful, they cannot be used by themselves. Before drilling, they may be used as guidelines for intelligent and objective placement of test holes. Subsequent to drilling and coring, they may be used to provide areal extension and interpretation of the results from single borehole findings.

Subsurface Geophysical Techniques. Subsurface geophysical methods involve well or test hole logging using electrical resistivity, radiation, calipers, temperature, flow meters, and cameras. Theory, use, and instrumentation associated with the various logging techniques are beyond the scope of this paper (see Keys 1968; USGS 1980), and interpretation of the logs is often qualitative. For instance, differences encountered in the electrical and radiation logging are related to properties of both the geologic materials and the fluid within them. Depending on the need for knowledge at an individual site, the investigator may consider use of any or all of these logging techniques in conjunction with other techniques presented previously and subsequently. Paillet and Keys (1984) and Jones et al. (1984) report on studies using borehole geophysics in conjunction with hydraulic testing to evaluate fractured crystalline bedrock in Manitoba and Arizona.

A relatively new and valuable development which may be used in conjunction with the variety of logging techniques or separately, is the miniaturized down-hole TV camera. Recently marketed cameras are available in diameters as small as 40.5 mm (<1.75 in.), and are useful for direct examination of bedrock within small test holes that are less expensive to drill. The visual record derived is especially valuable for determining lithology, the orientation and location of fractures, and interpreting the response of observation wells or piezometers.

Boundary Conditions

Based on the aquifer geometry derived and the problem being analyzed, boundary conditions around the region of flow must be described to allow for solution of the equations of flow and transport within the region. The common boundary conditions used in analysis of subsurface flow systems are no-flow or specified-flux (Neumann-type boundaries) and constant head (Dirichlet-type boundaries). Sources or sinks within the flow region (e.g., recharge or pumping wells) may also be considered part of the boundary specifications for modeling efforts. Specific boundary conditions chosen are usually based on both determination of aquifer geometry and its interaction with the land surface and adjacent flow systems which would have been determined in earlier investigation. For instance, areas determined to be water table divides are usually specified as no-flow boundaries, positions where the aquifer discharges to the land surface or stream channel can be specified as constant head boundaries, and if it is determined that the aquifer receives or loses water to an adjacent unit, this boundary can be treated as a specified flux boundary. The flux rate imposed may be determined from hydraulic considerations (i.e., conductivities, gradients, etc.). Recharge to the ground water system, or pumpage from the system at various locations, is treated as a known source or sink at the node(s) affected.

Choosing appropriate boundary conditions is a very difficult part of ground water modeling. Where possible, physically based boundary conditions should be those employed. Yet, dominant and obvious natural boundaries controlling the flow system of concern may be at such distances that their inclusion in the model is difficult. A practical alternative is to make the overall region being modeled somewhat larger than the area of concern, thereby minimizing effects of the boundaries on the simulation. If this approach is taken, sensitivity analyses should be done to determine the effect of boundary choice on results of the simulation.

Flow System Testing

Theoretical or laboratory-based analyses of flow and contaminant transport through porous media may consider flow within individual pores or fractures, but from the practical perspective, consulting engineers and others working on field-scale problems usually resort to use of the continuum approach. Thus, lab or field testing procedures used to characterize media, either "classical" granular material or fractured bedrock, should be applied at a scale including sufficient pores or fractures to treat the media as a REV.

Hydraulic properties of porous media may be determined in the laboratory using samples extracted from the field, and the properties can also be determined in situ by testing of wells and/or piezometers. These same installations may later be used for both observing water levels for initial conditions and model verification, and for providing access to the ground water for water quality sampling. Following is a brief description of laboratory-based aquifer testing, well and piezometer testing, and benefits and drawbacks of each of these techniques.

Laboratory tests for hydraulic properties basically provide point determinations of aquifer properties. They are applied to small samples collected during drilling (disturbed samples) or coring (undisturbed samples). Generally, core-derived samples are preferred for laboratory analyses, but for granular material, disturbed samples may still be somewhat representative of field conditions. Hydraulic conductivity is determined using either constant-head or falling head permeameters, whereby water or another fluid (or gas) is forced through the sample under known flowrate and gradient. Application of Darcy's law yields hydraulic conductivity. Klute (1965) indicates that a constant-head test is better for media with conductivity greater than about 0.01 cm/min, while the falling head test is better for media having lower conductivities.

Porosity is generally determined by indirect techniques using measurements of bulk sample volume and saturated and unsaturated weight, or bulk volume and assumed particle density to infer void volume. In media with small pores, modifications of the first technique may be necessary to induce infiltration of water into all pores by vacuum, or through use of other liquids which infiltrate the available pore space more easily than does water. For hydraulic conductivity testing methodology see Black (1965), ASTM (1967), and Jaeger (1972), while Vomocil (1965) details porosity testing methodology. The variety of laboratory tests used for characterization of granular media are virtually useless in fractured systems because of scale. Further, testing results may be confusing because of the high degree of variability at the local scale (see Rogowski 1989).

Well testing determines in situ hydraulic parameter values, and because of the nature of the testing, the values are averaged and integrated over relatively large aquifer volumes, eliminating the problems associated with laboratory-based testing. However, pumping tests can be expensive to install and run, and their interpretations are somewhat subjective due to the influences of unknown boundary conditions encountered by relatively large cones of depression. Well testing for aquifer hydraulic parameter determination generally involves installation of a test well and observation well(s). The test well is pumped at a known flowrate, and by monitoring the resulting variation of water levels in the observation wells, the aquifer's transmissivity (T) and storativity (S) can be determined. Transmissivity is defined as the product of hydraulic conductivity and saturated thickness, while storativity is defined as the volume of water released from storage per unit surface area per unit decline of hydraulic head. Storativity and transmissivity have their origins in analysis of confined aquifers. For unconfined systems, the parameters of concern are hydraulic conductivity and specific yield, S_y, the volume of water that an unconfined aquifer releases from storage per unit decline in the water table. S_y represents drainage of pores, and is related to, but usually not identical to, porosity (see Urban and Gburek 1987). Interpretation of pumping test data has traditionally been done using graphical techniques, but recent innovations involve computer analysis of the data. The two most commonly used techniques for interpretation are the Theis (1935) time-drawdown analysis and the Jacob time-drawdown modified nonequilibrium technique (Cooper and Jacob 1946). Theory, design, analysis, and limitations of pumping tests can be found in Kruseman and deRidder (1970) and Stallman (1971), while Walton (1987) presents a number of analyses including computer techniques (and a PC diskette) for analysis of results.

By assuming that fractured rock behaves as a continuum, most of the standard hydraulic testing techniques can be applied to wells and piezometers situated in fractured media to evaluate its bulk hydraulic characteristics (see Urban and Gburek 1987). As in granular media testing, a pumping test in fractured media should be designed to sample the stratigraphy of concern. If the investigator has done borehole logging or coring, and finds the fracture system pervasive through the entire depth of the aquifer, the common time- or distance-drawdown techniques will yield hydraulic properties reflecting the overall length of well sampled. If however, the well is sampling a layered fractured system, interval testing by means of well packers may be required to differentiate hydraulic characteristics of the fractured and nonfractured layers. Depending on level of interpretation desired, corrections may be applied to the traditional tests to account for the nature of the fractured matrix (e.g., Smith and Vaughan 1985; Sen 1986), or cross-hole testing techniques may be used to evaluate the anisotropy of hydraulic conductivity in fractured systems (Hsieh and Neuman 1985; Hsieh et al. 1985).

