REPORT SUMMARY

SUBJECTS	Solid-waste studies / Solid by-product disposal/reuse / Water quality/aquatic resources / Water quality control
TOPICS	Dispersions; Groundwater; Transport; Hydrodynamics; Solid-waste disposal; Leachate migration
AUDIENCE	Environmental scientists and engineers / R&D scientists and planners

A Review of Field-Scale Physical Solute Transport Processes in Saturated and Unsaturated Porous Media

Hydrodynamic dispersion is the key physical process that determines how groundwater plumes of chemicals develop. This literature review defines the status of data on field-scale groundwater transport mechanisms and represents a first step in developing and validating methods for predicting the fate of leached chemicals.

BACKGROUND EPRI is conducting a multiphase effort to quantify migration patterns of chemicals released from solid wastes at utility disposal sites. Before researchers can develop accurate models for predicting the groundwater transport of solutes, reliable data on field-scale transport parameters are needed.

OBJECTIVES To conduct a critical review of literature reporting field-scale dispersion data and to analyze selected data for interpreting and identifying data gaps.

APPROACH Researchers manually searched the literature and identified numerous published and unpublished studies yielding field-scale values of dispersivity. The primary targets for this search were dispersion parameters because they control the degree of spreading and dilution of a solute plume. Both saturated and unsaturated subsurface environments were included. After compiling the data, the researchers summarized information on the longitudinal, horizontal, and vertical dispersivities; aquifer material; average aquifer thickness; hydraulic conductivity; apparent heterogeneity; and effective porosity. They then analyzed data from selected sites, interpreting the information and illustrating the use of alternative mathematical methods in the analysis.

RESULTS The literature review yielded the following information on field-scale dispersivities. For the saturated zone, researchers found 55 sites (studies) in which a total of 99 values of longitudinal dispersivity was reported. In contrast, 22 and 7 values of horizontal and vertical dispersivities, respectively, were reported. Only 2 sites provided data extensive enough to allow a reanalysis for illustrating the use of alternative methodologies and interpretations. In

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ABSTRACT

The development of accurate mathematical models to predict field-scale solute transport in the saturated and unsaturated zones is hampered by the lack of reliable data on field-scale transport parameters. A critical review of the available literature on studies conducted at 55 saturated zone and 28 unsaturated zone sites produced 99 and 8 longitudinal dispersivity values, respectively.

In the saturated zone, the scale of observation for all the data ranged from 0.75 m to 100 km with longitudinal dispersivities from 0.01 to 5500 m. However, only five sites produced highly reliable dispersivity data, based on an evaluation of the test configuration, the tracer monitoring, and the data analysis method for each site. The largest scale of high reliability dispersivities was only 115 m. The high reliability data subset indicates that the dispersivity initially increases with the scale of observation. But it is not clear whether the dispersivity increases indefinitely with scale or reaches an asymptotic value as is assumed in classical modelling and predicted by recent stochastic theories.

In the unsaturated zone the dispersivity ranged from 1 mm to 0.7 m and appeared to increase with the scale of observation from 1 m to 20 m; however, most experiments were at scales of about 2 m. The transport process is dominated by the lateral movement of solutes in dry, high tension soils whereas in nearly saturated soils the solutes and water can move rapidly downward through the the macrostructures.

There is a clear need to conduct controlled large-scale field experiments in both the saturated and unsaturated zones to obtain reliable dispersivities at increasing scales and to identify the controlling transport mechanisms.

ACKNOWLEDGMENTS

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CONTENTS

SUMMARY

The first part of the report examines solute transport processes in the saturated zone. The principal objectives of that section are to (1) review all available literature where values of field-scale dispersivity were reported; (2) perform detailed analyses of data from selected sites to illustrate alternative methods and data interpretation; (3) evaluate the data collected and make a judgement regarding its reliability; and (4) draw conclusions and provide recommendations relating to the design of field experiments.

From the literature search, 55 sites were found where a total of 99 values of longitudinal dispersivity were reported. These ranged from 0.01 m to more than 5500 m at scales of 0.75 m to 100 km. From a first look at the data, without consideration of the quality of the data, it would appear that dispersivity increases indefinitely with scale. It should be noted that there is a paucity of data in the literature on horizontal and vertical transverse dispersion. Only 22 and seven values of horizontal and vertical dispersivity, respectively, were reported. This is an important data gap in light of the fact that contaminant plumes are three-dimensional in nature.

In addition to dispersivity values, a number of parameters for each site were tabulated to determine whether there was a relationship between these parameters and dispersivity. Upon examination of the data, there do not appear to be any such relationships.

Two sites having extensive, good quality data were re-evaluated to illustrate alternative methodologies and interpretations. Data from the Borden site in Canada were used to illustrate the method of second moment analysis to estimate three-dimensional dispersion characteristics and to demonstrate dependence of the dispersivities on displacement distance. Data from the Bonnaud site in France were used to demonstate the application of stochastic theories for predicting macrodispersion in aquifers.

The data collected from the general review were critically evaluated to determine how well field-scale dispersivity was actually measured by the tests conducted.

spreading of moisture in stratified soils due to capillary forces. The second is fast vertical flow in fingers or channels due to gravity. The observations show that lateral movement is more pronounced in relatively dry soils with correspondingly high capillary tensions. This lateral spreading is in accordance with the results of the recent stochastic theory. Fast gravity flow down through macro-structures seems to be important only near saturation conditions.

A critique of conceptual modelling approaches to field-scale unsaturated flow and transport emphasized the effects of field heterogeneity and techniques for evaluating large-scale effective parameters.

In unsaturated media, field-scale longitudinal dispersivity data from eight different field sites show dispersivities ranging from 1 mm to 0.7 m for scales from 1 m to 20 m. These data show a general increase of longitudinal dispersivity with the scale of the experiment.

The results of this review point to the need for carefully designed large-scale experiments in the unsaturated zone. The proposed experiments should be carried out at vertical scales of several tens of meters. Traceable solutes should be introduced with controlled water applications over areas of comparable horizontal extent. Because of its critical role in field-scale transport, three-dimensional spatial variability of soil properties should be measured as part of the experiments.

Section 1

INTRODUCTION

The purpose of this review is to summarize the current state of knowledge on field-scale physical processes that affect the transport of chemically inert solut s in the natural subsurface environment. More specifically, the results of this review are to serve as a guide in the planning and designing of comprehensive field experiments which are being proposed to explore field-scale transport processes in the saturated and unsaturated zones (Waldrop and Gelhar, 1984).

The term "field-scale" refers to vertical dimensions of tens to hundreds of meters and horizontal dimensions of tens of meters to kilometers. Scales of this order are typically those of concern in evaluations of the environmental consequences of waste disposal activities. Typical time scales associated with these spatial dimensions will be on the order of years or decades depending on the hydraulic conductivity and hydraulic gradient. As a consequence, it is impractical to carry out direct experiments to determine the pertinent transport parameters at these scales for each disposal site. Rather what is required are methods predicting the essential transport properties from practical small-scale measurements or extrapolation from a limited number of field-scale measurements. The overall goal of this research is to evaluate methods of predicting these field-scale parameters through comparison with large-scale field experiments.

The dispersion coefficient is the primary parameter of concern because it controls the degree of spreading and dilution of a solute plume. The effects of other parameters, including effective porosity, molecular diffusion, density and viscosity, are discussed briefly. The review is divided into sections dealing with the saturated and unsaturated zones. The approaches for reviewing the available data in these two zones differ somewhat because of differences in the current state of knowledge. In the saturated zone there is a substantial amount of field data which may establish trends and can be subjected to critical evaluation to determine its reliability. In the unsaturated zone the data are much more limited, and field experimentation is emphasized more in the context of a descriptive understanding of controlling process. Also, in the case of the saturated zone, theories of

$$Sn\frac{\partial C}{\partial t} + q_i\frac{\partial C}{\partial x_i} = \frac{\partial}{\partial x_i}\left(nSD_{ij}\frac{\partial C}{\partial x_j}\right) \qquad i, j = 1, 2, 3 \tag{1-2}$$

It is important to note that some investigators (Robertson and Barraclough, 1973; and Bredehoeft and Pinder, 1973) alternatively define the dispersion coefficient tensor as:

$$D^*_{ij} = nD_{ij} \tag{1-3}$$

In this review, when it was clear that D^*_{ij} was used in a study, we converted to the more common D_{ij} form using (1-3).

When the x_1 coordinate axis is aligned in the direction of mean fluid flow ($q_1 \neq 0$; $q_2 = q_3 = 0$), the dispersion coefficient tensor is often approximated in forms equivalent to

$$D_{ij} = \alpha_{ij}v + \delta_{ij}D_d \tag{1-4}$$

where

- α_{11} = α_L the longitudinal dispersivity
- α_{22} = $\alpha_{33} = \alpha_T$ the transverse dispersivity
- α_{ij} = 0 for $i \neq j$
- v = seepage velocity in the x_1 direction ($= q_1/nS$)
- D_d = effective diffusion coefficient
- δ_{ij} = Kronecker delta ($= 1$ if $i = j$, $= 0$ if $i \neq j$)

This form of the dispersivity, with isotropy of transverse dispersion, is strictly valid only for an isotropic porous medium (Bear 1972).

Many laboratory experiments have demonstrated that the classical equations, (1-2) and (1-4), are valid for homogeneous porous media in the laboratory for displacements on the order of about one meter. However, in the field, where the movement of solutes over distances of hundreds of meters is of interest, natural porous materials are not homogeneous. This, coupled with the fact that field-observed dispersivities are orders of magnitude larger than those found in small-scale laboratory tests (see e.g., Fried, 1975; Anderson, 1979), has recently led some investigators (Gelhar et al., 1979; Matheron and de Marsily, 1980; Gelhar and Axness, 1983) to question the validity of modeling field-scale solute transport only in the direction perpendicular to perfect layering. Stratified models have been developed using deterministic methods (e.g., Marle, 1967; Dieulin, 1980; Molz et al., 1983) and stochastic methods (Mercado, 1967; Gelhar et al., 1979; Matheron and de Marsily, 1980). Most of these approaches are similiar to the original work of G. I. Taylor (1953) on dispersion in tubes, but are of doubtful applicability on the field-scale (10^2 to 10^4 m) because real aquifers are not perfectly stratified over these dimensions.

A second general approach presumes that the spatial statistics of a heterogeneous velocity field are given, and determines the dispersion coefficient from the Lagrangian covariance of the velocity field (e.g., Dieulin, et al., 1981; Simmons, 1982; Tang et al., 1982; Winter, 1982). These methods are similar to the original work of G. I. Taylor (1921) on turbulent diffusion, although mathematical techniques are more complex. The key weakness of this velocity-based approach is that it requires the velocity covariance to be known in order to predict the dispersion coefficient. The velocity covariance is not directly measurable, and even if it could be, there remains the problem that the covariance of velocity will change as flow conditions change. The velocity-based method can be used in a nonpredictive mode by adjusting the velocity covariance function until agreement with an observed plume is accomplished; this approach was attempted by Simmons (1982) for the data from the Borden site.

A third general approach is based on the statisical description of the spatial variability of hydraulic conductivity. This approach has been developed using perturbation methods (Dagan, 1982; Gelhar and Axness, 1983; Gelhar, 1984) and Monte Carlo simulations (Warren and Skiba, 1964; Heller, 1972; Smith and Schwartz, 1980). The three-dimensional theory by Gelhar and Axness (1983) is the most general in that it includes the effects of local dispersion (or molecular diffusion) and anisotropy of the hydraulic conductivity covariance. Dagan (1982) treats the isotropic case with no local dispersion; his results agree exactly with those of Gelhar and Axness (1983) for this special case. An intermediate result of Gelhar and Axness (1983) also is identical to the velocity-based result of Winter (1982). The theory of Gelhar and Axness (1983) also produces the stratified case as a special case. Another conceptualization of the transport process is analogous to the so-called matrix diffusion model of transport in a single fracture bounded by a porous but impervious rock matrix (e.g., Neretnieks, 1980; Grisak and Pickens, 1980; Gillham, et al., 1984). In this case the aquifer is represented by a layer of permeable material through which all of the flow occurs, bounded by a practically impermeable

Section 2

SATURATED TRANSPORT PROCESSES

This portion of the review focuses on field-scale physical processes that control the transport of chemically inert solutes in saturated natural earth materials. Laboratory scale observations are not covered because these are not directly pertinent to field problems and are already reviewed extensively in textbooks and review articles (e.g., Fried and Combarnous, 1971; Bear, 1972; Fried, 1975; Dullien, 1979; Greenkorn, 1983). This review emphasizes dispersion coefficients but also considers the role of effective porosity, molecular diffusion, and fluid viscosity and density in the dispersion process. Included are (1) a general compilation of data from field sites, (2) detailed analyses for two selected sites, (3) evaluations of the available data and interpretations relating to modelling concepts, and (4) conclusions and recommendations relating to the design of field experiments.

SUMMARY OF FIELD OBSERVATIONS

A thorough review of published literature and numerous unpublished sources yielded field-scale values of dispersivity for various sites across the United States and in ten foreign countries. The review included consideration of data sources identified by Lallemand-Barres and Peaudecerf (1978), Anderson (1979), Pickens and Grisak (1981), Isherwood (1981), Mercer, et al. (1982), as well as others available to the authors. The Mercer et al. review dealt with aquifer thermal energy storage. In most of the studies cited by Mercer the dispersion coefficient is reported as a sum of thermal dispersion and hydrodynamic dispersion. Only when these values were reported separately could we determine the value of hydrodynamic dispersivity used in a particular study and include it in our summary table. An additional source emphasizing recent studies in Germany (Schröter, 1983) was located late in this work and could not be obtained for evaluation; therefore these sites are not included in the compilation. The findings of Schröter are similar, however, to the results presented in our literature summary.