Piezometer tests are less expensive to install and conduct and simpler to run than pumping tests. These tests are done in a single piezometer, an installation which is open to the aquifer over only limited depth, and sealed over the remainder of its length. Water level in the piezometer is not necessarily representative of the water table in the aquifer, rather, it represents hydraulic head at the position of the piezometer opening. Piezometer tests necessitate rapid introduction or removal of a known volume of water from the bore, and subsequent measurement of the rate of water level recovery. Piezometer tests may be successfully used in fractured systems, but concern must be given to the size and location of the piezometer tip in relation to the spacing and density of the fracture system. The Hvorslev (1951) method is used for analysis of a "point" piezometer (open over only a very small distance), while the method of Cooper et al. (1967) or Papadopoulos et al. (1973) is applicable to a piezometer open to the entire aquifer thickness. In both cases, the test results represent conditions relatively near the borehole. While application of the latter technique is relatively straightforward in fractured media, application of the point piezometer technique may be risky because the sampling point must be long enough to intersect sufficient fractures to sample an REV. Thus, piezometer design and testing should be integrated with coring and/or borehole logging.

Generally, pumping or piezometer tests will ultimately be employed to characterize an aquifer's hydraulic properties. The pumping test characterizes a larger volume of the aquifer's response to stress over a longer period of time, while the piezometer test characterizes a smaller volume of the aquifer responding to a more rapid stress. The space and time scales of the problem being investigated must be considered in choosing the appropriate testing technique (see Urban and Gburek 1987). Because of their smaller scale sampling, piezometer tests may give indications of spatial variations in aquifer hydraulic characteristics which are masked by pumping tests. At the same time, the random variability inherent in small-scale sampling of a natural system may make interpretation of results from piezometer tests difficult. The recently emphasized spatial variability analyses, such as those discussed by Rogowski (1989), may aid in these interpretations.

Mathematical Model of Solution

Many useful analytic and semianalytic solutions of the subsurface flow and transport equations can be found in the literature. However, these solutions are generally for simple geometry and boundary conditions. The primary tool used for analysis of subsurface flow systems under field conditions is the numerical model. Numerical models are based on discretization of the flow and transport equations, and of the continuum that is the subsurface flow system. The region of flow is broken down into a number of small "units," each assumed to be homogeneous with its own hydrogeologic characteristics. The manner in which the units are chosen depends both on the system being modeled (i.e., the degree of heterogeneity exhibited and that which is necessary to include in the solution), and the type of mathematical model used. Finite-difference models, one of the two major classes of models commonly used, require rectangular gridding, and the boundaries, either internal or external, are represented by a stair-step type of geometry when the region of flow is discretized. Finite-difference models are relatively easy to program and are computationally efficient.

The finite-element method, on the other hand, represents the region of flow as a series of small interconnected subregions, commonly triangles, quadrilaterals, or rectangles. These "finite elements" can be assembled in various ways to faithfully represent almost any internal geometry and external boundaries. Finite-element models are more difficult to program than are finite-difference models, and generally take longer to execute on the computer, but their solutions are generally considered more mathematically accurate. System discretization, although conceptually easier with finite elements, is more difficult to set up for computer entry. Choice of model is based on the importance of the variety of factors discussed above and the objectives of the simulation.

Two of the more commonly used and readily available finite-difference subsurface saturated flow models are PLASM (public domain; Prickett and Lonnquist 1971), and the USGS Modular Ground-Water Flow Model (public domain; McDonald and Harbaugh 1984). These are two-dimensional (areal) models which can be used in a quasi-three-dimensional mode to simulate layering. Potter and Gburek (1987a,b) have modified PLASM to simulate seepage face development. Two readily available finite-element models of saturated flow are FEWA (public domain; Yeh and Huff 1983) and SEFTRAN (proprietary, Huyakorn et al. 1984). PLASM and FEWA have associated contaminant transport models. A version of PLASM is available which contains a particle-tracking subroutine to simulate transport (convection, dispersion, and source/sinks) through the subsurface flow system being modeled (Prickett et al. 1981), while FEMA (Yeh and Huff 1985) uses FEWA geometry and flow system simulations as the base for a finite-element simulation of chemical transport (convection, dispersion, sorption, degradation, and radioactive decay). SEFTRAN contains chemical transport code (convection, dispersion, adsorption and first-order decay), but the output from the flow field simulation simply feeds into the transport simulation; the two solutions are not done concurrently.

A variety of two-dimensional (vertical cross-section) saturated/unsaturated models may also be considered for a more integrated treatment of the problem of ground water contamination. Two commonly used models of this type, both employing finite-element techniques, are UNSAT2 (Davis and Neumann 1983), and FEMWATER (Yeh and Ward 1980).

One problem of specific concern to the coastal areas of the NE, seawater intrusion, is not as amenable to solution by generic models as is the generalized problem of ground water contamination. Cooper et al. (1964) present a summary of analytic solutions applicable to analysis of seawater intrusion, while Pinder and Cooper (1970) present a numerical solution technique for position of the salt water front in a confined aquifer. Case studies of salt water intrusion can be found in Lee and Cheng (1974) and Segol and Pinder (1976).

A basic and practical introduction to ground water modeling can be found in Wang and Anderson (1982), while an overview of subsurface flow models available can be found in Javandal et al. (1984). Textbook treatments of subsurface flow modeling can be found in Remson et al. (1971), Pinder and Gray (1977), and Huyakorn and Pinder (1983). Huyakorn and Pinder (1983) also present the theory of numerical analysis of the transport equations, and Pinder (1973) and Bredehoeft and Pinder (1973) contain typical numerical solutions of these equations in field situations.

In general, models are already available to solve almost any subsurface flow problem envisioned, even contaminant transport within complex saturated/ unsaturated flow systems. The basic limitation in modeling subsurface flow systems continues to be an inability to describe both the system geometry and the internal parameters controlling flow. The results of system analysis using a numeric (or any kind of) model are no better than the modeler's perception of the system boundaries and parameters. However, if such models are used with supporting geologic and hydrologic investigations and the advice of geohydrologists and hydrologists, model results can be used successfully in analysis and management of NE ground water flow systems.

Flow System Monitoring and Sampling

Installation of ground water monitoring and sampling sites/networks for determination of initial conditions and model calibration or verification is perhaps the most subjective and poorly documented of all facets of ground water modeling. Such sites are relatively expensive compared to other hydrologic monitoring and sampling (e.g., raingages, streamgages). Additionally, knowledge of the subsurface system is much less detailed. Of concern here are those installations used for determining ground water levels and gradients, and for sampling of ground water for water quality analysis.

Monitoring wells and/or piezometers are installed for determination of water levels and energy gradients. Wells in unconfined aquifers indicate the position of the water table, while piezometers indicate hydraulic head at the depth of the piezometer opening. A network of observation wells can be used to determine the water table configuration over an area of concern. This can be used as initial condition information or for model calibration or verification. Piezometers installed in the vertical and/or horizontal can be used to determine gradients within the ground water body for determination of recharge and discharge zones as shown in Fig. 3. Water levels in both wells and piezometers can be read intermittently using a variety of devices, or can be continuously monitored using a variety of automated water level recorders (USGS 1980).

Detailed information is available regarding design and installation of wells or piezometers once a site has been determined (e.g., Johnson Division, UOP 1972; Barcelona 1983). However, siting of the monitoring/sampling installation or design of the network is quite subjective. Well networks intended to monitor the variability and areal distribution of water levels over a study area must be designed based on some a priori knowledge of the variabilities and areal distributions. For large-scale monitoring, enough information may be available from existing maps and water supply wells to get an initial indication of the variabilities and distributions. For small-scale problems however, less may be known a priori regarding site characteristics, and documentation of small-scale heterogeneities as they influence network design are important. For contaminant transport problems, those wells or piezometers used to monitor water levels and gradients may not be in the appropriate locations for withdrawal of water quality samples from the flow system. As an example, sampling wells/piezometers may need to be localized for both vertical and horizontal documentation of a contaminant plume, even though the flow system through which the plume is moving is much larger in scale and requires monitoring wells to be installed at the larger scale. Further, wells constructed for water quality sampling may have to be constructed of specialized materials based on the contaminant being sampled for (see Barcelona et al. 1983). An additional complication of water quality sampling involves withdrawal of a representative sample from the well bore. Suggestions range from removing a given number of "well volumes" of water from the bore before extracting the sample, to continuously monitoring concentration of selected quality constituents (e.g., dissolved oxygen, pH) as the well is being pumped and withdrawing the sample when these concentrations stabilize (see Pionke and Urban 1987). As in design of a sampling network, guidelines are vague and choice of sampling procedure is highly subjective. The network and procedure should be designed based on the problem and area of concern as best understood.

In some cases, crude simulations can be run using the more simplistic of the models discussed earlier and geometry and parameter values determined from the "literature." The results may then be used to indicate the critical areas which deserve sampling or monitoring installations. In an iterative type of procedure then, the investigator can go back and forth between simulation, sampling, resimulation incorporating the sampling results, etc., until the monitoring/sampling network is finalized to meet the objectives of the project.

Design of well and/or piezometer monitoring and sampling sites and networks must be based on knowledge of the geology and flow system in the area of concern, and a well-analyzed and formulated objective of the sampling and simulation (see Urban and Gburek 1988). In this overall area of modeling and sampling, the services of a hydrogeologist are invaluable. This author's recommendations are that no such study should be undertaken without this support.

A CASE STUDY

At the Northeast Watershed Research Center, efforts are underway to characterize shallow (<15 m) and deep (15-100 m) fracture layer controls on subsurface flow and contaminant transport within a 7.2-km^2 east-central Pennsylvania watershed. Topography and geology of the watershed are typical of upland watersheds in the unglaciated, intensely folded and faulted Appalachian Ridge and Valley Province. The watershed is underlain by interbedded shales, siltstones and sandstones. Detailed descriptions of the study area can be found in Urban (1977).

Based on previous findings within the watershed and other studies within the Appalachian region (Urban 1965; Ferguson 1967; Wyrick and Borchers 1981; Gerhart 1984), the shallow weathered fracture zone appears to have the potential to transmit large quantities of subsurface flow. This zone is normally fully drained in the uplands, where the potential for leakage to deeper ground water from the weathered zone exceeds root zone percolate. However, ground water return flow sustains near-surface saturation in the low-elevation areas of the watershed. Here, leakage to the deeper aquifer is near zero or negative (i.e., upward flow), so that lateral saturated flow is permanent within the shallow weathered fracture layer. Thus, methods available to characterize the layer's hydraulic properties in this zone are those of conventional well and piezometer tests and tracer observations.

Two cross-sections within a ground water discharge zone have been selected for detailed study of the physical and hydrologic controls imposed by the shallow weathered fracture layer at the small scale. Each section extends across a small stream valley.

Rock Cores

Rock cores were taken from each cross-section, one directly in the stream channel to a depth of 30 m, and two

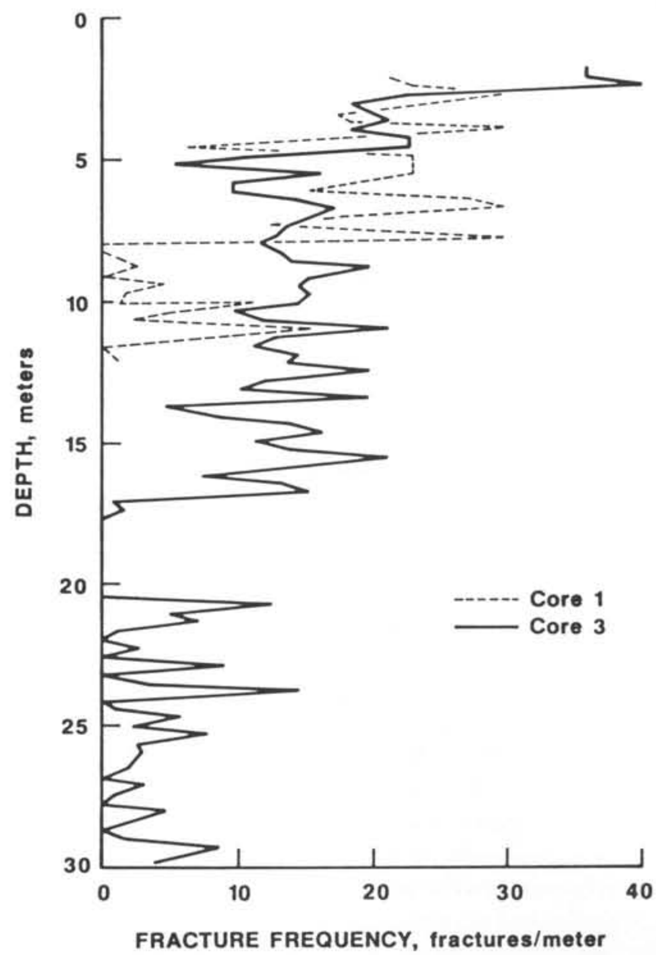

Figure 10. Fracture frequency with depth for two cores.

on either side up to lateral distances of 75 m and depths of 20 m. Fracture frequency for two cores from the east cross-section is shown in Fig. 10. Locations of these cores, along with a qualitative indication of fracture frequency based on visual interpretation for all cores in the section, are shown in Fig. 11. The bedrock is seen to be more highly fractured in the upper 10 to 15 m than at lower depths across the entire section, and depth of fracturing appears to be greatest directly under the channel. Wyrick and Borchers (1981) report this same pattern within the Appalachian Plateau Physiographic Province of West Virginia.

The weathered fracture layer appears as a dramatic visual feature of the cores; brown, red, and light-colored rock materials and fragments occur in the upper section of the cores, whereas the zone below contains the dark blues and grays of unweathered less-fractured parent material (clay shale). The contact between the two colors is distinct and consistent from core to core. Fracture width in the weathered zone appears to range from a fraction of