Fifty-five sites were found where investigators reported longitudinal dispersivities (A_L) at specified scales. Since observations at many sites included results from multiple tracer tests, a total of 99 values of longitudinal dispersivity were recorded. In the few cases where values of transverse horizontal and vertical

Table 2-1

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific ation of reliabili of $A_L/A_T/$ (I,II,III
Ahlstrom et al. (1977) Hanford, WA	Glacio-fluviatile sands and gravels	64	Multi-layered				Ambient	Multiple sampling wells (contam.)	2D	^3H	Numerical simulation using discrete-particle-random-walk algorithm.	20,000	30.5/18	III
Biershenk (1959) Cole (1972) Hanford, WA	Glacio fluviatile sands and gravels	64	Multi-layered	$5.7x10^{-4}$ to $3.0x10^{-2}$ m/s	10	26 31	Ambient	1 injection well, 2 obs. wells (pulse input)		Fluor-scein	Breakthrough curve pulse width.	3500 4000	6 460	III III
Bredehoeft and Pinder (1973) Brunswick, GA	Limestone	50	Multi-layered	$1.7x10^{-1}$ m^2/s	35		Artificial	Radial con-verging (contam.)	2D	Cl^-	Calibration of numerical model.	1000	170/52*	III
Bentley and Walter (1983) WIPP	Fractured dolomite	5.5		$6.5x10^{-7}$ to $8.6x10^{-7}$ m^2/s	18	0.3	Artificial	2-well re-circulating (continuous input)	2D	PFB,SCN	Grove and Beetem analysis (1971).	23	5.2	III
Claesen and Cordes (1975) Amargosa, NV	Fractured Dolomite	15	Bilayered	$5x10^{-2}$ to $11x10^{-2}$ m^2/s		0.14-3.4	Artificial	2-well re-circulating (pulse input)	2D	^3H	Grove and Beetem type analysis.	122	15	III
Iris, P. (1980) Campuget (Gard, France)	Alluvial deposits	9		10^{-3} m/s		0.2	Artificial	Radial, al-ternating diverging, converging	3D	Heat, Cl^-	Fitting of a numerical model.	50	1	II
Daniels (1981) Nevada Test Site	Alluvium derived from tuff	500		$1.7x10^{-5}$ m/s		0.04	Artificial	Radial converging (contam.)	2D	^3H	Type-curve matching for pulse input (after Sauty, 1977).	91	10-30	III
Dieulin (1981) La Cellier (Lozere, France)	Fractured granite	20		$3x10^{-4}$ to $9x10^{-4}$ m/s	2-8	3	Artificial	Radial con-verging (pulse input)	2D	Cl^- I^-	Analytical solution after Sauty (1977).	6	0.5	II

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific-ation of reliability of $A_L/A_T/A_V$ (I,II,III)†
Dieulin (1980) Torcy (France)	Alluvial deposits	6		$3x10^{-4}$ m/s		0.5	Ambient	1 injection well, 3 observation wells (pulse input)	2D (re-sistiv-ity)	Cl^-	1-D solution to advection-dispersion equation.	15	3	III
Fenske (1973) Tatum Salt Dome, MI	Limestone	53		$4.7x10^{-6}$ m/s	23	1.2	Artificial	Radial divergent (pulse input)		^3H	1-D solution to advection-dispersion equation.	91	11.6	III
Fried (1971) Rhine Aquifer	Sand, Gravel and cobbles	12				9.6	Artificial	Radial divergent (pulse input)		Cl^-	Numerical solution to 1-D radial advection dispersion equation to find A_L.	6	11	II
Fried (1975) Rhine Aquifer (Salt Mines) Southern Alsace, France	Alluvial; mixture of sand, gravel, and pebbles with clay lenses	125		10^{-3} m/s			Ambient (regional)	Multiple sampling wells (contam.)	3D	Cl^-	Calibration of bidimensional, horizontal, monolayer, hydrodispersive model to simulate salt concentrations.	800	15/1	III
Fried (1975) Lyons, France (Sanitary Landfill)	Alluvial, with sand and gravel and slightly stratified with clay lenses	20	Locally multi-layered, globally non-layered			5.0	Ambient (regional)	Multiple wells downstream of landfill (contam.)	2D	Conduc-tivity	Calibration of bidimensional, horizontal, monolayer, hydrodispersive model to simulate salt concentrations.	600-1000	12/4	III
Goblet (1982) "Site B" France	Fractured granite	50		10^{-5} to 10^{-7} m/s		84	Artificial	Radial converging (pulse input)	2D	RhWt SrCl	Convolution solution accounting for borehole flushing effects; uses uniform flow eqn.	17	2	III
Grove (1977) NRTS, Idaho	Basaltic lava and sediments	76		$1.4x10^{-1}$ to $1.4x10^{1}$ m^2/s	10		Ambient	Regional multiple sampling wells (contam.)	2D	Cl^-	Numerical simulation using Galerkin finite element model.	20,000	91/91	III

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific-ation of reliability of $A_L/A_T/A_V$ (I,II,III)
Grove and Beetem (1971) Eddy County (near Carlsbad), New Mexico	Fractured Dolomite	12			12	3.5	Artificial	2-well re-circulating (pulse in-put)	2D	^3H	Approximate numerical solution of advection-dispersion equation.	55	38.1	III
Gupta et al. (1975) Sutter Basin, California	Sandstone, shale, sand, and alluvial sediments		Multi-layered				Ambient (regional)	Multiple sampling wells (envir. input)		Cl^-	Numerical simulation using Galerkin finite element model.	50,000	80-200/ 8-20	III
Halevy and Nir (1962) Landa and Zuber (1970) Nahal Oren, Israel	Dolomite	100			3.4	4.0	Artificial	Radial con-verging (pulse input)	2D	^{60}Co	Landa and Zuber analysis (1970).	250	6	II
Harpaz (1965) Southern Coastal Plain, Israel	Sandstone with silt and clay layers	90	Multi-layered			14	Artificial	Radial di-vergent (continuous input)	2D	Cl^-	1-D analytical solution to radial advection-dispersion equation.	28	0.1-1.0	II
Helweg and Labadie (1977) Bonsall Sub-basin, Calif.							Ambient	Multiple sampling wells (contam.)		Dis-solved solids	No data analysis; dispersivities chosen based on unpublished literature sources.	14,000	30.5/9.1	III
Hoehn (1983) Lower Glatt Valley, Swit-zerland	Layered gravel and silty sand	25	Multi-layered	$9.2x10^{-4}$ m/s to $6.6x10^{-3}$ m/s		3.4 1.8 1.2	Ambient	(River in-filtrating into ground-water) One well for dye injection, two sampling wells	2D	Uranine	Modified Landa and Zuber's approach by using Ivanovitch and Smith's (1978). Breakthrough curves show three peaks; author attributes these to varying K.	4.4	Layer 1 0.1 2 0.01 3 0.2	III
						8.6 4.1 1.7						10.4	1 0.3 2 0.04 3 0.7	III

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific-ation of reliability of $A_L/A_T/A_V$ (I,II,III)†
Kreft et al. (1974) Zn-Pb deposits Poland	Fractured dolomite	57	Two-layered	2.5×10^{-4} to 4.7×10^{-4} m/s	2.4	7.5 100	Artificial	Radial convergent (pulse input)		^{131}I	Lenda and Zuber analysis (1970).	22	44-110	II
		48	Two-layered	2.5×10^{-4} to 4.7×10^{-4} m/s	2.4	60.1 22.7	Artificial	Radial convergent (pulse input)		^{131}I	Lenda and Zuber analysis (1970).	21.3	2.1	II
Kreft et al. (1974) Sulfur deposits	Limestone	7	Multi-layered	1.1×10^{-4} m/s	12.3	10 10.8	Artificial	Radial convergent (pulse input)		^{58}Co	Lenda and Zuber analysis (1970).	27	2.7-27	II
	Limestone	7	Multi-layered	1.1×10^{-4} m/s	12.3	8.6	Artificial	Radial convergent (pulse input)		^{58}Co	Lenda and Zuber analysis (1970).	41.5	20.8	II
Lau et al. (1957) U. Cal. Berkeley	Sand and gravel	1.5	Contains localized clay lenses	9×10^{-4} m/s	30	7	Artificial	Radial divergent (continuous input) with multiple obs. wells	2D	Cl^-	Numerical/graphical solution to radial advection-dispersion equation.	19	2-3	I
Lee, et al. (1980) Perch Lake, Ontario (lakebed)	Sand		Layered sand containing small-scale heterogeneities	3.2×10^{-5} m/s		0.14	Ambient	5 injection wells, multiple obs. wells (pulse input)	3D	Cl^-	Breakthrough curve pulse width used to find A_L (1-D solution used).	≤6	0.012	III
Leland and Hillel (1981) Amherst, MA	Fine sand and glacial till	0.75	Multi-layered	2.4 to 3×10^{-5} m/s	40	.3-.6	Ambient	Multiple injection and sampling wells (pulse input)	3D	Cl^-	Analytical solution to 2-D advection-dispersion equation.	4	.05-.07	III

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classification of reliability of $A_L/A_T/A_V$ (I,II,III)
Ivanovich and Smith (1978) Dorset, England	Fractured chalk			2.2×10^{-3} m/s (fast pulse)	0.5	57.6	Artificial	Radial con-vergent, fast (pulse input)		^{82}Br	Curve fitting using two-pulse analytical model.	8	3.1	III
	Chalk			3.6×10^{-4} m/s (slow pulse)	2.3	9.6	Artificial	Radial con-vergent, slow (pulse input)		^{82}Br		8	1.0	III
Kies (1981) New Mexico State University Las Cruses, NM	Sand		Vertical variation of param-eters ob-served	9.55×10^{-5} m/s	42 (total porosity)		Ambient	Grid of samples screened at the water table. Tracer input was the out-flow of a tracer test in the un-saturated zone (contam.)	2D	Nitrate	Computer fit to break-through curve based on data from one sample site.	25	1.6/0.76	III
Klotz, et al. (1980) Dornach, Germany	Fluvio-glacial gravels	14				20	Ambient and artificial	Radial con-vergent (pulse input)	2D	^{82}Br Uranine	Moment method but assumed A_L constant. Landa and Zuber analysis (1970).	10	5, 1.9	II
Konikow (1976) Rocky Mt. Arsenal	Alluvium				30		Ambient	Multiple obs. wells (contam.)		Cl^-	Calibration of numerical model.	13,000	30.5	III
Konikow and Bredehoeft (1974) Arkansas River Valley (at La Junta), Colorado	Alluvium, in-homogeneous clay, silt, sand, gravel	25		2.4×10^{-4} to 4.2×10^{-3} m/s	20		Ambient	Multiple obs. wells (contam.)	2D	Dis-solved solids	Calibration of numerical model.	18,000	30.5/9.1	III
Kreft et al. (1974) Poland	Sand	2.5		3.1×10^{-5} to 1.5×10^{-4} m/s 1.2×10^{-4} m^2/s	24	29	Artificial	Radial con-vergent (pulse input)		^{131}I	Landa and Zuber analysis (1970).	5-6	0.18	II

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classification of reliability of $A_L/A_T/A_V$ (I,II,III)
Mercado (1966) Oavne Region, Israel	Sand and sandstone with some silt and clay	~ 80	Multi-layered	2.1×10^{-8} to 2.4×10^{-8} m^2/s	23.3	0.84-3.4	Artificial	Single well in-jection - withdrawal (continuous input)	3D	^{60}Co Cl^-	Analytical solution to 1-D advection-dispersion equation.	≤115 (obs. wells)	0.5-1.5 (injection phase)	I
Meyer, et al. (1981) Koeberg Nuclear Power Station, South Africa	Sand	20	Multi-layered			0.12	Ambient	1 injec-tion well; 3 obs. wells (pulse input)	3D	^{131}I	Solution after Lenda and Zuber (1970), modified to reflect tracer concentrations at various depths.	2-8	.01,.03,.01, .05 for layers; .42 for depth avg. (whole aquifer)	III
Molinari (1977) Sauty (1980) Sauty (1977) Bonnand, France	Sand	3	Some vertical layering	8.3×10^{-4} to 1.1×10^{-3} m^2/s	38		Forced Uniform Flow Field	Multiple sampling wells at different distances from source (pulse input)	2D	Heat	Fitting of axis-symmetric numerical model.	13	1.0	II
						2.7				I^-	Sauty's (1977) curve fitting techniques.	13	0.79	I
				1.0						^3H		13	1.27	I
				2.4						^{131}I		13	0.72	I
				1.0						^{131}I		26	2.23	I
				2.0						^{131}I		33.2	1.94/0.11	I
				2.0						^{131}I		32.5	2.73/0.11	I
New Zealand Ministry of Works and Development (1977) Heretaunga Aquifer, New Zealand														
Roya Hill Site	Greywacke gravel with cobbles	100	Highly hetero-geneous	$0.29 m^2/s$	22	150-200	Ambient	Multiple obs. wells (pulse input)	3D	^{131}I, RhWT, ^{82}Br, Cl^-, E. Coli	3D analytical solution for pulse input.	54-59	1.4 - 11.5/ 0.1 - 3.3/ 0.4 - 0.10	II

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific ation of reliabili of $A_L/A_T/$ (I,II,III
Flaxmere Site 2	Greywacke alluvium (gravels)	120	Highly hetero-geneous	0.37m^2/s	22	20-25	Ambient	Multiple obs. wells (pulse input)	3D	RhWT, ^{82}Br		25	0.3 - 1.5/-/ 0.06	II
Hastings City Rubbish Dump	Greywacke alluvium (gravels)		Highly hetero-geneous	.14,.35 m^2/s		20	Ambient	Multiple obs. wells (contam.)		Cl^-		290	41/10/0.07	III
Naymik and Barcelona (1981) Meredosia, IL (Morgan County)	Unconsol-idated sand and gravel	27		$2.2x10^{-2}$ to $4.3x10^{-2}$ m^2/s			Ambient	Multiple wells (contam. event)	2D	Ammonia	Calibration of 2D solute transport model.	16.4	2.13 - 3.35/ 0.61 - 0.915	III
Oakes and Edworthy (1977) Clipstone, UK	Sandstone	44		$2.4x10^{-6}$ to $1.4x10^{-4}$ m/s	32-48	5.6,4.0 9.6 2.4 3.6	Artificial	1)radial diverging and 2)converg-ing (pulse input)	2D 2D	^{82}Br Cl^-,I^-	Numerical solution of of advection-dispersion equation.	6 3 6 3	0.16, 0.38 0.31 0.6 0.6	II
Papadopulos and Larson (1978) Mobile, Alabama	Medium to fine sand interspersed with clay and silt	21	Assumed none	$5x10^{-4}$ m/s (horiz.) and $5.1x10^{-5}$ m/s (vert.)	25	0.05	Artificial	Radial di-vergent (continuous input)	2D	Heat	Numerical simulation using finite difference techniques.	57.3	1.5	II
Pickens and Grisak (1981) Chalk River	Sand	8.5	Small-scale cross bedding	$2x10^{-5}$ to $2x10^{-4}$ m/s	38	0.15	Artificial	2-well re-circulating (continuous input)	3D	^{51}Cr	Analysis after Grove and Beetem (1971).	8	0.5	III

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific-ation of reliability of $A_L/A_T/A_V$ (I,II,III)†
Pickens and Grisak (1981), cont'd.	Sand	8.5	Small-scale cross bedding	$2x10^{-5}$ to $2x10^{-4}$ m/s	38	0.15	Artificial	1-well injection withdrawal (continuous input)	3D	^{131}I	Analysis after Mercado (1966) at obs. wells.	3	0.002-0.09 for layers (0.007 = mean for layers)	III
											Analysis after Mercado (1966) and Gelhar and Collins (1971) at pumping well.	3	0.03	III
											Analysis after Mercado (1966) at obs. wells.	5	0.004-0.015 for layers (0.008 = mean for layers)	III
											Analysis after Mercado (1966) and Gelhar and Collins (1971) at pumping well.	5	0.09	III
Pinder (1973) Long Island	Glacial outwash	43	Multi-layered	$7.5x10^{-4}$ m/s	35	0.43	Ambient (regional)	Multiple wells (contam.)	3D	Cr^{+6}	Simulation of chromium concentrations using Galerkin-finite element approach.	10^3	21.3/4.2	III
Roberts, et al. (1981) Palo Alto Baylands	Sand and gravel, silt	2	Multi-layered	$1.25x10^{-3}$ m^2/s (lower aquifer) $5.0x10^{-4}$ m^2/s (upper aquifer)	25	15.5 12.0 3.5 25.6 7.9	Artificial	Radial divergent (continuous input)	2D	Cl^-	Visual matching of observed values with type curves given by Ogata and Banks (1961), using a 1-D uniform flow solution.	11 20 40 16 43	5 2 8 4 11	III III III III III
Robertson (1974) Robertson and Barraclough (1973) NRTS, Idaho	Basaltic lava and sediments	76		$1.4x10^{-1}$ to $1.4x10^{1}$ m^2/s	10	1.5-8	Ambient	Regional multiple sampling wells (contam.)	2D	Cl^-	2-D solute transport used and solved numerically by the method of characteristics and compared to field data.	$2x10^4$	910/1370*	III

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific-ation of reliability of $A_L/A_T/A_V$ (I,II,III)†
Robson (1974 and 1978) Barstow, CA	Alluvial sediments	27	Multi-layered	2.1×10^{-4} to 1×10^{-2} m^2/s	40		Artificial	2-well re-circulating (continuous input)	2D	Cl^-	Digital program used to calculate breakthrough curves, using analysis after Grove (1971).	6.4	15.2	III
					40	3	Ambient	Regional multiple sampling wells (contam.)	2D	Dis-solved solids	Numerical model after Bredehoeft and Pinder (1973).	10^4	61/18	III
Robson (1978) Barstow, CA	Alluvial sediments	30.5	Multi-layered	5×10^{-4} m/s	40		Ambient	Regional multiple sampling wells (contam.)	3D	Dis-solved solids	Simulation using vertical 2-D solute transport model.	3200	61/-/0.2	III
Rabinowitz and Gross (1972) Roswell Basin New Mexico	Fractured limestone	61	Multi-layered	1.1×10^{-2} to 2.9×10^{-1} m^2/s	1	11-21	Ambient	Regional-multiple sampling wells (envir. input)	2D	^3H	Assumed input function determined A_L by matching 1-D pulse solution with concentration data.	~ 32,000	20-23	III
Rousseiot (1977) Blyes-Saint-Vulbas near Lyon, France	Clay, sand and gravel	12	Multi-layered	6.5×10^{-3} to 1.5×10^{-2} m/s	14 2.1-18 1.8-5.9 11-24	18 11.5,3.8 46.7,16 24	Artificial	Radial converging (pulse input)	2D	I^-	Type curve matching to individual layers after Sauty (1977).	9.3 5.3 10.7 7.1	6.9 0.3, 0.7 0.46, 1.1 0.37	II III III II
Sauty (1977) Corbas	Sand and gravel	12				125,100 15.5,78 6.9	Artifical	Radial converging (pulse input)	2D	I^-	Type curve matching to individual layers.	25 50 150	11, 1.25 25, 6.25 12.5	III III II
Segol and Pinder (1976) Cutler Area, Biscayne Bay Aquifer, Florida	Fractured limestone and calcareous sand stone	30.5	Assumed none	0.45×10^{-2} m/s (horiz.) and 0.09×10^{-4} m/s (vert.)	25	20	Ambient	Regional multiple sampling wells (envir. input)	3D	Cl^-	2-D Galerkin finite element simulation.	490	6.7/0.67	III

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific-ation of reliability of $A_L/A_T/A_V$ (I,II,III)†
Sudicky et al. (1983) Borden	Glacio-fluvial sand	7-27	Moderate	4.8×10^{-5} to 7.6×10^{-5} m/s	38	0.07-0.25	Ambient	Multiple sampling wells (pulse input)	3D	Cl^-	Analytical solution to 3-D advection-dispersion equation with approach after Dieulin (1980).	11 0.75	0.08/0.03 0.01/0.005	I I
	Glacio-fluvial sand	7-27	Moderate	10^{-5} to 10^{-7} m/s	38	0.01-0.04	Ambient	Multiple wells (continuous input-25yrs) (envir. input)	3D	^3H	Simulation using 1-D solution to the advection-dispersion equation.	600	30-60	III
Sykes et. al (1982b and 1983) Borden	Sand			5.8 to 7.2 x 10^{-5} m/s	35		Ambient	Multiple sampling wells (pulse input)	3D	Cl^-	Simulation using multi-dimensional finite difference model.	700 700	7.6/ /0.076 -1.5 (3D model) 1.5-7.6/-/0 -0.15 (2D model)	III III
Sykes et al. (1983) Mobile, Alabama	Sand, silt and clay	21		5×10^{-4} m/s (horiz.) and 2.5×10^{-5} m/s (vert.)	25	0.05	Artificial	Radial divergent (continuous input)	3D	Heat	Simulation using a 3-D finite difference model.	57.3	0.76/-/0.15	II
Vaccaro and Bolke (1983) Spokane Aquifer, Washington and Idaho	Glacio-fluvial sand and gravel	152		9×10^{-5} m^2/s to 6.5 m^2/s	7-40	0.003-2.8	Ambient	Multiple obs. wells (contam.)		Cl^- (from land use activities)	Calibration of numerical model.	43,400	91.4/27.4	III
Valocchi et al. (1981) Palo Alto Baylands	Sand and gravel, silt	2	Multi-layered	1.25×10^{-3} m^2/s (lower aquifer) 5.0×10^{-4} m^2/s (upper aquifer)	25	27	Artificial	Radial divergent (continuous input)		Cl^-	2-D advection-dispersion simulation model used (Galerkin finite-element).	16	1.0/0.1	I

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific-ation of reliability of $A_L/A_T/A_V$ (I,II,III)†
Walter (1983) WIPP	Fractured dolomite	7		8.0×10^{-5} m^2/s	0.7 and 11 (along separate paths)	4.5, 2.4	Artificial	Radial con-vergent (pulse input)	2D	MTFMB, PFB MFB, para-FB	Type curve matching after Sauty (1977).	30	10-15	III
Webster, Procter and Marine (1970) Savannah River Plant, SC	Crystalline, fractured schist and gneiss	76		3.6×10^{-7} m/s		1.3 21.4	Artificial	2-well (pulse input) (recir-culating flow)	2D	^{85}Sr ^{82}Br	Grove and Beetem analysis (1971).	538	134	III
Werner, et al. (1983) Hydrothermal Test Site Aefligen, Switzerland	Gravel	20		6×10^{-3} m/s	17	9.1	Ambient	Single injection well with monitoring network	3D	Heat	Calibration of numerical model.	700 37 105 200	130-234 131 208 234	III III III III
Wiebenga (1967) Lenda and Zuber (1970) Burdekin Delta, Australia	Sand and gravel	6.1		5.5×10^{-3} m/s	32	29	Artificial	Radial convergent (pulse input)		^{131}I ^{3}H	Analysis after Lenda and Zuber (1970).	18.3	0.26	II
Wilson (1971) Robson (1974) Tuscon, AZ	Unconsoli-dated gravel sand, and silt	4.3	Multi-layered	5.75×10^{-3} m^2/s	38		Artificial	2-well test (non-recirculat-ing) (con-tinuous input)	3D	Cl^-	Digital program used to calculate breakthrough curves.	79.2	15.2	III

Table 2-1 (continued)

SUMMARY OF FIELD OBSERVATIONS FOR THE SATURATED ZONE

Reference and Site Name	Aquifer Material	Average Aquifer Thickness (m)	Apparent Hetero-geneity	Hydraulic Conductivity (m/s) or Transmissivity (m^2/s)	Effective Porosity (percent)	Velo-city (m/d)	Flow System	Test Configu-ration	Moni-tor-ing	Tracer Used	Method of Data Interpretation	Scale of Test (m)	Disper-sivity $A_L/A_T/A_V$ (m)	Classific-ation of reliability of $A_L/A_T/A_V$ (I,II,III)†
Wood (1981) Aquia Formation Southern MD	Sand	1000		2.9×10^{-4} to 8.7×10^{-4} m^2/s	35	0.0003-0.0007	Ambient	Envir. input		Na^+	Concept of hydrochemical facies is used. Aquifer is viewed as a large column; dispersion is determined by observing solute concentration contours in space and using a 1-D solution to the advection-dispersion equation after Ogata and Banks (1961).	10^5	5,600-40,000	III
Wood and Erlich (1978) and Bassett, et al. (1980) Lubbock, TX	Sand and gravel	17	Multi-layered	3.2×10^{-3} to 4.4×10^{-3} m^2/s		78	Artificial	Radial convergent (pulse input)	2D	I^-	Solution after Gelhar and Collins (1971).	1.52	0.015	II

Notes

A_L = Longitudinal dispersivity
A_T = Horizontal transverse dispersivity
A_V = Vertical transverse dispersivity

* a porosity-corrected dispersivity value.
† For description of classification criteria, see text.

PFB = Pentafluorobenzoate
SCN = Thiocyanate
MTFMB = Metatrifluoromethylbenzoate
OFB = Orthofluorobenzoate
MFB = Metafluorbenzoate
Para-FB = Parafluorobenzoate

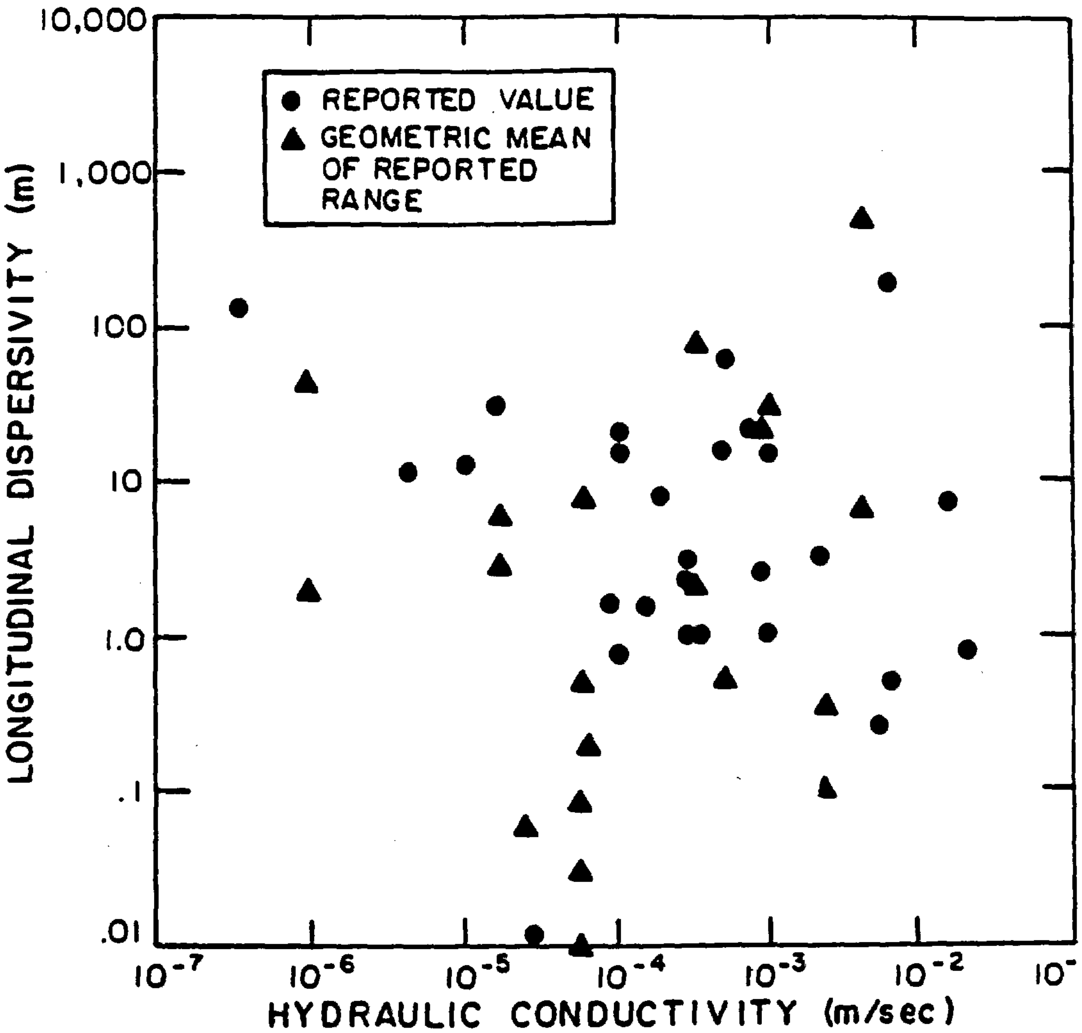

Figure 2-2. Hydraulic Conductivity Versus Longitudinal Dispersivity

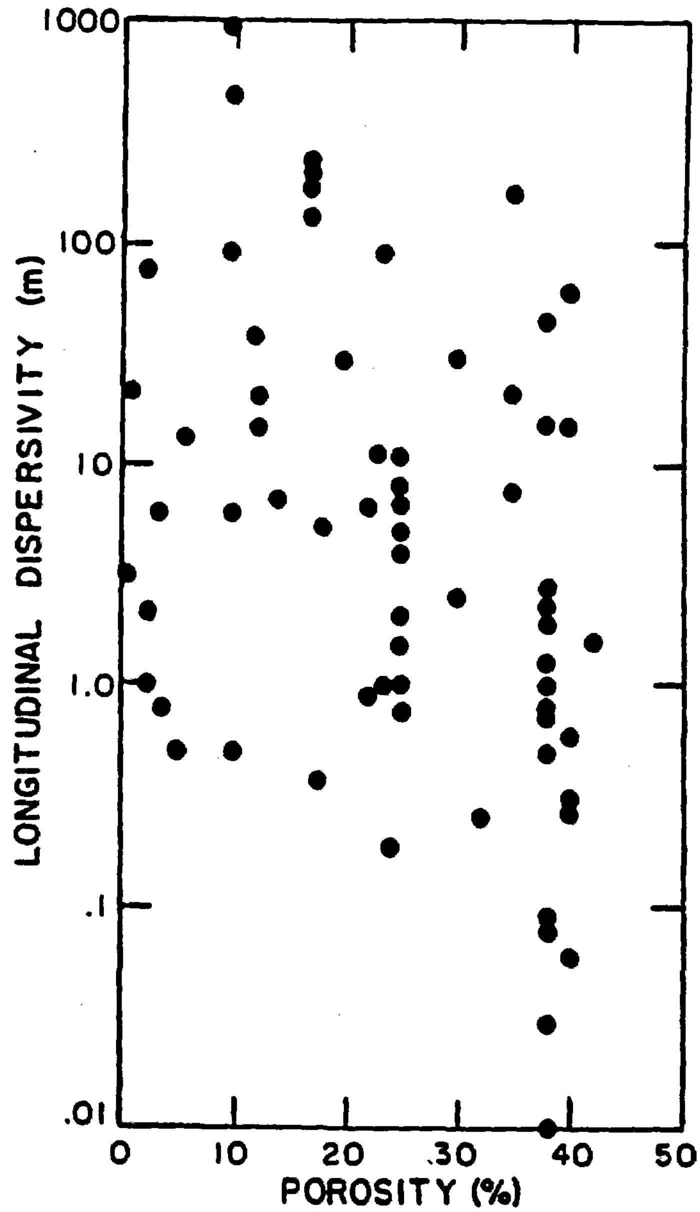

Figure 2-3. Porosity Versus Longitudinal Dispersivity

the 55 sites examined, artificial flow systems were used in roughly half of the experiments.