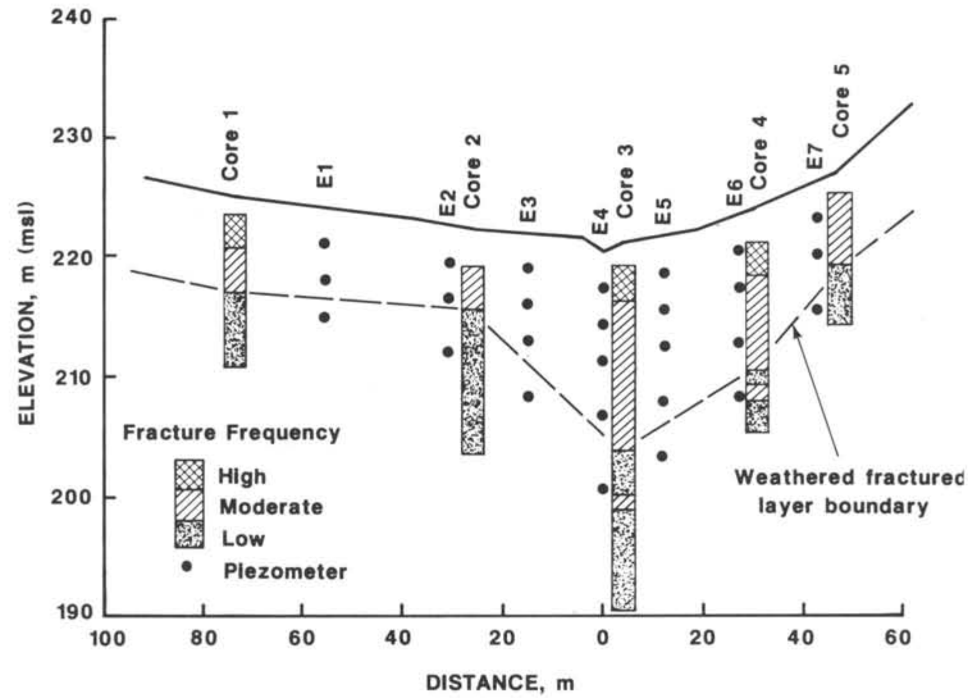

Figure 11. Fracture distribution and piezometer locations in east section.

a millimeter to about 10 millimeters, while widths in the unweathered zone appear too small to be measurable. Occasionally, there are single or groups of isolated fractures within the unweathered zone which appear to be weathered and are of the same dimensions as those in the weathered zone.

Matrix characteristics of the fractured bedrock within these cross-sections were evaluated by the Mineral Sciences Laboratory of the Pennsylvania State University using samples from the cores. Interconnected matrix hydraulic conductivity was too small to be measured by the equipment at the laboratory, and effective matrix porosity was extremely low in all cases. The shale bedrock matrix does not appear to constitute a significant transport media for ground water, nor does it have great potential to store water, so within the shallow weathered fracture layer, flow and chemical transport is governed by fracture geometry.

Geophysical Methods

Ground penetrating radar (GPR) and seismic refraction surveys were used in attempts to characterize the soil, weathered bedrock, and unweathered bedrock layers. GPR was totally ineffective in determining characteristics of the weathered zone near the cross-section study areas. However, the seismic investigation showed there to be a significant difference between the seismic signatures of the weathered fractured and the unweathered less-fractured layers. Computed depths to the weathered zone-unweathered zone contact based on seismic findings compared favorably with weathered zone depths observed in the rock cores, so the field-determined seismic signatures for the soil, weathered bedrock, and unweathered bedrock zones appear to provide a useful tool for mapping hydrogeologic units within the valley study area.

Cross-section Piezometers

Each cross-section has been instrumented with a network of piezometers to determine hydraulic head distribution within and immediately below the fracture layer, and to provide access for hydraulic testing and water quality sampling. Each piezometer point was completed as an unscreened, 61-cm long, 10.2-cm dia. hole. Based on fracture frequencies found during coring, this size samples the fracture-field at a scale which represents the fracture-controlled properties of the bedrock.

The piezometer net installed in the east section is shown in Fig. 11; piezometer lines made up of three to five piezometers at different depths are identified. Generally, the piezometers furthest from the stream indicate recharge conditions in the shallow zone, i.e., head decreases with depth. Those closer to the stream, where the effects of flowline convergence are more pronounced, indicate ground water discharge conditions; heads increase with depth. Figure 12 shows head distribution

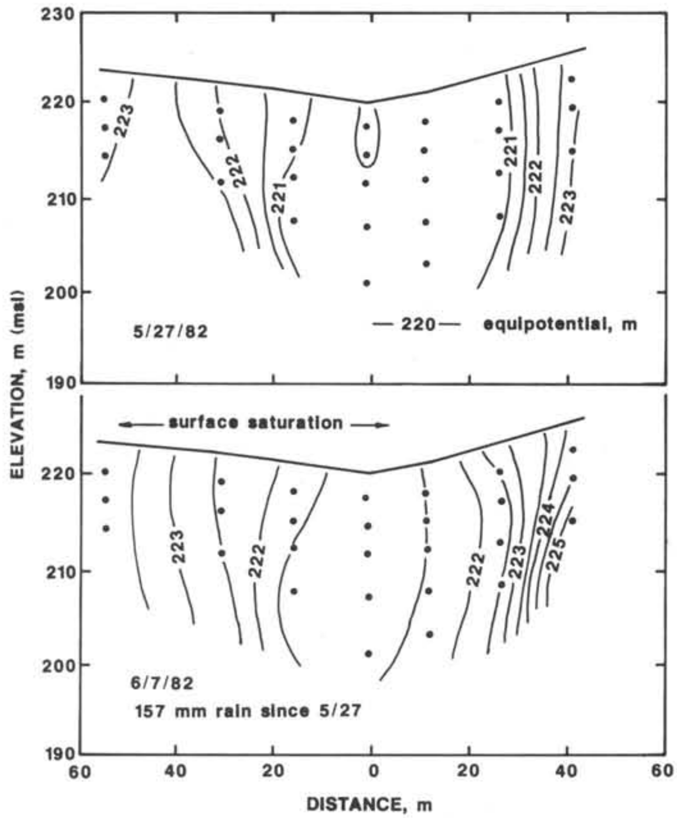

Figure 12. Typical head distributions within east section.

within the cross-section under differing hydrologic regimes. The 5/27/82 head distribution was during a relatively dry condition, while the 6/7/82 distribution was after 157 mm of rain over 11 days. The shallow fracture layer exhibited basically lateral flow (normal to the equipotentials) under the dry conditions, but the flow lines were bent toward the stream at depth. This implies upward flow from the deeper less-fractured zone to the shallow fracture layer which was acting as a drain. After the series of storms, the equipotential lines had migrated toward the channel, and there was more of a lateral flow pattern at all depths, with increasing gradients near the channel discharge point. In each case, lateral flow dominated to within approximately 15 m of the channel, where the flowlines converged to the stream channel. After the 157 mm of rain, hydraulic heads in all piezometers on the north (left) side of the section were greater than the land surface, resulting in the formation of an extensive seepage face discharging ground water directly to the land surface.

Hydraulic Conductivity Distribution

A total of 30 slug tests were performed in the section's piezometers, and results were analyzed using the Hvorslev (1951) technique. The tests yielded a mean hydraulic conductivity of 25×10^{-4} cm/sec for the shallow fractured weathered zone, but a standard deviation of 25×10^{-4} cm/sec indicates a wide spread within the values. Unfractured-zone tests show a mean hydraulic conductivity of 8×10^{-4} cm/sec with a 5×10^{-4} cm/sec standard deviation. Thus, the fractured zone averages about three times more permeable than the unfractured zone when both are considered as units.

Interaction of Land Use and Geology on Ground Water Quality

The land use-geology-flow system interactions exert a direct effect on shallow ground water quality at this small scale. Results of limited water quality sampling within the sections are given schematically in Fig. 13, and indicate a direct effect of overlying land use on ground water quality within the shallow fractured zone. High nitrate concentrations are observed under cropland, while under meadow, concentrations are almost zero. In the vicinity of the stream, where the two shallow lateral flows converge, the flow field also influences the nitrate concentrations. Here, concentrations being input to the stream are

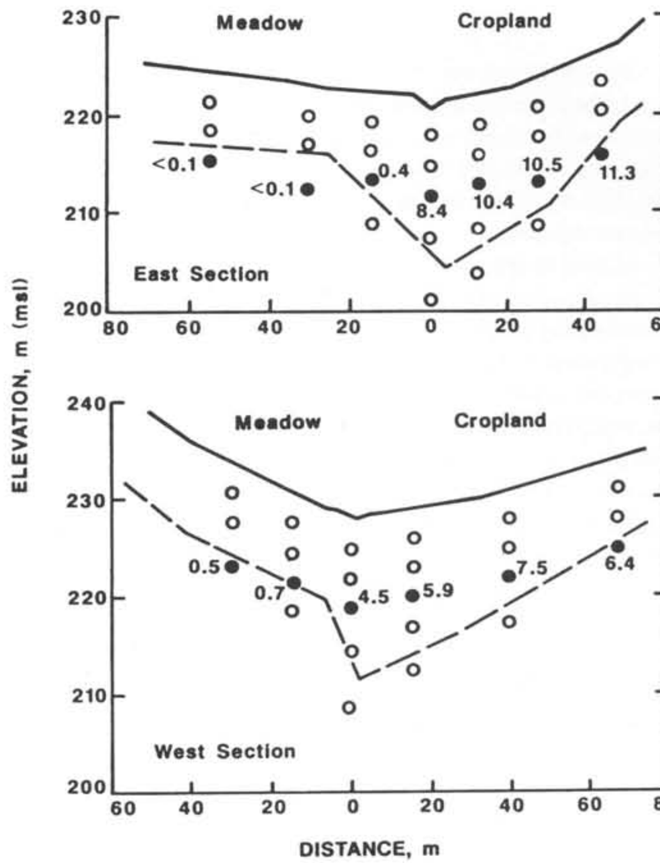

Figure 13. Nitrate concentrations (mg/l NO_3-N) in 9.1 m piezometers.

intermediate to those of the two fracture zone-derived lateral flows as a result of their mixing.

Implications for Characterizing Northeastern Ground Water Systems

The open character of the shallow, weathered, fracture zone in the study area marks a horizon of high hydraulic conductivity and a water transfer zone of major importance within upland watersheds. The layer acts as a direct conduit to the stream for ground water discharging upward from the regional aquifer and that flowing laterally within the shallow fracture zone itself. Water quality within this zone may be affected directly by overlying land use, and indirectly by the quality of the regional ground water flow; thus, subsurface flow input to the stream is a composite of both these sources from both sides of the stream.

To properly design models and ground water monitoring and sampling networks for water quantity and quality simulations of such a system, knowledge of the geologic controls on water movement is required. Further, where all components of the hydrologic cycle are intimately connected, as in the NE, modeling and sampling system design must also be based on the generalized ground water flow system. The analyses must consider subsurface watershed boundaries, effective aquifer depths, hydraulic conductivity and porosity distributions, and locations of the dominant ground water units of recharge, discharge, and lateral flow. Variabilities within aquifer parameters controlling the regional hydrogeologic relationships appear to be at the 10's of square-kilometers scale - 1000's of meters distance, and 10's of meters depth. Characterization of flow and contaminant transport within the regional system requires sampling at this scale. Important subcomponents of geology and hydrogeology though, such as the shallow, intensively fractured, weathered rock layer, may exert controls at a smaller scale. Variabilities within these systems must be examined at the scale of 100's of square meters - distances of 10's of meters and depths of a few meters.

COMPLICATIONS AND CONSTRAINTS

The Northeast was originally blessed with abundant and clean surface water. Man's impact resulted in deterioration of the quality of this resource, and the drought of the mid-60's also pointed out the quantity problems associated with surface-derived water supplies, even in precipitation-rich areas like the NE. Ground water is the alternative, and this fact, coupled with the increased emphasis on land disposal of wastes, makes knowledge and management of the ground water resource critical in the NE. However, legal and political aspects pertaining to the subsurface flow system can hinder our ability to comprehensively develop and manage the ground water resource in conjunction with other resources, such as surface water. Following are brief discussions of some of these aspects.

Generally, the riparian doctrine of water use had been recognized in the NE, providing for unrestricted use of ground water by each landowner. However, as continuing legal tests relating to this doctrine have been made, the trend has been toward the American modification of this English rule, i.e., toward a doctrine of reasonable use. This doctrine restricts the water rights of the landowner in relation to the needs of other landowners affected. The current transitory stage in development of ground water law tends to impede a rational approach to management of the ground water resource.

Many contend that a well-managed conjunctive-use program is the most appropriate for conditions such as that encountered in the NE. Conjunctive use involves balancing the use of ground water and surface water reservoirs based on both availability and demand. When excess surface water exists, the surplus may be used to recharge the ground water reservoir, while during low-flow periods, ground water is pumped to augment surface supplies. This concept has traditionally been understood to apply to the problem of water supply, but there is no reason why the concept can't and shouldn't be extended to encompass the problem of waste management and disposal also. However, thè political structure of the NE is not necessarily conducive to this. For instance, in Pennsylvania local government is the seat of power, and within the local governmental bodies, control of water supply and waste disposal is commonly vested within different agencies, departments, or authorities. This dispersal of control tends to discourage large-scale resource management.

Even the current emphasis on state-level management and protection of the ground water resource may not be sufficient to respond to "subsurface" problems of the NE. While aquifers themselves may be a "state-scale" problem, the fact that these aquifers provide a high percentage of flow to streams expands the effects of an individual aquifer to a much larger scale. For instance, nutrient input to Chesapeake Bay is currently a concern. While Pennsylvania is thought to contribute a large portion of the nutrients, especially nitrate, through the subsurface route, other states are also contributing via the same route. Regulation of ground water quality at the state level may not be sufficient when the large-scale subsurface flow systems combine to produce a very-large-scale surface water system transcending state boundaries.

Northeastern views of the subsurface flow systems are, in some cases, outdated. Water witching (dousing) is still encountered, sometimes related to relatively large-scale ground water development. Conversely, conjunctive-use and management of the ground water resource must consider innovative approaches to ground water recharge as an integrated part of the overall plan. In the urban Northeast, this will require acceptance and application of techniques such as use of porous asphalt (see Urban and Gburek 1981, and Gburek and Urban 1981). There is an obvious dichotomy between some perceptions and other needs.

Any approach to ground water management in the NE must include consideration of its effects on the near-stream environment and wetlands. Wetlands of the NE are generally zones of ground water discharge and, to varying degrees, zones of nutrient removal. Ground water contamination and/or management will thus have a direct effect on such wetlands and their outflows in context of the Northeast's interconnected flow systems. The riparian zone can also serve as a site for water quality management. Like wetlands, it provides ideal conditions for denitrification (reduction) of higher nitrate levels in the ground water inflow before they impact the stream. Such connections and controls must be quantified.

Finally, agricultural land use impacts on ground water in the NE must generally be analyzed in context of a watershed system approach. Agricultural impacts are "nonpoint" in nature, and their effects are felt at scales other than local and in more than one flow component of the subsurface system. When the typical NE watershed scale is considered however, the system being dealt with is one of mixed land use, typically consisting of intermixed urban, suburban, crop, and forest land uses contributing to the same subsurface flow system. Further, as detailed earlier, these effects are felt not only in the ground water at the local scale, but they are also felt in the streams draining the upland watershed of concern, and in their aggregation with other such streams draining other upland watersheds to form a larger scale river system.