Ambient flow systems are important in determining regional dispersivities, i.e., on scales of greater than approximately 100 m. Usually dispersivities are calculated for such regional systems by calibrating a numerical model of a contaminant or natural tracer that is observed in multiple wells in the study area. Controlled tracer tests have not been performed on the scale of the regional models because of the lengthy observation times; however, some smaller scale tests (less than 100 m) have been performed in ambient flow systems. Ambient (or "natural gradient") flow systems were used in 27 of the sites examined.

A third type of flow system that was used in only one case, Bonnaud, France (Sauty, 1977) is referred to as "forced uniform flow." In this type of system, a uniform flow field is generated between two lines of equally spaced wells, one line recharging and one line pumping, with both screened to the full depth of the aquifer.

Monitoring configuration (two dimensional vs. three dimensional) was noted due to the influence of vertical mixing in an observation well on the concentration of tracer in a water sample. Several studies (Meyer, et al., 1981; Pickens and Grisak, 1981) have shown that when a tracer is not injected over the full aquifer depth, vertically-mixed samples underestimate the tracer concentration and consequently, the longitudinal dispersivity is overestimated. This can occur when the tracer occupies only a portion of the vertical thickness, but a sample from the entire thickness is taken. The true tracer concentration is then diluted with essentially tracer-free water. If one then attempts to model the diluted ("measured") concentration, the dispersivity must be overestimated in order to obtain enough tracer spreading (dispersion) to yield the "measured" concentration. Where point sampling was reported, it is noted as "three-dimensional"; this was the case at 19 sites. Although two-dimensional (vertically-mixed) sampling was reported for most other sites, there were many sites for which no indication of whether point or fully-mixed sampling was performed.

Noting information on test configuration is important because some tests are of questionable value in that they may not be providing a measure of actual field dispersivity. The various test configurations used to generate artificial flow systems are: (1) radial convergent, where the center well is pumped and the tracer is introduced into the observation well; (2) radial divergent, where the tracer solution is injected into the center well and the tracer breakthrough is observed at the observation well; (3) single well injection - withdrawal, where the tracer solution is injected into the center well for a period of time and then pumped back out of the same well with tracer measurement during the pumping cycle; (4) two-well or "doublet", where the tracer solution is injected in one well with measurement of the tracer concentration as a function of time in the pumping well. The details of these tests are discussed by Thompson (1981).

Typically at least one observation well is used for sampling tracer concentrations with these tests, but sometimes multiple observation wells are used. Also noted in Table 2-1 is whether the tracer input was a pulse or step. This information is important in evaluating the interpretation of the data from each tracer test.

The tracers used to measure transport at each site were also recorded as a factor to be considered in evaluating the reliability of the calculated dispersivities. The tracers used to determine dispersivity can be classified as artificial, environmental, or contaminant. Artificial tracers are those which are purposely injected into the groundwater, and therefore the input conditions are usually well defined. Also they are not usually present naturally in the groundwater, or are injected at concentrations very much higher than natural background levels. Artificial tracers used in the published literature reviewed included: ^{85}Sr, ^{90}Sr, Cl^-, I^-, uranine, ^{131}I, rhodamine, fluorscein, pentafluorobenzoate (PFB), thiocyanate (SCN) anions, metatrifluoromethylbenzoate (MTFMB), orthofluorobenzoate (OFB), metafluorobenzoate (MFB), parafluorobenzoate (para-FB), ^{51}Cr, temperature (hot or cold water), ^{60}Co, ^{58}Co, and ^{82}Br.

Environmental tracers and chemical contaminants can also be used to determine dispersivities, although their input conditions are not as well defined as those of artificial tracers. Environmental tracers are constituents associated with uncontrolled natural changes occurring in the groundwater before the start of a study. Tritium, sodium, and chloride were the cases examined. Contaminant tracers typically result from accidental events such as seepage of industrial wastes from disposal ponds or roadside chemical spills. Hexavalent chromium and ammonia are examples of contaminant-type tracers used to determine field-scale dispersivity that were found in the literature search.

Also noted for each test situation was the type of data interpretation used by the author(s) in determining dispersivity. Data interpretations were nearly as varied as the number of tests conducted; they ranged from one-dimensional analytical solutions to numerical methods, or sometimes some combination of several techniques. This information was collected to use as part of the subsequent judgement of the reliability of the dispersivity values.

As a result of this general review, several observations can be made based on the data presented in Table 2-1 and Figure 2-1. First, it would appear that dispersivity increases indefinitely with scale. Field observations of dispersivity ranged from 0.01 m to approximately 5500 m at scales of 0.75 m to 100 km. These values of field-scale dispersivity do not appear to be related to the type of aquifer material (porous media versus fractured media) or the thickness of the aquifer evaluated; for example, at each distance there was frequently a two-order magnitude range in dispersivity for similiar aquifer materials.

From the examination of the cases where three-dimensional measurements of solute concentrations were made, it is clear that tracers tend to travel in zones of limited vertical extent, and that vertical mixing of the tracer as it travels through the aquifer is often very small (Sudicky, Cherry, and Frind, 1983; LeBlanc, 1982).

DETAILED ANALYSIS OF SELECTED SITES

Two sites having extensive, good quality data were selected for detailed analysis illustrating methodology and additional interpretation. The Borden landfill site in Canada is used to illustrate the method of second moment analysis to determine three-dimensional dispersion characteristics and demonstrate displacement dependence of the dispersivities. A generalized moment method of this type will be used to analyze the proposed field experiments; the observations are designed for this purpose. The Bonnaud site in France is analyzed to illustrate how data on spatial variability of hydraulic conductivity can be used to predict dispersivities.

Borden Landfill, Canada

Sudicky, Cherry, and Frind (1983) reported the results of a natural gradient tracer test using chloride as a tracer at the Borden Landfill in Ontario, Canada. Point sampling was used to determine tracer breakthrough concentration, and a three-dimensional analytical solution to the advection-dispersion equation was fitted to the field data to obtain estimates of longitudinal, transverse, and vertical dispersivities. Values of 0.08 m and 0.03 m were obtained for longitudinal and transverse dispersivities, respectively, at a distance of 11.0 m from the source. Vertical dispersion was found to be very weak and was attributed to molecular diffusion.

The data gathered for this study are considered to be of high quality because multilevel sampling was employed and the test was conducted under ambient field conditions. However, the values derived for the dispersivities are questionable because the analytical solution used assumes that dispersivities are constant in time and space. The purpose of this section is to use the method of moments (see e.g., Fischer, et al., 1979) as an alternate method of analysis to estimate the dispersivities. This method is advantageous in that it does not make the a priori assumption that the dispersivities are constant in time (or displacement distance); conversely, it can be used to make observations about the nature of the dispersivities, e.g., whether they are constant or are growing with time (mean travel distance) from the source.

For steady uniform saturated (S = 1) flow in the x_1 direction, where x_1, x_2, and x_3 are the principal axes of the dispersion tensor, and D_{ij} is spatially constant, (1-2) can be written in a moving coordinate system as:

$$\frac{\partial C}{\partial t} = D_{ij} \frac{\partial^2 C}{\partial \xi_i \partial \xi_j} \tag{2-1}$$

where

$$\xi_1 = x_1 - q_1 t/n$$
$$\xi_2 = x_2$$
$$\xi_3 = x_3$$

Multiplying (2-1) by ξ_k^2 and integrating

$$\iiint_{-\infty}^{\infty} \xi_k^2 \frac{\partial C}{\partial t} d\xi_1\, d\xi_2\, d\xi_3 = \frac{d}{dt} \iiint_{-\infty}^{\infty} \xi_k^2\, C\, d\xi_1\, d\xi_2\, d\xi_3 =$$

$$\iiint_{-\infty}^{\infty} D_{ij}\, \xi_k^2 \frac{\partial^2 C}{\partial \xi_i \partial \xi_j} d\xi_1\, d\xi_2\, d\xi_3$$

Noting that $D_{ij} = 0$ when $i \neq j$, and integrating the last term by parts using the condition $C \to 0$ as $\xi_k \to \infty$, yields the following expression relating the second moment of the concentration to the dispersion coefficient

$$\frac{d\sigma_k^2}{dt} = 2D_{kk} \qquad k = 1, 2, 3 \text{ (no sum)} \tag{2-2}$$

where

$$\sigma_k^2 = \frac{\iiint_{-\infty}^{\infty} (x_k - \bar{x}_k)^2 \, C \, d\xi_1 \, d\xi_2 \, d\xi_3}{\iiint_{-\infty}^{\infty} C \, d\xi_1 \, d\xi_2 \, d\xi_3} \tag{2-3}$$

where

$\bar{x}_k$ = mean displacement of the tracer in the x_k direction

Using the dispersivity relationship of (1-5) and taking the mean velocity $v = q_1/n$ as constant,

$$\frac{d\sigma_k^2}{d\bar{x}} = 2A_k \tag{2-4}$$

where $\bar{x} = vt$ and A_k denotes one of the three principal components of the dispersivity.

Given that the x_1 axis has been aligned with the mean flow in the development of (2-2), $\bar{x}_2 = \bar{x}_3 = 0$.

If values of σ_k^2 versus $\bar{x}$ are plotted, then the nature of A_k as a function of time (or mean travel distance) can be obtained. For example, if σ_k^2 versus $\bar{x}$ plots as a straight line, then A_k can be considered a constant. Rearranging (2-4) and integrating with A_k constant gives

$$\sigma_k^2 = 2 A_k \bar{x} + \text{const} \tag{2-5}$$

so that the macrodispersivity is given by

$$A_k = m/2 \tag{2-6}$$

where m is the slope of the σ_k^2 versus $\bar{x}$ graph.

The values of A_k so obtained can be interpreted as average macrodispersivities between two locations.

If values of σ_k^2 versus $\bar{x}^2$ plot as a straight line, then A_k can be viewed as growing linearly with distance. Rearranging and integrating (2-4) with A_k a linear function gives

$$\sigma_k^2 = a + b\bar{x}^2 \qquad a, b = \text{constants} \tag{2-7}$$

and A_k can be expressed as

$$A_k = b\,\bar{x} \tag{2-8}$$

where b is the slope of the straight line.

This type of analysis was carried out using the data reported by Sudicky, et al. (1983); the moments were determined from the data for the "slow zone" in their Figures 4 and 5 for 29 and 121 days plus an estimate based on the initial injection configuration. A plot of σ_k^2 versus $\bar{x}$ (where $x_1 = x$, $x_2 = y$, $x_3 = z$) is shown in Figure 2-4; σ_k^2 versus $\bar{x}^2$ is shown in Figure 2-5. Because there are only two data points, it is difficult to ascertain which behavior the dispersivities may be exhibiting (e.g., constant vs. growing linearly with mean distance). Earlier reports (Sudicky and Cherry, 1979, Figure 10; Sudicky, et al., 1983, Figure 15) of more detailed data for the "fast zone" clearly show increases of longitudinal dispersivity with distance at the Borden site.

Shown in Table 2-2 are the values of longitudinal, transverse, and vertical dispersivities derived from these analyses compared with those reported by Sudicky, et al. (1983). Assuming a linear change in dispersivity with displacement (the center column of Table 2-2), the value obtained for A_x is more than four times, and the value obtained for A_y is almost three times, larger than the values obtained for dispersivities using the three-dimensional analytical solution. The moment analysis therefore indicates that *a priori* assumptions regarding dispersivities could be incorrect. Based on the fact that field dispersivities apparently

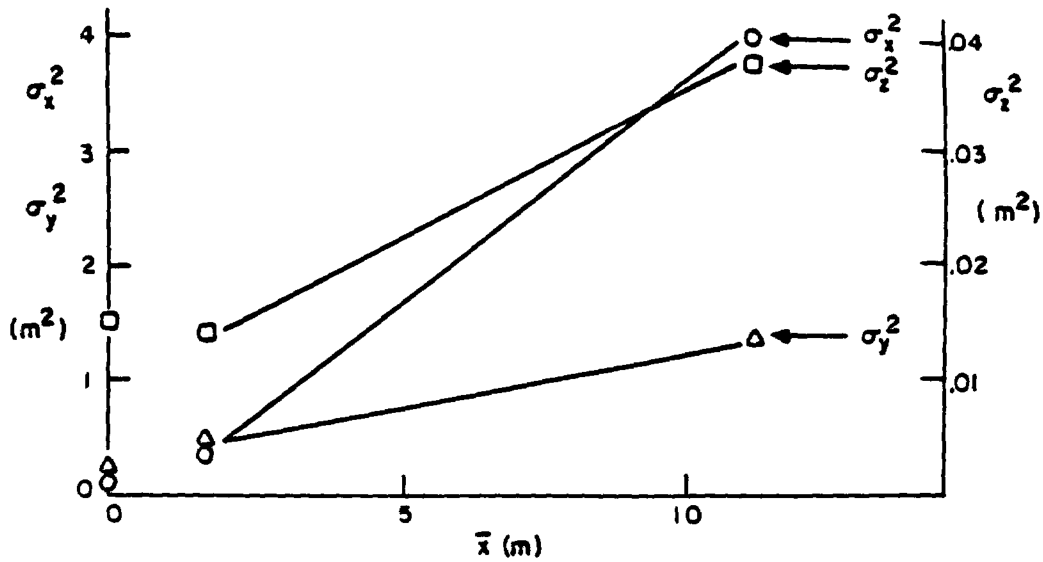

Figure 2-4. Plot of σ_K^2 Versus $\bar{x}$ for Borden Data

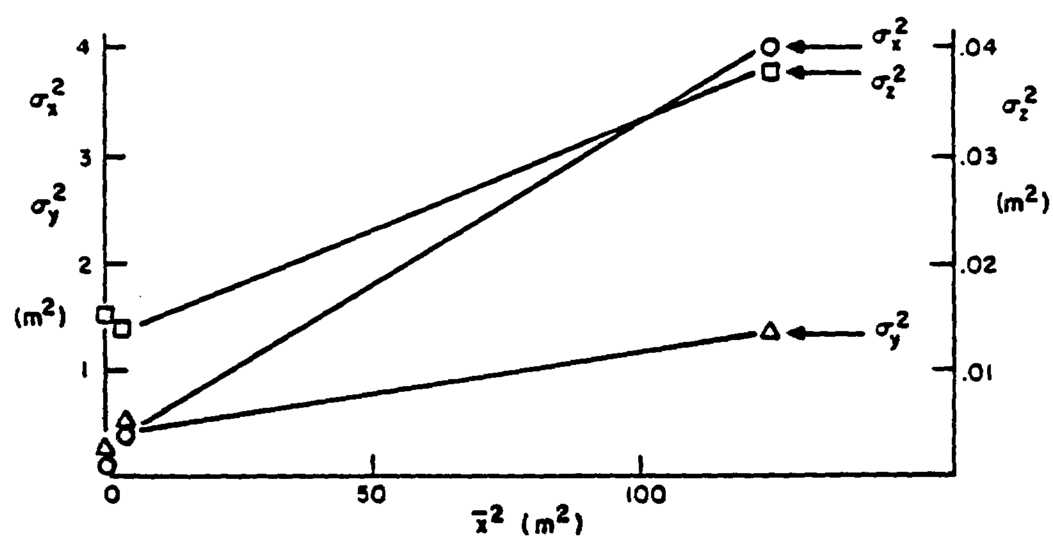

Figure 2-5. Plot of σ_K^2 Versus $\bar{x}^2$ for Borden Data

Table 2-2

DISPERSIVITIES DETERMINED FOR BORDEN LANDFILL, ONTARIO

	Method of Moments		
	Constant A_r (Figure 2-4 and Equation 2-6) 1.7 to 11.2 m	Linear A_r (Figure 2-5 and Equation 2-8) at 11.2 m	Analytical Solution by Sudicky, et al. 1983* Scale = 11.2 m
A_x(m)	0.19	0.34	.08
A_y(m)	0.045	0.084	.03
A_z(m)	0.00125	0.0019	9.3×10^{-5}
A_x/A_y	4.2	4.0	2.7
A_y/A_z	36	44	321

* Based on $D_{tv} = 1.0 \times 10^{-10}$ m^2/s with $v = 1.07 \times 10^{-6}$ m/s (Sudicky, et al., 1983, p. 105).

increase with scale, this alternate method of analysis allows dispersivities to be calculated without assuming that they are constant.

Note that the value of vertical dispersivity for Sudicky, et al. (1983) in Table 2-2 was based on their value attributed to molecular diffusion ($D_{tv} = 1.0 \times 10^{-10}$ m^2/s) which they claim produced "best agreement" with measurements. The moment analysis shows that the vertical dispersivity A_z is an order of magnitude larger, though still rather small (1 to 2 mm). This small vertical dispersivity is consistent with the small values predicted from stochastic theory (see Gelhar and Axness, 1983, Table 3, Case 2).

Bonnaud, France

The French Bureau of Geologic and Mining Research (BRGM) carried out a number of tests to determine aquifer parameters at a field site near Bonnaud, France. Tracer tests in a confined shallow sand aquifer under conditions of pulse injection and forced uniform flow yielded data from which dispersivities could be calculated. Using Sauty's type curve matching techniques, values of dispersivity were found to be: $A_L = 2.7$ m; $A_T = 0.11$ m (Sauty, 1977).

Gelhar and Axness (1983) have presented a stochastic theory for predicting macrodispersion in aquifers. The data from Bonnaud is used here to demonstrate an application of Case 2 (Horizontal Stratification with Horizontal and Vertical Anisotropy, p. 170) of that theory. The appropriate predictive equation is Equation 69 of Gelhar and Axness (1983); Section 7 of that paper thoroughly illustrates the application of the theory with several sample calculations. Readers interested in the details of the stochastic approach should consult that source.

The parameters required for application of the stochastic theory were determined as follows.

Logs of vertical fluid velocity in pumped boreholes (Peaudecerf, et al., 1975) were used to estimate the autocovariance function of the log hydraulic conductivity assuming that the hydraulic conductivity was proportional to the local inflow rate along the borehole. From the autocovariance, the variance of lnK and the vertical correlation scale (λ_3) were estimated to be 1.20 and 0.50 m, respectively.

From the hot water injection tests, maps of hydraulic head were constructed by BRGM (see e.g., Fabris, 1978, PL 5-6). This information provided an estimate of $\phi = 60°$, the angle between the direction of flow and the major axis of the hydraulic conductivity tensor. In addition, measurement of the axes of the hydraulic head ellipsoid provided an estimate of $K'_{11}/K'_{22} = 1.76$. Knowing this ratio, the value for σ^2_{lnK}, and assuming a value of λ_1/λ_3 of 20, λ_1/λ_2 was estimated to be 19 from Equation 69 and Figure 4a of Gelhar and Axness (1983). The implied configuration of the heterogeneous sediment bodies, with the horizontal scale λ_2 almost the same as the vertical scale λ_3, is not at all realistic. We expect λ_2 to be much larger than λ_3 in stratified sediments. With these initial parameter estimates, the longitudinal dispersivity A_{11} is equal to 0.56 m (from Equation 69) which is a factor of 5 smaller than that determined from the tracer test ($A_{11} = 2.7$ m, Sauty, 1977). We suspect that the lnK covariance and the resulting parameters, σ^2_{lnK} and λ_3 determined from the borehole flow meter, are not reliable. Hufschmied (1983) has shown that ambient vertical hydraulic gradients can have a very important effect in such borehole flow measurements. The effects of vertical gradients were not accounted for in the measurements at the Bonnaud site.

Assuming the following plausible values of the parameters ($\sigma^2_{lnK} = 1.2$, $\lambda_3 = 1$ m, $\lambda_1 = 20$ m, $\lambda_2 = 8$ m), Equation (2-9) yields $A_{11} = 3.1$ m, $A_{22}/A_{11} = 0.04$ (the ratio of longitudinal to transverse dispersivity), and $K'_{11}/K'_{22} = 1.1$, which are in reasonable agreement with the observed values ($A_L = 2.7$ m; $A_T = 0.11$ m; $K'_{11}/K'_{22} = 1.76$). In any case, the main point is to illustrate the method of analysis which can be used to determine field-scale dispersivities. This analysis also demonstrates the need for developing reliable field methods of determining the statistical parameter of hydraulic conductivity variations.

A number of dispersivity determinations have been done at the Bonnaud site. Molinari and Peaudecerf (1977), in Table 1, report longitudinal dispersivity values ranging from less than 0.5 m to over 2 m for displacement distances of 13 m to 35 m. These data clearly show an increase of dispersivity with distance (see also Figure 9, Gelhar, et al., 1979). Single-well hot water injection tests (Sauty, et al., 1982) have also been used to determine the longitudinal dispersivity for heat transport. In these experiments, which involved displacement of the thermal front out to 13 m, the dispersivity was found to be in the same range as for chemical tracers (Molinari and Peaudecerf, 1977).

EVALUATION AND INTERPRETATION

Reliability of Dispersivity Data

The information collected on test conditions and data interpretations was used primarily to assess the reliability of the reported dispersivities. The objective was to determine, insofar as possible, whether field-scale dispersivity was actually measured by the tests conducted. Before proceeding, we wish to make a clear distinction between the reliability of the reported dispersivity and the worth of a study. Often, the purpose of a study was something other than the determination of dispersivity. Our classification of dispersivity is not intended as a judgement on the quality of a study as a whole, but rather to give us some criteria with which to screen the large number of data values obtained. By using the more reliable data, conclusions which evolve from the data will be more soundly based and alternative interpretations may become apparent. The primary factors considered in making this assessment included: the monitoring detail employed (two-dimensional as opposed to three-dimensional), tracer used, type of test run, and data interpretation.

The reliability of the reported dispersivities was classified as either high, intermediate, or low. Table 2-3 lists the criteria used to classify a dispersivity as either high or low reliability. No specific criteria were defined for the intermediate classification because it was intended to encompass those dispersivity values that simply did not fall into the high or low group. As defined, the classifications do not place a strict numerical confidence limit on reported dispersivity, but rather are intended to give an order of magnitude estimate of the confidence we placed on a given value. In general, we consider high reliability dispersivities to be accurate to within a factor of about two or three. Low reliability data are considered to be no more accurate than within one or two orders of magnitude. Intermediate reliability falls somewhere between the extremes.

High Reliability Dispersivity Values. For a reported dispersivity to be classified of high reliability, each of the following criteria must have been met.

1. The tracer test was either uniform flow, diverging radial flow or a two-well pulse test (without recirculation).

   These three test configurations produce breakthrough curves which are sensitive to the dispersion coefficient and appear to work well in field applications. The radial converging flow test is generally considered less satisfactory because breakthrough curves at the pumping well for convergent radial flow tests frequently show tailing that complicates the interpretation of these tests. Some researchers attribute this behavior to two or more discrete layers and try to reproduce the observed breakthrough curve by the superposition of breakthrough curves in each layer, where the properties of each layer may

Table 2-3

CRITERIA USED TO CLASSIFY THE RELIABILITY OF THE REPORTED DISPERSIVITY VALUES

High Reliability Dispersivity Values

- The tracer test was either uniform flow, radial diverging flow, or two-well instantaneous pulse test (without recirculation).
- The tracer input was well defined.
- The tracer was conservative.
- The spatial dimensionality of the tracer concentration measurements was appropriate.
- The analysis of the tracer concentration data was appropriate.

Low Reliability Dispersivity Values

- The two-well recirculating test with a step input was used.
- The single-well injection-withdrawal test with tracer monitoring only at the single well was used.
- The tracer input was not clearly defined.
- The tracer breakthrough curve was assumed to be the superposition of breakthrough curves in separate layers.
- The measurement of tracer concentration in space was inadequate.
- The equation used to obtain dispersivity was not appropriate for the data collected.

differ (e.g., Ivanovitch and Smith, 1978; Sauty, 1977). The problem with this interpretation is that there are typically numerous heterogeneities on a small scale that cannot be attributed solely to identifiable layers. One possible explanation of the tailing in radial convergent tests could be borehole flushing -- that is, the possible tailing produced by slow flushing of the tracer out of the borehole by the ambient flow. Goblet (1982) measured the slow flushing of tracer out of the borehole and used a convolution integral solution to reproduce the tailing observed at the pumping well. In cases where borehole flushing was observed and accounted for, dispersivities obtained from a radial convergent flow test were not excluded from the high quality category.

2. The tracer input must be well defined.

Both the input concentration and the temporal distribution of the input concentrations must be measured. If not, the input is another unknown in the equation, and we are less confident in the dispersivity.

3. The tracer must be conservative.

A reactive or nonconservative tracer complicates the equations and results in additional parameters that must be estimated. Consequently, we are less confident in the resulting dispersivity. Tracers such as Cl^-, I^-, Br^-, and tritium were considered to be conservative.

4. The dimensionality of the tracer concentration measurements was appropriate.

A tracer, or solute, introduced into an aquifer will spread in three dimensions. High reliability dispersivities were judged to be those where three-dimensional monitoring was used in all cases except where the aquifer tracer had been injected and measured over the full depth of the aquifer; in this case two-dimensional monitoring was acceptable. In all other cases -- where the dimension of measurement was either not reported, or where two-dimensional or one-dimensional measurements were used where three-dimensional measurements should have been used -- the data were judged to be of lower reliability.

5. The analysis of the concentration data was appropriate.

Since the interpretation of the tracer data is necessarily linked to the type of tracer test to which the interpretation method is applied, these two features of the field studies were evaluated together. The four general categories of data interpretations can be grouped as: (1) breakthrough curve analysis, usually applied to uniform flow tests and radial flow tests (e.g, Sauty, 1980); (2) type curve matching, usually applied to uniform flow tests, radial tests, and two-well tests (e.g., Grove and Beetem, 1971; Ogata and Banks, 1961; Sauty 1977; and Gelhar, 1982); (3) method of spatial moments, uniform flow tests (Fischer, et al., 1979); and (4) numerical methods, applied to contamination events (e.g., Pinder, 1973; and Konikow and Bredehoeft, 1974). Many of the specific problems with most data interpretation methods are discussed in the sections on high reliability (point 1) and low reliability (points 1, 2, 3, 4, 6) dispersivity values. However, two pertinent points are discussed below.

One obvious difficulty with the data interpretation is that in using either breakthrough curves or type curve matching to determine dispersivity, the solution usually assumes that the dispersivity is constant. The field data collected by this literature search suggest this assumption is not valid, at least for small-scale tests (tens of meters). At larger scales (hundreds of meters) an asymptotic constant dispersivity is valid theoretically. However, at most sites the displacement distance after which the dispersivity is constant is not known. Therefore, high quality data were judged to be those for which no _a priori_ assumptions were made regarding the dispersivity.

A second major problem with many of the analyses reviewed was that a one- or two-dimensional solution to the advection-dispersion equation was frequently used when, in fact, the spreading of the plume under consideration was three-dimensional in nature. High quality dispersivities were those for which the dimensionality of the solute plume, the solute measurements, and the data analyses were consistent.

Low Reliability Dispersivity Values. The classification of a reported dispersivity as being of low reliability resulted if one of the following criteria was met.

1. The two-well recirculating test with a step input was used.

The problem with this configuration is that beyond initial times, the breakthrough curve is not strongly influenced by dispersion, but rather is determined by the different travel times along the flow paths established by injection and pumping wells. As a result, the two-well test with a step input is generally insensitive to dispersivity. For this reason all tests of this type were considered to produce data of low reliability.

2. The single-well injection-withdrawal test was used with tracer monitoring at the pumping well.

A difficulty encountered in the small-scale single-well injection-withdrawal test (where water is pumped in and out of the same well) is that if observations are made at the production well, the dispersion process observed is different than one of unidirectional flow. The problem stems from the fact that macrodispersion near the injection well is due to velocity differences associated with layered heterogenity of the hydraulic conductivity of the medium. In the single-well test with observations made at the production well, one is simply observing the effects of reversing the velocity of the water. If the tracer travels at different velocities in layers as it radiates outward, it will also travel with the same velocity pattern as it is drawn back to the production well. As a result, the mixing process is partially reversible and the dispersivity would be underestimated relative to the value for unidirectional flow. Heller (1972) has carried out experiments which demonstrate this reversibility effect, on a laboratory scale.

3. The tracer input was not clearly defined.

When a contamination event or an environmental tracer is used, the tracer input (both quantity and temporal distribution) is not well defined and hence becomes another unknown in the equation.

4. The tracer breakthrough curve was assumed to be the superposition of breakthrough curves in separate layers.

These studies generally assume the material is perfectly stratified which, especially at the field scale, may not be valid. At a small scale (a few meters) where continuous layers may be a reasonable assumption, the dispersivity of each layer does not represent the field-scale parameter. The field-scale dispersivity is a result of the spreading due to the different velocities in each layer. Therefore, modelling separate layers is not a good method to determine field-scale dispersivities.

5. The measurement of tracer concentration in space was inadequate.

   In uniform flow the tracer is often distributed in three-dimensional space, but if the measurements are two dimensional or one dimensional then the true tracer cloud cannot be modelled due to a lack of data. If the tracer is introduced over the entire saturated thickness, then two-dimensional measurements would be adequate.

6. The equation used to obtain dispersivity was not appropriate for the data collected.

   Various assumptions regarding flow and solute characteristics are made in obtaining a solution to the advection-dispersion equation. To apply a particular solution to the data from a field experiment, the assumptions in that solution must be valid for the experiment. One common example where the assumptions are not valid for the experiment is the case of a one-dimensional uniform flow solution applied to a radial flow test, because the converging (or diverging) flow field around the well is clearly nonuniform.

In this classification process very few dispersivity values were judged to be of high reliability. These few included Bonnaud, Borden, Palo Alto Baylands, University of California-Berkeley, and Yvane Region, Israel. There were 38 sites for which the data were judged to be of low reliability for one or more of the reasons discussed above; 18 sites provided data that were judged to be of intermediate reliability. Figure 2-6 depicts the longitudinal dispersivity data replotted with symbols reflecting the reliability classification; the largest symbols indicate data judged to be of highest reliability.