SUMMARY AND CONCLUSIONS

The humid climate of the Northeast assures us of continuing abundant supplies of water. The nature of the shallow ground water flow systems insures that they will be "refilled" by this precipitation annually. The most important problem relating to ground water at present is past and potential contamination by both point and nonpoint sources of pollution. Because of the nature of the NE region, with agriculture and industry coupled with a heavy population density, management of ground water quality is especially important. Development of the capability to characterize, model, and sample NE ground water flow systems for both water and chemical movement is critical if we are to make progress in protecting and preserving this valuable natural resource.

LITERATURE CITED

ASTM. 1967. Permeability and Capillarity of Soils. Amer. Soc. of Testing Materials, Philadelphia, PA, 210 pp.

Barcelona, M.J., J.P. Gibb and R.A. Miller. 1983. A guide to the selection of materials for monitoring well construction and ground-water sampling. Contract Report 327, Illinois State Water Survey, Urbana, IL, 78 pp.

Bear, J. 1972. Dynamics of Fluids in Porous Media. American Elsevier, New York, NY.

Black, C.A., ed. 1965. Methods of Soil Analysis, Part 1. Amer. Soc. of Agronomy, Madison, WI, 770 pp.

Bredehoeft, J.D. and G.F. Pinder. 1973. Mass transport in flowing groundwater. Water Resour. Res. 9:194-209.

Burkholder, H.C. 1976. Methods and data for predicting nuclide migration in geologic media. Intern. Symp. on Management of Wastes from the LWR Fuel Cycle, Denver, CO.

Caswell, W.B. and S. Ludwig. 1978. Maine coastal area water supply and demand. Maine Geological Survey and State Planning Office, 244 pp.

Claassen, H.C. and E.H. Cordes. 1975. Two-well recirculating tracer test in fractured carbonate rock, Nevada. Hydro. Sciences Bull., v. XX, pp. 367-382.

Cline, G.D. 1968. Geologic factors influencing well yields in a folded sandstone-siltstone-shale terrain within the East Mahantango Creek Watershed. M.S. Thesis, Pennsylvania State University.

Cooper, H.H., Jr. and C.E. Jacob. 1946. A generalized graphical method for evaluating formation constants and summarizing well field history. Trans. Amer. Geophys. Union 27:526-534.

Cooper, H.H., J.D. Bredehoeft and I.S. Papadopulos. 1967. Response of a finite-diameter well to an instantaneous charge of water. Water Resour. Res. 3:263-269.

Cooper, H.H. Jr., F.A. Kohout, H.R. Henry and R.E. Glover. 1964. Sea water in coastal aquifers. U.S. Geol. Survey Water Supply Paper 1613-C, 84 pp.

Cybriwski, Z.A., E.N. Levine and M.N. Taksoz. 1984. Detection of permeable rock fracture zones within crystalline bedrock by 3D vertical seismic profiling. Proc. Eastern Reg. Ground Water Conf., Nat. Water Well Assn., pp. 274-293.

Davis, L.A. and S.P. Neuman. 1983. Documentation and User's Guide: UNSAT2 Variably Saturated Flow Model (Including 4 Example Problems). Final Report, NUREG/CR-3390 WWL/TM-1791-1, Water, Waste and Lane, Inc., Fort Collins, CO.

DeAngelis, D.L., G.T. Yeh and D.D. Huff. 1984. An integrated compartmental model for describing the transport of solute in a fractured porous media. Environ. Sci. Div. Publication No. 2292, Oak Ridge National Laboratory, Oak Ridge, TN, 141 pp.

Dobrin, M.B. 1960. Introduction to Geophysical Prospecting. McGraw-Hill, New York, NY, 446 pp.

Emsellem, Y. and G. deMarsily. 1971. An automatic solution for the inverse problem. Water Resour. Res. 7:1264-1283.

EPA. 1985. Overview of State Ground-Water Program Summaries: Volume 1. Office of Ground-Water Protection, U.S. Environmental Protection Agency, Washington, DC., 28 pp.

Ferguson, H.F. 1967. Valley stress relief in the Allegheny Plateau. Assn. of Engr. Geol. Bull. 4:63-68.

Freeze, R.A. and J.A. Cherry. 1979. Groundwater. Prentice-Hall Inc., Englewood Cliffs, NJ, 604 pp.

Gburek, W.J. and J.B. Urban. 1980. Storm water detention and groundwater recharge using porous asphalt - initial results. Int. Symp. on Storm Runoff, July 28-31, 1980, Lexington, KY, pp. 89-97.

Gerhart, J.M. 1984. A model of regional ground-water flow in secondary-permeability terrain. Ground Water 22:168-175.

Grisak, G.E. and J.F. Pickens. 1980. Solute transport through fractured media: 1. The effect of matrix diffusion. Water Resour. Res. 16:719-730.

Grisak, G.E., J.F. Pickens and J.A. Cherry. 1980. Solute transport through fractured media: 2. Column study of fractured till. Water Resour. Res. 16:731-739.

Handman, E.H., I.G. Grossman, J.W. Bingham and J.L. Rolston. 1979. Major sources of ground-water contamination in Connecticut. U.S. Geol. Survey Water Resources Investigations Report 79-1069, 59 pp.

Heath, R.C. 1984. Ground-Water Regions of the United States. U.S. Geol. Survey Water Supply Paper 2242, 78 pp.

Hickey, J.J. 1984. Field testing the hypothesis of Darcian flow through a carbonate aquifer. Ground Water 22:544-547.

Hobba, W.A. 1985. Water in Hardy, Hampshire, and western Morgan counties, West Virginia. West Virginia Geol. and Econ. Survey Environ. Geology Bull. EGB-17.

Hsieh, P.A. and S.P. Neuman. 1985. Field determination of the three-dimensional hydraulic conductivity tensor of anisotropic media: 1. Theory. Water Resour. Res. 21:1655-1665.

Hsieh, P.A., S.P. Neuman, G.K. Stiles and E.S. Simpson. 1985. Field determination of the three-dimensional hydraulic conductivity tensor of anisotropic media: 2. Methodology and application to fractured rocks. Water Resour. Res. 21:1667-1676.

Hubbert, M.K. 1940. The theory of groundwater motion. J. Geol. 48:785-944.

Huntley, D. and H.M. Mishler. 1984. Relationship between permeability and electrical resistivity in granular and fractured rock aquifers. Surface and Borehole Geophys. Methods in Ground Water Invest., Nat. Water Well Assn., pp. 18-36.

Huyakorn, P.S., R.W. Broome, A.G. Kretschek and J.W. Mercer. 1984. SEFTRAN: A Simple and Efficient Flow and Transport Code. Technical Report prepared for Holcomb Research Institute International Ground Water Modeling Center by Geotrans, Inc., Herndon, VA, 142 pp.

Huyakorn, P.S. and G.F. Pinder. 1983. Computational Methods in Subsurface Flow. Academic Press, New York, NY, 473 pp.

Huyakorn, P.S., B.H. Lester and C.R. Foust. 1983. Finite element techniques for modeling groundwater flow in fractured aquifers. Water Resour. Res. 19:1019-1035.

Huyakorn, P.S., H.W. White, Jr., V. Guvanasen and B.H. Lester. 1986. TRAFRAP: A two-dimensional finite element code for simulating fluid flow and transport of radionuclides in fractured porous media. Geotrans, Inc., Reston, VA.

Hvorslev, M.J. 1951. Time lag and soil permeability in ground-water observations. Bulletin 36, Waterways Experiment Station, Corps of Engr., Vicksburg, MI.