The general compilation of all of the dispersivity data in Figure 2-1 indicates that the dispersivity increases indefinitely with scale but, after critically evaluating the data in terms of reliability as shown in Figure 2-6, it becomes evident that this trend may be more apparent than real. The largest high-reliability dispersivity is around 3 m (Bonnaud) and the largest scale of high reliability values is on the order of 100 m (Yvane, Israel). It is not clear from the data whether dispersivity increases indefinitely with scale or whether the relationship becomes asymptotic for very large scales, as would be predicted by the stochastic theory of Gelhar and Axness (1983). This points to a need for field experiments designed to observe the displacement dependence of dispersivity at scales up to 100 m or more.

Another important point that is highlighted by this analysis is the three-dimensional nature of solute plumes. There is a need to recognize this fact and use three-dimensional solutions in determining dispersivity. All the sites which had extensive three-dimensional monitoring clearly show three-dimensional character of the plumes with very limited vertical mixing (e.g., Borden (Sudicky, et al., 1983); Cape Cod (LeBlanc, 1982); Long Island (Perlmutter and Lieber, 1970; Pinder, 1973); New Zealand (New Zealand Ministry of Works and Development, 1977); Aefligen, Switzerland (Werner, et al., 1983)). Future field experiments should be designed to observe this three-dimensional structure.

Other Transport Parameters

Although the primary emphasis of this review was the dispersive mixing characteristics, some other important physical transport parameters were also considered. Effective porosity is a key transport parameter that determines the mean rate of movement of an inert solute in an aquifer. The data summarized in Table 2-1 show that effective porosity can vary over a wide range. For example, values of effective porosity for alluvial sand and gravel aquifers ranging from 20 to 40 percent have been used in modelling. Frequently, values near the upper end of this range (Pinder, 1973; Robson, 1978) or the lower end of this range (New Zealand Ministry of Works and Development, 1977; Konikow and Bredehoeft, 1974) are found, but in many cases the values seem to be arbitrarily chosen. This uncertainty in effective porosity means that predictions of mean contaminant movement can involve large errors. There is a need to develop methods of measuring in-situ effective porosity at the field scale. Field experiments should be designed to evaluate methods of determining field-scale effective porosity. Specifically, it should be determined to what extent small-scale tracer tests (on the order of a few meters) can be used to predict large-scale effective porosity.

Molecular diffusion may affect the contaminant transport process in an aquifer. The dispersion coefficient is usually expressed as the sum of molecular diffusion and mechanical dispersion due to the local velocity variations in the porous medium. It is usually assumed on a field scale that the effects of molecular diffusion on the dispersion coefficient are negligible compared to mechanical dispersion, so that field dispersion is solely a function of the product of the dispersivity and the mean velocity. The experiments at the Bonnaud site provide some field evidence that this is the case. In tests reported by Sauty (1977) and Sauty (1980) hot water and the chemical tracer ^{131}I each provided identical dispersivities at the same distance, using the same test. Because the effective thermal diffusivity is several orders of magnitude larger than diffusion coefficients for chemical solutes, these results support the assumption that molecular diffusion does not affect the field-scale dispersion coefficient.

None of the field tests summarized in Table 2-1 were specifically designed to evaluate the effect of molecular diffusion, and we have not detected any effects in the observations which can be definitely attributed to molecular diffusion. On the other hand, some studies have attributed an important role to molecular diffusion. Sudicky, et al. (1983) interpret molecular diffusion as being the dominant mechanism for vertical dispersion, although our reanalysis of the data does not support this interpretation. Future field experiments should be designed to evaluate the effects of molecular diffusion. This can be done by simultaneously using tracers with distinctly different molecular diffusion coefficients.

The fluid properties viscosity and density may be significantly affected by solutes at high concentrations as well as by temperature. None of the field studies explicitly consider effects of varying density or viscosity. Some field observations for landfill plumes (Kimmel and Braid, 1974; Sudicky and Cherry, 1979) seem to indicate that the density excess associated with the solute concentrations on the order of 10^4 ppm causes the plumes to sink, but this effect has not been investigated quantitatively. Rather, numerical modelling of the landfill plume at the Borden site ignored this effect (Sykes, et al., 1982). It can be expected that density and viscosity changes will also affect dispersive mixing. A stable density stratification is likely to suppress vertical mixing (Dagan, 1966). Enhanced mixing can be expected when a lower viscosity fluid displaces one of higher viscosity. This viscous instability effect is discussed in the review by Wooding and Morel-Seytoux (1976). However, the classical treatments of viscous and/or gravitational instability consider only homogeneous porous media and as such are of little value in quantifying field-scale phenomena. Ongoing theoretical research at MIT sponsored by the National Science Foundation is exploring the effects of density and viscosity differences in heterogeneous media using stochastic methods. Many solutes that are of environmental concern occur at low concentrations that will not produce significant density or viscosity changes. Future field experiments should be designed to minimize the effects of density and viscosity changes.

CONCLUSIONS AND RECOMMENDATIONS

The results of the general review of field-scale dispersion data show that longitudinal dispersivities range from 1 cm to 5 km over scales from 0.75 m to 100 km; the dispersivity appears to increase indefinitely with scale (Figure 2-1). After assessing the reliability of the dispersivity data (Figure 2-6), the trend of increasing dispersivity with scale is much less definitive. These results point to the need for field experiments designed to observe the displacement dependence of dispersivities at scales of at least 100 m.

The detailed analysis for the Borden site, as well as the general results of several sites with three-dimensional monitoring, clearly demonstrate the three-dimensionality of plumes and the very small vertical transverse dispersivity which is common to stratified sedimentary deposits. Field experiments should be designed to monitor the three-dimensional structure of the tracer plume.

The field-scale effective porosity, an important parameter determining the mean rate of movement of an inert solute, was found to vary over a wide range, even for similar materials (e.g., 20 to 40 percent for sand and gravel). Field experiments should be designed to evaluate methods of determining field-scale effective porosity from small-scale tracer tests. No systematic effects of molecular diffusion on dispersive transport were detected in the data reviewed. For very heterogeneous media with a significant amount of fine-grained material, some influence on molecular diffusion can be anticipated. Field experiments should be designed to evaluate the effect of molecular diffusion by using tracers with distinctly different molecular diffusion coefficients.

The three-dimensional stochastic theory of Gelhar and Axness (1983) provides a general predictive framework relating the macrodispersivity tensor to the spatial statistical properties of hydraulic conductivity. Field experiments should be designed to measure the spatial variability of hydraulic conductivity and evaluate the predictive capabilities of the stochastic theory.

Section 3

UNSATURATED TRANSPORT PROCESSES

This portion of the review emphasizes field-scale physical processes which affect the transport of inert solutes through unsaturated natural porous earth materials. Laboratory-scale phenomena are not emphasized because these are covered in textbooks, (e.g., Hillel, 1980) and review articles (Arnold, et al., 1982), and are not applicable to quantify field-scale behavior. We emphasize the classical case of isothermal liquid water flow in ideal porous media whose physical properties are not altered by water or solutes. Several non-classical effects are discussed by Nielsen and Biggar (1982). Numerical models also are not emphasized because these are reviewed by Oster (1982).

Although numerical modeling of unsaturated flow and transport at field scales of tens or hundreds of meters is attempted, there is no evidence that the dominant unsaturated flow and transport processes in such field settings are presently quantifiable. Some theories and such models exist, but their validity and reliability has not been demonstrated (Arnold, et al., 1982, pp. 340-342). A serious obstacle to the application of field-scale flow and transport models is the difficulty of determining the necessary parameters in view of the extreme heterogeneity of natural media. Furthermore, the proposed theories and models have not been tested adequately with field data at scales greater than a few meters. Even though the existing field information is not extensive, an analysis and review of these observations is important to identify some fundamental mechanisms of field-scale flow and transport in unsaturated zones to better direct future experiments.

The objectives of this review are threefold. The first is to summarize the existing field observations which are presently scattered in a wide variety of publications. The second objective is to interpret some of the existing field observations to determine, when possible, the controlling factors of large-scale unsaturated flow and transport. The review emphasizes effects of natural soil heterogeneity on field-scale flow and transport processes, and on possible effective large-scale parameters. The third objective is to draw conclusions and make recommendations for future experiments, based on the results derived under the first two objectives.

SUMMARY OF FIELD OBSERVATIONS

The summary of existing field observations on unsaturated flow and transport is given in the form of three tables. Table 3-1 offers a brief description of the observations in terms of the scale (spatial extent of field experiments) of observation (both lateral and vertical scales), the soil type, the tracers, the source of liquid, and the type of observation (controlled versus uncontrolled). Table 3-2 summarizes the sampling methods used to obtain measurements of soil moisture content (θ), capillary tension head (ψ), solute concentration (c), etc. Table 3-3 includes some points of interest pertinent to the observations, including the methods of analysis that were used and the conclusions that were drawn.

Table 3-1 shows that most of the controlled experiments are limited to small-scale displacements on the order of 1-2 m. Very few large-scale observations exist, all of them obtained from uncontrolled experiments. The observations of Trautwein, et al. (1983) involve the largest scale (with depth of approximately 100 m). The soils vary from relatively homogeneous (agricultural soils), to extremely heterogeneous (e.g., stratified sediments). The most often used tracers in the controlled experiments are ^3H, Cl^-, NO_3^- and Br^- .

Table 3-2 shows that soil moisture content is usually measured by neutron probes, while ψ is usually measured by tensiometers. Solute concentrations are obtained by extraction and analysis of water samples using soil-moisture suction cups. In a few cases gamma logs or electrical conductivity measurements, using salinity sensors, were obtained.

Table 3-3 indicates that particular types of flow (such as lateral movement or rapid gravity flow) have on certain occasions been observed. As for analysis of data, Table 3-3 shows that a simple one-dimensional model is usually adopted for estimation of dispersivities. Table 3-3 is useful but very brief; some field observations which deserve more attention are indicated with a star (*) and will be discussed in greater detail in the next section.

DISCUSSION OF FIELD OBSERVATIONS

It was mentioned in the Introduction that unsaturated flow and transport is a highly complicated problem. Two of the factors that complicate the problem are its dimensionality (three-dimensional) and the fact that soil properties vary highly in space. Existing large-scale field observations are very limited and often have not been analyzed thoroughly. The objective of this section is to examine and discuss

Table 3-1

DESCRIPTION OF FIELD OBSERVATIONS FOR THE UNSATURATED ZONE

Author	Vertical Scale (m)	Lateral Scale	Soil Type	Tracer	Source of Liquid	Type of Observation
Andersen, et al. (1968)	22.0	Wells located 15 m apart	Fine to medium sand and gravels	Environmental ^3H	Rainfall	Uncontrolled
Biggar and Nielsen, (1976)	1.83	20,6.5x6.5m plots on a 150ha field	Panoche clay loam	Cl^- NO_3^-	Irrigation	Controlled
Brissaud, et al. (1983)	1.60	0.50 m	Sandy loam	D_2O	Irrigation	Controlled
Crosby, et al. (1968)	20.0	10.0 m	Stratified glacial outwash deposits	Cl^- NO_3^-	Leakage from a drain field	Real contamination site
DePoorter, et al. (LASL)	6.0	3.0 m	Crushed tuff, sand and gravel		Irrigation	Caisson controlled
Foster, et al. (1980)	20.0	Boreholes at max 10km apart	English chalk	Environmental ^3H	Rainfall	Uncontrolled
Isaacson, et al. (1974)	70.0	Wells located 15 m apart	Silty sand	Environmental ^3H	Rainfall	Uncontrolled
Johnson, et al. (1981)	10.0		Silty sand, coarse sand and gravel	Cl^-, etc.	Leakage from sanitary landfills	Real contamination site

Table 3-1 (continued)

DESCRIPTION OF FIELD OBSERVATIONS FOR THE UNSATURATED ZONE

Author	Vertical Scale (m)	Lateral Scale	Soil Type	Tracer	Source of Liquid	Type of Observation
Johnston, et al. (1983)	8.0	4.4x2.3m plot	Weathered doleritic and granitic profile, sandy topsoil	Cl^- Rhodamine	Irrigation	Controlled
Jury, et al. (1982)	3.05	0.64 ha field	Tujunga loamy sand	Br^-	Irrigation and rainfall	Controlled tracer, Natural rainfall
Kies, (1981)	2.00	7x7 m plot	Glendale clay loam	Cl^- ^3H NO_3	Irrigation	Controlled
Knoll and Nelson, (1962)	3.0	7 m	Relatively homogeneous fine sand	Sr^{85} NO_3^-	Irrigation	Controlled
Mann, (1976)	7.0	3x3 m plots	Alluvial deposits	--	Irrigation	Controlled
Oakes, et al. (1977)	20.0		English chalk	Environmental ^3H NO_3	Rainfall	Uncontrolled

Table 3-1 (continued)

DESCRIPTION OF FIELD OBSERVATIONS FOR THE UNSATURATED ZONE

Author	Vertical Scale (m)	Lateral Scale	Soil Type	Tracer	Source of Liquid	Type of Observation
Phillips, et al. (1979); Jones, et al. (1983)	7.5	0.6 m	Homo-genized sand	Cobalt 60/EDTA ^3H	Irrigation	Controlled Caisson experi-ment
Price, et al. (1979)	30.0	80 m	Glacial flu-vial deposits	Pu Am	Leakage from a disposal crib	Real con-tamination site
Prill, et al. (1977)	10.0	20 m	Layers of sand & gravel clay, silty clay and loam	—	Irrigation	Controlled
Purtymum, et al. (1973)	20.0	15 m	Ashflows of rhyolite tuff	^3H	Wastes from storage shafts	Real con-tamina-tion site
Quisenberry and Phillips (1976)	0.90	2.13x 2.13m plots	I.Maury silty loam II.Hunting-ton silty clay loam	Cl^-	Irrigation	Controlled
Routson, et al. (1979)	25.0	50 m	Stratified glaciofluvial deposits	^{106}Ru ^{137}Cs ^{144}Ce	Leakage from radioactive waste stor-age tank	Real con-tamina-tion site

Table 3-1 (continued)

DESCRIPTION OF FIELD OBSERVATIONS FOR THE UNSATURATED ZONE

Author	Vertical Scale (m)	Lateral Scale	Soil Type	Tracer	Source of Liquid	Type of Observation
Saffina, et al. (1979)	1.50	I.0.85 ha field II.Lysimeters	Plainfield loamy sand	Cl^- NO_3^-	Irrigation	Controlled
Smajstrla, et al. (1976)	1.80	2,2x2 m plots	I.Lakeland fine sand II.Derby loam	Cl^-	Irrigation	Controlled
Starr, et al. (1978)	3.00	I.1.8 m diam. cylinder II.4,4.6x6.1m plots	Fine sandy loam over coarse sand	I.dye soln. II. Cl^-	Ponded irrigation	Controlled
Supkow, (1974)	6.0	15.0 m		No tracer	Irrigation	Controlled
Trautwein, et al. (1983)	120.0	Size of pond 500 m x 1000 m	Layers of sand and clay	Organic carbon metals	Leakage from a waste disposal pond	Real contamination site
Van de Pol, et al. (1977)	1.52	8x8 m plot	Layers of clay, silty clay, silty loam & sand	Cl^- ^3H	Irrigation	Controlled
Warrick, et al. (1971)	1.80	6.1x6.1 m plot	Panoche clay loam	Cl^-	Irrigation	Controlled
Wellings and Bell, (1980)	3.0		Fine sandy loam over fine sand	Cl^- NO_3^-	Rainfall	Controlled tracer, natural rainfall

Table 3-2

DATA ACQUISITION TECHNIQUES FOR UNSATURATED ZONE PARAMETERS

Author	Soil Moisture θ	Capillary Tension head Ψ	Tracer Concentration C	Temperature T
Andersen, et al. (1968)	Neutron probe		Laboratory analysis of soil samples	
Biggar and Nielsen, (1976)		Tensiometers	Ceramic suction cups	
Brissaud, et al. (1983)	Neutron probe		Gramma neutron probe	
Crosby III, et al. (1968)	• Laboratory analysis of soil samples • Geophysical borehole logging		Laboratory analysis of soil samples	
DePorter, et al. (LASL)	Neutron probe	Tensiometers	Ceramic suction cups	Thermocouples
Foster, et al. (1980)	Analysis of soil samples		Laboratory analysis of soil samples	
Isaacson, et al. (1974)	Laboratory analysis of soil samples	Thermocouple Psychrometers	Laboratory analysis of soil samples	Diode temperature transducers
Johnson, et al. (1974)			Pressure-vacuum lysimeters	
Johnston, et al. (1983)	Laboratory analysis of soil samples		Soil samples, visual inspection of Rhodamine	
Jury, et al. (1982)	Neutron probe	Tensiometers	Suction probes	
Kies, (1981)	Neutron probe	Tensiometers	Ceramic suction cups	
Knoll and Nelson, (1962)	Neutron probe			

Table 3-2 (continued)

DATA ACQUISITION TECHNIQUES FOR UNSATURATED ZONE PARAMETERS

Author	Θ	Ψ	C	T
Oakes, et al. (1977)	Laboratory analysis of soil samples		Laboratory analysis of soil samples	
Mann, (1977)	Neutron probe			
Phillips, et al. (1979) and Jones, et al. (1983)	Neutron probe	• Heat dissipation units • Thermocouple psychrometers • Electrical resistance units • Tensiometers	• Radiological counting systems • Laboratory analysis of samples	• Thermistors • Thermocouples
Prill, et al. (1977)	Neutron probe			
Price, et al. (1979)			• Portable radiation survey instruments • Laboratory analysis of soil samples	
Supkow, (1974)	Neutron probe			
Purtymum, et al. (1973)	Laboratory analysis of samples		Laboratory analysis of soil samples	
Quinsenberry and Phillips, (1976)	Laboratory analysis of soil samples		Laboratory analysis of soil samples	

Table 3-2 (continued)

DATA ACQUISITION TECHNIQUES FOR UNSATURATED ZONE PARAMETERS

Author	θ	Ψ	C	T
Routson, et al. (1979)			• Gamma scintillation probe • Laboratory gamma energy analysis (GEA) of sediment samples	
Saffina, et al. (1979)			Soil sampling excavation of profile	
Smajstrla, et al. (1976)	Neutron probe	Tensiometers	Electric conductivity soil salinity sensors	
Starr, et al. (1978)	Neutron probe	Tensiometers	Ceramic suction cups	
Trautwein, et al. (1983)	Laboratory analysis of soil samples		Lysimeters and monitor wells sample water in unsaturated zones	
Van de Pol, et al. (1977)	Neutron probe	Tensiometers	Ceramic suction cups	
Warrick, et al. (1971)		Tensiometers	Ceramic suction cups	
Wellings and Bell, (1979)	Neutron probe	• Mercury manometer tensiometers • Gypsum resistance blocks • Pressure transducer tensiometers	Laboratory analysis of soil samples	

Table 3-3

REMARKS ON UNSATURATED ZONE PARAMETERS

Author	Remarks on Observations in the Unsaturated Zone
*Andersen, et al. (1967)	Soil moisture profiles show a downward movement of the moisture front 3-3.5 m/mo. The true traveling velocity of water, determined by ^3H, is 4.5 m/year. The apparent velocity of water is due to a pressure wave. A displacement with dispersion model fits ^3H data quite well.
Brissaud, et al. (1983)	Steady-state and transient experiments were performed. Dispersivities were evaluated by approximately fitting model breakthrough curves to experimental ones. Concentration measurements indicate lateral flow is significant below approximately 1 m.
*Biggar and Nielsen, (1976)	Fitting dispersion coefficient (D) and apparent velocity (v_s) of a one-dimensional model to breakthrough curves at each location and depth. Statistical analysis of D and v_s shows that they are log-normally distributed. D is approximately linearly related to v_s.
*Crosby, et al. (1968)	Despite substantial additions of water in the drain field of the nursing home, practically dry alluvium was found at depths 40 m or less. This suggests that water must be flowing laterally in the finest beds and is removed by evapotranspiration.
DePorter, et al. (LASL)	Caisson experiments. Soil is excavated, the caissons emplaced and backfilled with sandy material. The caissons are instrumented with horizontal neutron probe access tubes, tensiometers, porous cups and thermocouples.
Foster, et al. (1980)	It is suggested that up to 20 percent of environmental ^3H is transported to groundwater by a preferential (more rapid) flow through fissures. The bacteriological contamination of groundwater supplies after high intensity rainfalls is an evidence of such mechanisms.
Isaacson, et al., (1974)	^3H concentration decreases logarithmically from the surface to a depth of 4.5 m. Archaic water is found below 7 m. This suggests that precipitation water does not percolate to the water table. It is suggested that temperature gradients are of importance in arid regions.

Table 3-3 (continued)

REMARKS ON UNSATURATED ZONE PARAMETERS

Author	Remarks on Observations in the Unsaturated Zone
Johnson, et al. (1981)	Tracer data beneath landfills indicate alternation of leachate in the unsaturated zone with depth. Lateral moisture flow is speculated.
*Johnston, et al. (1983)	Movement of water through preferential paths (roots) is observed. Water moves laterally more in the finer-textured dolerite saprolite than in the granite saprolite. Preferential flow is observed on the granite saprolite only.
*Jury, et al. (1982)	Average concentration breakthrough curves indicate displacement with dispersion. A transfer function model is fitted to the data.
*Kies, (1981)	Fitting dispersion coefficient (D) and apparent velocity (v_s) of a one-dimensional model to breakthrough curves at each location and depth. D, v_s are log-normally distributed. D is related to v_s approximately linearly. In some probes tracers reach deeper depths before arriving at shallower depths.
*Knoll and Nelson, (1962)	Soil moisture maps show lateral movement of water. Lateral movement is more pronounced at approximately 2 m depth, probably due to the existence of a finer-textured layer at this depth.
Mann, (1976)	Field tests in relatively homogeneous dry vadose zones indicate a downward movement of water with only minor lateral movement. Before water application, moisture in the vadose zone is below field capacity, after drainage moisture remains at field capacity.
Oakes, et al. (1977)	It is suggested that infiltration takes place at large fissures and that solutes diffuse between this moving water and the relatively static water in the chalk matrix. A mathematical model is proposed that reproduces observed distributions of ^3H.
Phillips, et al. (1979) Jones, et al. (1983)	Caisson experiments. Soil is excavated, the caissons emplaced and back filled with uniform sand. In long term tests cobalt dissociates from EDTA and is sorbed onto the soil. ^3H is transported approximately 4 m.

Table 3-3 (continued)

REMARKS ON UNSATURATED ZONE PARAMETERS

Author	Remarks on Observations in the Unsaturated Zone
*Prill, et al. (1977)	In infiltration pond experiments, insignificant lateral movement of soil moisture was observed. Soil is stratified, initial soil moisture is relatively high and size of pond is large.
*Price, et al. (1979)	Actinide distribution data show lateral spread of waste liquid in the stratified sedimentary units below the crib. Waste liquid is more prone to move laterally in a medium to very fine sand unit than to move deeper to a coarse sand unit.
Purtymum, et al. (1973)	Soil moisture and ^{3}H data indicate lateral movement. Lateral movement is amplified in the contact between two ashflows. The low values of moisture content indicate moisture transport in the vapor phase.
Quisenberry and Phillips, (1976)	Rapid movement of water is observed, due to macropores. This movement depends on the initial moisture content. For soil initially dry, little of the applied water moves past the 90 cm depth; if soil is relatively wet a large percentage percolates past 90 cm depth.
*Routson, et al. (1979)	Data on radioactive pollutant concentrations between 1973-1978 indicate lateral spread of soil moisture. Soil is highly stratified, which enhances lateral spread. Waste movement is initially rapid, but it decreases with time.
Saffina, et al. (1979)	Sprinkler-irrigation field and lysimeter experiments. Large localized variability in solute distributions beneath and between plants was observed. At least part of this variability must be due to natural heterogeneity.
Smajstrla, et al.	Use of ceramic and platinium electrode salinity sensors is described. Solution concentrations with depth are given but no analysis or discussion of data is offered.
*Starr, et al. (1978)	Soil is layered with a fine-textured layer overlying a coarse-textured one. Both lysimeter and plot data indicate that water in the coarse layer moves rapidly in small fingers of flow. Frontal type of flow may not be valid in such soils.

Table 3-3 (continued)

REMARKS ON UNSATURATED ZONE PARAMETERS

Author	Remarks on Observations in the Unsaturated Zone
Supkow, 1974)	Soil moisture distributions in the unsaturated zone show lateral flow. Speculation is that moisture movement is by piston flow, while some mixing occurs near the contact of displacing and displaced waters.
*Trautwein, et al. (1983)	20 years after pond construction, pond water is found at depths at 100 m below pond. The thick sand layer below 100 m seems to stop vertical migration. Effective K_{sat} obtained by fitting a one-dimensional flow model are much greater than laboratory values.
*Van de Pol, et al. (1977)	Fitting dispersion coefficient and apparent velocity of a one-dimensional model to breakthrough curves at each location and depth. D, v_s are log-normally distributed. Some breakthrough curves indicate that solute arrives at deeper locations before arriving at shallower ones.
*Warrick, et al. (1971)	v_s is calculated from irrigation and soil moisture data. D is fitted to the maximum of the solute versus depth curves. Calculated distributions do not penetrate soil as much as the observed. This implies a large solute velocity probably due to lack of complete mixing.
*Wellings and Bell, (1980)	A piston flow model is assumed and it is postulated that it can reasonably predict the movement of solutes. However, the information given to support the model does not seem to be sufficient.

* Discussed in detail in text.

some field observations in an attempt to better understand field-scale flow and transport mechanisms. It is of particular interest to investigate the effects of soil heterogeneity.

The discussion consists of two parts. First, observations offering a physical insight into the mechanisms of field-scale flow and transport are discussed. Second, publications addressing the practical problem of modelling such systems and estimating field-scale parameters are discussed. Both aspects of the problem (physical insight and modelling) are closely related.

Mechanisms of Field-Scale Flow and Transport

Hydraulic properties of soils vary highly in space, even in what appear to be relatively homogeneous soils (Nielsen, et al., 1973). Thus, in many field-scale problems, a large degree of variation of hydraulic properties can be expected. Water in unsaturated soils moves under gravity and surface forces. Because of the variation of the hydraulic properties of the soil, the soil moisture content, etc., the direction of the total force, and thus the direction of the flow, is spatially variable. As a result, flow and transport in unsaturated soils is generally a three-dimensional process, as is the case for the saturated zone. Depending on the type of soil variability (e.g., layering, macropores) and/or initial and boundary conditions, particular types of flow may be more pronounced. Two such types of flow, of particular interest in waste disposal control, are lateral flow and fast gravity flow. Several field observations are examined in this section to investigate under what conditions these types of flow occur.

Lateral Flow. Theoretical analysis of field-scale unsaturated flow (Yeh and Gelhar, 1983; Yeh, et al., 1982) indicates that in stratified soils the effective hydraulic conductivity tensor is anisotropic with a tension-dependent degree of anisotropy. The ratio of the horizontal hydraulic conductivity to the vertical hydraulic conductivity increases with tension. One thus expects that at high tensions (dry soils) lateral flow may be important.

Routson, et al. (1979) investigated the time history of leakage below a radioactive waste storage tank at Hanford, Washington. The soil formation consists of glaciofluvial deposits with the principal units consisting of pebbly and medium sand. The deposits are bedded, and sharp boundaries often exist between sediment types. Bedding consists of thin, nearly horizontal, discontinuous laminations and cross-stratified sedimentary units. The climate where the tank is located is arid and the sediments have a relatively low soil-moisture content. In 1973 a leak from a storage tank was observed. Measurements of unsaturated zone contamination, using gamma radiation logs in wells around the tank, indicated significant lateral movement of wastes in the sediment layers (at least at the initial stages of the leak, between 1973-1974). At later stages of the leak, flow was so slow that lateral movement could not be detected, given that the radioactive decay of tracers is relatively rapid (106 Ru has a half-life of approximately one year). Plotted isopleths of 106 Ru (the main radioactive component of the waste liquid) also show lateral movement. Lateral spread is much larger than the diameter of the tank, while vertical movement is restricted to a few tens of meters. The report assumed that the lateral spread was due to unsaturated flow and sediment layering but no physical explanation that leads to such phenomena was given. Lateral movement is probably due to the relatively high tensions occurring in such dry materials. Horizontal stratification enhances such movement, since at high tension, hydraulic conductivites of fine textured materials are relatively high; and water may prefer to spread laterally in fine beds than to move vertically through coarser ones (see Yeh, et al., 1982). Unfortunately, the report does not give the stratigraphy of the sedimentary units so a direct comparison between unit type and lateral spread is not possible. Several small-scale experiments were conducted by Routson, et al. and showed significant lateral spread in such layered soils.

Price, et al. (1979) reported the movement of leachates in the unsaturated zone below a waste disposal crib at Hanford. The sedimentary units below the crib are stratified and consist of layers of medium to very fine sand; pebbly very coarse to medium sand; and sandy silt. The crib is located in an arid environment and the initial soil moisture of the sediments is relatively low. Sediment samples were analyzed for radioactivity, and isopleths of Pu and Am were plotted. These data show a lateral movement of wastes in the unsaturated zone below the crib extending to a width of 10 m, encompassing the crib perimeter. The waste liquid was more prone to spread laterally in the medium to very fine sand unit than to move deeper into the pebbly, very coarse to medium sand unit. This behavior can be explained by the fact that in dry soils with high tensions, hydraulic conductivity of fine materials is larger than the hydraulic conductivity of coarse ones.