Jaeger, J.C. 1972. Rock Mechanics and Engineering. Cambridge University Press, London, 523 pp.

Jammallo, J.M. 1984. Use of magnetics to enhance identification of bedrock fracture trace zones for well locations. Proc. Eastern Reg. Ground Water Conf., Nat. Water Well Assn., pp. 105-133.

Javandal, I., C. Doughty and C.F. Tsang. 1984. Groundwater transport: Handbook of mathematical models. Water Res. Monograph 10, Amer. Geophys. Union, 228 pp.

Johnson, R.H. 1973. Hydrology of the Columbia (Pleistocene) deposits of Delaware. Delaware Geological Survey Bulletin 14, 78 pp.

Johnson Division, UOP. 1972. Ground Water and Wells. Johnson Division, Universal Oil Products Co., Saint Paul, MI, 440 pp.

Jones, J.W., E.S. Simpson and S.P. Neuman. 1984. Geophysical investigation of fractured crystalline rock near Oracle, Arizona. Surface and Borehole Geophys. Methods in Ground Water Invest., Nat. Water Well Assn., pp. 877-888.

Keys, W.S. 1968. Logging in groundwater hydrology. Ground Water 6:10-18.

Klute, A. 1965. Laboratory measurement of hydraulic conductivity of saturated soils. Methods of Soil Analysis, Part 1, C.A. Black. (ed.) Amer. Soc. of Agronomy, Madison, WI, pp. 210-221.

Kruseman, G.P. and N.A. deRidder. 1970. Analysis and evaluation of pumping test data. Intern. Inst. Land Reclamation and Improvement, Bulletin 11, Wageningen, The Netherlands.

Krynine, D.P. and W.R. Judd. 1957. Principles of Engineering Geology and Geotechnics. McGraw Hill Book Co., Inc., New York, NY, 730 pp.

Lattman, L.A. and R.R. Parizek. 1964. Relationship between fracture traces and the occurrence of ground water in carbonate rock. J. Hydrol. 2:73-91.

Lee, C.H. and T.S. Cheng. 1974. On seawater encroachment in coastal aquifers. Water Resour. Res. 10:1039-1043.

Leonard-Mayer, P.J. 1984. Development and use of azimuthal resistivity surveys for jointed formations. Surface and Borehole Geophys. Methods in Ground Water Invest., Nat. Water Well Assn., pp. 52-91.

Maloszewski, P. and A. Zuber. 1985. On the theory of tracer experiments in fissured rocks with a porous matrix. J. Hydrol. 79:333-358.

McDonald, M.G. and A.W. Harbaugh. 1984. A modular three-dimensional finite-difference ground-water flow model. U.S. Geol. Survey Open-File Report 83-875, 528 pp.

Meinzer, O.E. 1923. The occurrence of ground water in the United States, with a discussion of principles. U.S. Geol. Survey Water Supply Paper 489, 78 pp.

Morey, R.M. 1974. Continuous subsurface profiling by impulse radar. Proc. Subsurface Explor. for Underground Excavation and Heavy Const., Amer. Soc. Civil Engr., pp. 213-232.

Neuman, S.P. 1973. Calibration of distributed parameter groundwater flow models viewed as a multiple-objective decision process under uncertainty. Water Resour. Res. 9:1006-1021.

Neuman, S.P. 1975. Role of subjective value judgement in parameter identification. Modeling and Simulation of Water Resources Systems, G.C. Vansteenkiste, ed., North-Holland, Amsterdam, pp. 59-82.

Ogata, A. 1970. Theory of dispersion in a granular medium. U.S. Geol. Surv. Prof. Paper 411-I.

Paillet, F.L. and W.S. Keys. 1984. Applications of borehole geophysics in characterizing the hydrology of fractured rock. Surface and Borehole Geophys. Methods in Ground Water Invest., Nat. Water Well Assn., pp. 743-761.

Papadopulos, S.S., J.D. Bredehoeft and H.H. Cooper. 1973. On the analysis of "slug test" data. Water Resour. Res. 9:1087-1089.

Parizek, R.R. and L.J. Drew. 1966. Random drilling for water in carbonate rocks. Proc. Symp. Computers Operations Res., Mineral Ind. Exp. Sta., Vol. 3, Pennsylvania State University, pp. 1-22.

Pinder, G.F. 1973. A Galerkin-finite element simulation of groundwater contamination on Long Island, NY. Water Resour. Res. 9:1657-1669.

Pinder, G.F. and H.H. Cooper Jr. 1970. A numerical technique for calculating the transient position of the saltwater front. Water Resour. Res. 6:875-882.

Pinder, G.F. and W.G. Gray. 1977. Finite Element Simulation in Surface and Subsurface Hydrology. Academic Press, New York, NY, 295 pp.

Pionke, H.B. and J.B. Urban. 1987. Sampling the chemistry of shallow aquifer systems - A case study. Ground Water Monitoring Rev. 7:79-88.

Potter, S.T. and W.J. Gburek. 1987a. Seepage face simulation using PLASM. Ground Water 25:722-732.

Potter, S.T. and W.J. Gburek. 1987b. Documentation and User's Guide: Seepage Face Modifications to PLASM. Northeast Watershed Research Center, USDA-ARS (in house), 79 pp.

Prickett, T.A. and C.G. Lonnquist. 1971. Selected digital computer techniques for groundwater resource evaluation. Bulletin 55, Illinois State Water Survey, Urbana, IL, 65 pp.

Prickett, T.A., T.G. Naymik and C.G. Lonnquist. 1981. A "random-walk" solute transport model for selected groundwater quality applications. Bulletin 65, Illinois State Water Survey, Urbana, IL, 103 pp.

Rasmuson, A. 1985. Analysis of hydrodynamic dispersion in discrete fracture networks using the method of moments. Water Resour. Res. 11:1677-1683.

Remson, I., G.M. Hornberger and F.J. Molz. 1971. Numerical Methods in Subsurface Hydrology. Wiley-Interscience, New York, NY, 389 pp.

Rogowski, A.S. 1989. Estimating the parameters controlling saturated and unsaturated flow in soils. C.R. Frink (ed). Ground Water in the Northeast. The Connecticut Agricultural Experiment Station, New Haven. Bulletin 876. 75pp.

Rolston, J.L., I.G. Grossman, R.S. Potterton Jr. and E.H. Handman. 1979. Places in Connecticut where ground water is known to have deteriorated in quality. U.S. Geol.

Survey Misc. Field Studies Map MF 981-G.

Ross, B. 1986. Dispersion in fractal fracture networks. Water Resour. Res. 22:823-827.

Sager, B., S. Yakowitz and L. Duckstein. 1975. A direct method for the identification of the parameters of dynamic nonhomogeneous aquifers. Water Resour. Res. 11:563-570.

Schwartz, F.W., L. Smith and A.S. Crowe. 1983. A analysis of macroscopic dispersion in fractured media. Water Resour. Res. 19:1253-1265.

Segol, G. and G.F. Pinder. 1976. Transient simulation of saltwater intrusion in southeastern Florida. Water Resour. Res. 12:65-70.

Sen, Z. 1986. Aquifer test analysis in fractured rocks with linear flow pattern. Ground Water 24:72-78.

Sharp, J.C., Y.N.T. Maini and T.R. Harper. 1972. Influence of groundwater on the stability of rock masses: 1. Hydraulics within rock masses. Trans. Inst. Min. and Met. Lond. 81:A13-A20.

Shih, S.F. and J.A. Doolittle. 1984. Using radar to investigate organic soil thickness in the Florida Everglades. Soil Sci. Soc. Am. J. 48:651-656.

Smith, E.D. and N.D. Vaughan. 1985. Aquifer test analysis in nonradial flow regimes: A case study. Ground Water 23:167-175.

Snipes, D.S., G.G. Padgett, W.B. Hughes and G.E. Springston. 1983. Ground water quantity and quality in fracture zones in Abbeville County, South Carolina. Technical Completion Report A-053-SC, Water Resources Research Institute, Clemson University, 54 pp.

Snow, D.T. 1968. Rock fracture spacings, openings, and porosities. J. Soil Mech. Found. Div., Proc. Amer. Soc. Civil Engrs. 94:73-91.

Snow, D.T. 1969. Anisotropic permeability of fractured media. Water Resourc. Res. 5:1273-1289.

Stallman, R.W. 1971. Aquifer-test design, observation, and data analysis. U.S. Geol. Survey Techniques of Water Resources Investigations, Book 3, Chapter B1, 26 pp.

Taylor, R.W. 1984. The determination of joint orientation and porosity from azimuthal resistivity measurements. Surface and Borehole Geophys. Methods in Ground Water Invest., Nat. Water Well Assn., pp. 37-51.

Theis, C.V. 1935. The relationship between the lowering of the piezometric surface and the rate and duration of discharge of a well using groundwater storage. Trans. Amer. Geophys. Union 2:519-524.

Thomas, H.E. 1951. The Conservation of Ground Water. McGraw-Hill, New York, 327 pp.

Tibbetts, B.L., C.R. Dunrud and F.W. Osterwald. 1966. Seismic refraction measurements at Sunnyside, Utah. U.S. Geol. Survey Prof. Paper 550D, pp. D132-D137.

Trescott, P.C. 1975. Documentation of finite-difference model for simulation of three-dimensional groundwater flow. U.S. Geol. Survey Open-File Report 75-438.

Tselentis, G. 1986. On-site assessment of rock discontinuities from resistivity logs. T-L log: A new logging technique. J. Hydrology, v. 83, pp. 269-283.

USGS. 1980. National Handbook of Recommended Methods for Water-Data Acquisition: Chapter 2-Ground Water. Office of Water Data Coordination, U.S. Geol. Survey.

USGS. 1984. National Water Summary 1984. U.S. Geol. Survey Water Supply Paper 2275, 467 pp.

Urban, J.B. 1965. Geologic and hydrologic significance of seeps and springs in eastern Ohio. J. Soil Cons. Soc. Am. 20:178-179.

Urban, J.B. 1977. The Mahantango Creek Watershed - Evaluating the shallow ground water regime. Watershed Research in Eastern North America, ed. D.L. Correll, Smithsonian Institute, Washington, DC, pp. 251-275.

Urban, J.B. and W.J. Gburek. 1980. Storm water detention and groundwater recharge using porous asphalt - experimental site, Inter. Symposium on Storm Runoff, July 28-31, 1980, Lexington, KY, pp. 81-87.

Urban, J.B. and W.J. Gburek. 1986. Determination of aquifer parameters at a ground water recharge site. Ground Water 26:39-53.

Urban, J.B. and W.J. Gburek. 1988. A geologic and flow-system-based rationale for ground-water sampling. Field Methods for Ground-Water Contamination Studies and Their Standardization, ASTM STP 963, A.G. Collins and A.I. Johnson (eds.) American Society for Testing Materials, Philadelphia, PA, pp. 468-481.

Vomocil, J.A. 1965. Porosity. Methods of Soil Analysis, Part 1., C.A. Black (ed.) Am. Soc. of Agronomy, Madison, WI, pp. 299-314.

Wagenet, R.J. and P.F. Germann. 1989. Concepts and models of water flow in macropore soils. C.R. Frink (ed). Ground Water in the Northeast. The Connecticut Agricultural Experiment Station, New Haven. Bulletin 876. 75pp.

Walton, W.C. 1987. Groundwater Pumping Tests: Design and Analysis. Lewis Publishers, Inc., Chelsea, MI, 201 pp.

Wang, H.F. and M.P. Anderson. 1982. Introduction to Groundwater Modeling. W.H. Freeman and Co., San Francisco, CA, 237 pp.

Watkins, J.S. and A.M. Spieker. 1971. Seismic refraction survey of pleistocene drainage channels in the Lower Great Miami River Valley, Ohio. U.S. Geol. Survey Prof. Paper 605B, pp. B1-B17.

Way, D.S. 1973. Terrain Analysis: A Guide to Site Selection Using Aerial Photographic Interpretation. Dowden, Hutchison and Ross, Stroudsburg, PA, 392 pp.

Witherspoon, P.A., C.H. Amick, J.E. Gale and K. Iwai. 1979. Observations of a potential size effect in experimental determination of the hydraulic properties of fractures. Water Resour. Res. 15:1142-1146.

Wittke, W. 1973. General report on the symposium "Percolation Through Fissured Rock." Bull. Intern. Eng. Geol., pp. 3-28.

Wyrick, G.G. and J.W. Borchers. 1981. Hydrologic effects of stress relief fracturing in an Appalachian valley. U.S. Geol. Survey Water Supply Paper 2117, 51 pp.

Yeh, G.T. and D.D. Huff. 1983. FEWA: A finite element model of water flow through aquifers. Environ. Sci. Div. Publication No. 2240, Oak Ridge National Laboratory, Oak Ridge, TN, 216 pp.

Yeh, G.T. and D.D. Huff. 1985. FEMA: A finite-element model of material transport through aquifers. Environ. Sci. Div. Publication No. 2428, Oak Ridge National Laboratory, Oak Ridge, TN, 80 pp.

Yeh, G.T. and D.S. Ward. 1980. FEMWATER: A Finite-Element Model of Water Flow Through Saturated-Unsaturated Porous Media. Env. Sci. Div. Publication No. 1370, Oak Ridge National Laboratory, Oak Ridge, TN, 153 pp.

LIST OF SYMBOLS

A - cross-sectional area of flow (L^2)

C - concentration (M/L^3)

D - dispersion coefficient (L/T)

h - hydraulic head (L)

i - hydraulic gradient (L/L)

K - hydraulic conductivity (L/T)

n - effective porosity (L^3/L^3)

Q - flow (L^3/T)

s_l and s_t - directions of ground water flow lines and normals to these lines respectively

S - storativity (L^3/L^3); volume of water released from storage by an aquifer per unit surface area per unit decline in hydraulic head, generally applied to confined aquifers

S_s - specific storage (l/L); volume of water released from storage by a unit volume of aquifer under a unit decline in hydraulic head

S_y - specific yield (L^3/L^3); volume of water released from storage by an unconfined aquifer per unit surface area per unit decline in the water table

t - time (T)

T - transmissivity (L^2/T); the product of hydraulic conductivity and saturated thickness, generally applied to confined aquifers

V - fluid velocity (L/T); the Darcy velocity or specific discharge

V_p - average linear fluid velocity within the pores (L/T)

x, y and z - coordinate directions (L)

The Connecticut Agricultural Experiment Station, founded in 1875, is the first experiment station in America. It is chartered by the General Assembly to make scientific inquiries and experiments regarding plants and their pests, insects, soil and water, and to perform analyses for State agencies. The laboratories of the Station are in New Haven and Windsor; its Lockwood Farm is in Hamden. Single copies of bulletins are available free upon request to Publications; Box 1106; New Haven, Connecticut 06504. ISSN 0097-0905