Crosby, et al. (1968, 1971) discussed measurements of soil moisture and pollutants below a septic tank drain field area in the Spokane Valley in Washington. The sediments below the drain field consist of glacial outwash deposits and are probably highly stratified. This environment is also arid. Moisture data in the unsaturated zone below the drain field indicated unexpectedly high tension conditions below a depth of 7-10 m. Under such tension, gravitational movement of water cannot be expected. Since water is continously added in the drain field, the law of conservation of mass requires an accounting for the lost moisture. The authors assumed that water must be moving laterally away from the drain field and removed to the atmosphere by evapotranspiration. A series of laboratory experiments conducted to investigate the possibility of lateral movement were discussed in Crosby et al., (1968). These experiments indicated that lateral movement may be important under certain soil conditions. Fine, medium, and coarse sands were bedded in a sand box model. Water was added to a square inch surface area. Under high water application rates, water essentially moved as saturated masses or ribbons to the bottom of the model. Under low water application rates, however, capillary dispersion of water in the finest layers was able to keep pace with water additions before the lateral boundaries of the model affected the flow. Crosby et al. concluded that the nature of interbedded materials and rate of water application are important factors controlling wastewater movement in unsaturated zones.

Prill (1977) discussed moisture movement in the unsaturated zone below four artificial groundwater recharge ponds. The alluvial deposits below the ponds consist of layers of sand and gravel interbedded with clay, silty clay, and loam layers. The ponds are circular with 15 m diameters. Before the start of ponding the soil moisture content in the sediment below the pond was relatively high (with a 70 percent degree of saturation). Measurements of moisture content beneath and around the pond (to a depth of 10 m) indicated vertical movement of the moisture front but no significant lateral spread. Wetting front patterns suggested that a major part of the applied water (estimated to be around 90 percent) moved downward beneath the pond. Lateral movement was very slow and was restricted to a short distance even in the finer texture layers. The implication of this experiment is that when the initial soil moisture is high, the water application rate is rapid, and the application area is large relative to the depth of observed unsaturated zone, lateral spread may not be important even in stratified soils. In such cases gravitational forces may be more important than capillary forces.

Knoll and Nelson (1962) described soil moisture movement beneath a six-inch square crib. The soils of the study area consisted of a relatively homogeneous fine sand, except for some thin irregularly placed lenses of a material of slightly different porosity. The soil matrix was initially relatively dry. Water application was controlled so that ponding was always maintained in the crib. The lateral spread of soil moisture was quite significant and was more pronounced at the 2 m depth. The authors suggested that this was probably due to a lens of a slightly more permeable material. This experiment indicated that, for initially relatively dry soil and small size of application area relative to the observed depth of unsaturated zone, lateral spread can be significant.

Palmquist and Johnson (1962) described a laboratory experiment which contains some pertinent observations. A tank model was filled with glass beads forming a porous matrix. Water was applied through a small crib at the upper surface of the porous matrix. The first model consisted of initially dry glass beads of 0.47 mm diameter (corresponding to medium sand size). After water application started, the wetted area was confined to a relatively narrow vertical column. A second model consisted of three layers of 0.036 mm diameter glass beads (silt size) separated by two layers of 0.47 mm beads. After water application started, water in the small-diameter bead layer initially moved away from the crib at nearly equal vertical and horizontal velocities, until it reached the top of the higher diameter bead layer. Water then tended to move laterally in the fine bead layer instead of moving into the coarser layer. After pressure built up, water eventually moved in the coarser layer in discrete fingers of small diameter.

The field observations discussed above show that lateral spreading of soil moisture in field settings is important in certain cases and indicate some factors enhancing such lateral spreading. These factors and the implications of lateral flow in waste disposal control will be discussed in some detail in the Conclusions and Recommendations part of this review.

Fast Gravitational Flow. In particular types of soils and under certain moisture conditions water appears to move rapidly under the influence of gravitational forces in saturated channels of relatively small diameter. Obviously such flow is of great concern in waste disposal control since wastes would reach the water table much faster than predicted from ordinary unsaturated flow models. Two types of such fast gravitational flow may be distinguished. The first type is fast vertical flow in discrete fingers in the coarse layers of stratified soils. The second type is preferential flow in soil structures such as macropores and fissures.

Starr, et al. (1978) provided experimental evidence for the existence of fast flow moving in discrete fingers in a coarse layer of a stratified soil. The soil consisted of a 60 cm layer of sandy loam over a layer of coarse sand. Two experiments were performed. In the first, a 1.8 m-diameter steel cylinder was driven into the soil to a depth of 3.6 m. A depth of 45 cm of water containing a green dye was infiltrated into the column. After infiltration ended, successive layers of soil were removed from the cylinder and the dye pattern of each newly exposed surface was photographed. In layers near the soil surface, a general green hue was observed over the cross-sectional area of the column. However, at cross-sectional areas below 1 m, 12 discrete fingers of flow were observed. These fingers ranged from 5 to 20 cm in diameter and occupied only 5 percent of the total cross-sectional area of the cylinder. These observations suggest that ponded flow of water through this layered, fine over coarse, soil moves in discrete three-dimensional fingers in the coarse subsoil, rather than as a one-dimensional front. The locations of such fingers may be controlled by natural heterogeneities of the soil, i.e., they may tend to occur in the coarser regions which offer less resistance to flow. Simple calculations show the flow in such fingers is very rapid. In a second experiment the solute movement under four 4.6 m by 6.1 m ponded plots adjacent to the first experiment, was studied. It was observed that several salt pulses reached depths of 120 cm and 180 cm very soon after they had reached the 60 cm depth. Such rapid movement may be due to fast flow in the coarse layer in discrete fingers of flow similar to the ones observed in the first experiment. The paper concludes that water moving through layered field soils may move rapidly in fingers of flow through coarse subsoils. The assumption of one-dimensional front-type flow under these field conditions may lead to gross errors if it is used to estimate the time the solutes arrive at the water table. A similar type of flow moving in discrete fingers in coarse layers of stratified soils was observed in the laboratory experiments discussed in Palmquist and Johnson (1962).

Quisenberry and Phillips (1976) studied percolation of rapidly surface-applied water in field soils with strong structure. Several 2.13 m by 2.13 m plots were established in two types of soil, Maury silty loam and Huntington silty clay loam. The experiments show that water movement through the profile was characterized by an initial rapid movement of water into the soil and subsequent movement to a depth depending, to a large extent, on the initial water content. The relative increase of water content throughout the profile corresponded very well with the increase in concentration of chloride used as a tracer. This suggests that water moved through the profile without displacing much of the initial water. It was assumed that water moved primarily in soil structures called macropores. The amount of displacement that occurred depended on the rate of water movement in the macropores as compared to the soil matrix.

Johnston, et al. (1983) reported preferential flow in pipe-like channels associated with root channels, in lateritic profiles in Western Australia. A 4.4 m by 2.3 m plot was established and a tracer solution, containing Rhodamine WT and NaCl, replaced natural rainfall in the plot. The soil profile underlying the plot consisted of a humus-rich sandy topsoil grading into sandy gravels that, in turn, overlay a weathering profile, developed on both granite and doleritic parent rocks, granitic on the west and doleritic on the east side of the plot. Granite saprolite was coarser grained than the dolerite, and included deep descending roots. Bright Rhodamine WT staining was found around the root channels of the granite saprolite, indicating preferential flow of water in these channels. The tracer solution also moved laterally from primary vertical channels in lesser soil structures. The tracer solution moved deeper in the coarser grained granite saprolite, but lateral dispersion was more pronounced in the finer grained dolerite. The lack of evidence of vertical flow in the dolerite is most likely associated with its finer texture. Large continuous voids observed in the dolerite saprolite probably do not conduct water, since these voids would only transmit water if their uppermost extension intersected a saturated layer. The paper concludes that the physical and hydraulic properties of both the clay rich matrix and the more permeable inclusions, and the areal frequency of preferential flow paths and their connection with an overlying source of free water are all of importance in the transport of solutes through these natural materials.

Thomas and Phillips (1979) also discussed the significance of water movement in macropores. They concluded that water movement in macropores is influenced by the rate of water addition, soil structure, relative sizes of pores, clay orientation, soil water content, and tillage. They further discussed the most important consequences of water movement in such macropores. They suggested that recharge of groundwater can begin long before soil reaches field capacity. Also some of the salts in the surface of a soil will be moved to a much greater depth by rain or irrigation, and because of this it is not likely that water will carry a surge of contaminants to groundwater at some time predictable by Darcian flow theory.

Bouma (1981) and Beven and Germann (1982) provided further discussion and evidence of the significance of macropores in vadose zone flow and transport. These papers suggest that rapid flow through macropores depends on moisture content and rate of water application. Large continuous voids will be filled and conduct water only at suctions close to zero (i.e., near saturation). In low moisture conditions these voids are empty and do not contribute to water flow.

The above papers show that fast gravity flow may, in certain cases, be significant. The conditions under which fast flow is important and the importance and implications of such flow on waste disposal control will be discussed in the Conclusions and Recommendations.

Dimensionality of Flow. The above discussion has indicated that flow in stratified and structured soils is three-dimensional and velocity is variable in space. In fact, three-dimensionality of flow and spatial variability of velocity occur in relatively homogeneous soils as well. Some field observations illustrate this point.

Van de Pol, et al. (1977) examined breakthrough curves at several depths in three locations established within an 8 m by 8 m plot. The soil in the top 0.7 m consisted of clay to silty clay. Examination of the breakthrough curves showed some peculiarities. Chloride breakthrough curves appeared at approximately the same time at depths 0.335 m and 0.635 m, while chloride distribution for the 0.46 m depth was considerably delayed.

All three points were on the same vertical line. This phenomenon is probably due to the three-dimensional spatial variability of the soil properties. Because of such variability, flow follows more permeable paths and tends to avoid low permeability zones (at a given moisture content). This results in spatially varying velocities and a three-dimensional flow field.

Starr, et al. (1978), discussed earlier, indicated a similar phenomenon. In the experiments under the ponded 4.6 m by 6.1 m plots, a large scatter of chloride concentration data was observed at all depths. They also observed that at some locations the solute breakthrough curve arrived at greater depths before arriving at shallower ones. The paper suggests a one-dimensional flow in the top 0.60 m sandy loam layer and flow in discrete fingers in the coarse layer below. However, the scatter of concentration data, even in the top 0.60 m layer, indicates that flow in this layer is probably three-dimensional due to spatial variability of soil properties within the layer. It is also possible that flow in the coarse layer below may be moving in channels, not because of flow instability, but because of natural heterogeneity, i.e., the flow is following paths of higher hydraulic conductivity at the existing moisture conditions.

In a similar experiment described by Kies (1981) a large degree of variability in breakthrough curves was also observed. It was also observed that in some locations breakthrough curves arrived earlier at deeper locations than at shallow ones. This also suggests a nonuniform velocity field and three-dimensionality of flow.

Modelling Approaches and Parameter Estimation

The previous discussion has shown that flow and transport in the field is generally three-dimensional and that lateral spreading and rapid gravity flow are often important in the unsaturated zone. Different types of models have been used to describe the behavior of this system. A distinction is made between local distributed parameter models and large-scale lumped parameter models. It is clear from the previous discussion that if a distributed parameter model is chosen to represent the system, the model should be three-dimensional since, at a local scale, nature varies three-dimensionally. Thus, models visualizing the system as a battery of one-dimensional, non-interacting columns (e.g., Dagan and Bresler, 1980, 1983; Milly, 1982) can be easily criticized. It is possible, however, that on a larger scale the local effects may average out and the system can be effectively described by a simplified model having uniform parameters and/or lower dimensionality. Such a possibility of simplifying the large-scale distributed systems seems probable but not yet proven. Efforts in this direction are being undertaken at MIT, using a stochastic theory.

The parameters used in the models describing the large-scale average behavior of the system are viewed as large-scale effective properties. These field-scale parameters are different from the local properties. Due to the fact that the local governing equations are parametrically nonlinear, the large-scale effective parameters cannot be estimated as a direct average of local property measurements.

Field-scale parameters are expected to depend, among other things, on the scale of the experiment, the nature of the heterogeneity, and the moisture content. Unfortunately, only a few field-scale parameters have been reported and often the analytical methods used in their estimation are questionable. Nevertheless, it may be useful to analyze reported parameters hoping that we may observe some basic relationships.

Biggar and Nielsen (1976) described experiments conducted in an agricultural setting. Breakthrough curves were obtained at six depths at the centers of twenty 6.5 m-square plots, following a pulse of Cl^- and NO_3^- solution. Then, a simple one-dimensional transport model with the constant parameters of apparent diffusion coefficient, D, and apparent pore-water velocity, v_s, was fitted to each breakthrough curve. A possible criticism of such parameter estimation procedures is that since the parameters D and v_s have been assumed to be uniform over depth they should be fitted simultaneously with the breakthrough curves at all depths (at each location). Fitting a model with different D and v_s at each depth, contradicts the assumption of constant D and v_s. The paper suggests that fitted values of D and v_s follow a log-normal distribution. D is correlated with v_s through an almost linear relationship ($D = 0.6 + 2.93\, v_s^{1.11}$). The paper further discussed the number of samples required to obtain a reliable estimate for the mean of D and v_s. This is another weak point of the paper since it implies that the mean values of D and v_s are the effective parameters that should be used in the mean models for estimation of mean concentrations. Such an assumption does not seem to be justified, since the system is non-linear in the parameters.

The experiments and methods for calibrating D and v_s reported in Van de Pol, et al. (1977) and Kies (1981) are similar to Biggar and Nielsen (1976). Thus the criticisim of the fitting D and v_s also applies. The fitted values again seem to follow a log-normal distribution. Kies' data also show that D is related to v_s, again through an almost linear relationship. These papers also imply that the mean values of D and v_s should be used in models predicting mean concentration. Again, this has not been proven. Kies' data show an increase in the mean values of D with depth below soil surface. He also observed that the average solute velocities, v_s, calculated by fitting v_s to the breakthrough curves, is larger than the mean pore-water velocities v_m. This is probably due to the fact that most of the water may be moving in larger structures with higher velocities than the average pore-water velocity (calculated by dividing average infiltration rate by average soil moisture content). He also observed that the ratio v_s/v_m increases with depth, and that ^3H travels faster and disperses more than NO_3^-.

Warrick, et al. (1971) reported a field experiment in a 6.1 m by 6.1 m plot in a Panoche clay loam soil. A solute pulse was applied to the surface of the soil and was observed as it moved through the soil profile. Experimental data and a numerical soil-moisture flow model showed that the infiltration rate approached steady state in a relatively short time. A one-dimensional governing equation for the solute concentration was solved analytically. Using this solution and the value of the velocity found from the infiltration rate, the maximum concentration was determined as a function of time and dispersion coefficient D. Comparing this model to experimental data, a value of $D = 0.07\ cm^2/min$ gave the best fit. They observed that a value of D close to $0.05\ cm^2/min$ better described the data at small times whereas $D = 0.1\ cm^2/min$ was more nearly correct at large times. This suggests that D may increase with the time or distance the solute has traveled. The approach used in this paper for evaluating D was fundamentally different from the one used in Biggar and Nielsen (1976), Van de Pol, et al. (1977), and Kies (1981). The velocity here was estimated from infiltration data and the parameter D was fitted to the maximum observed concentrations only and not to the whole breakthrough curves. A possible criticism of this approach is that information available regarding the shape of the breakthrough curves is not used in the estimation procedure, so the estimated parameters are possibly not optimal.

Warrick, et al. (1971) observed that the calculated distribution curves do not penetrate the soil profile as deeply as the measured ones. A similar phenomenon was observed in Kies (1981). This may be a reflection of preferential flow paths with most of the water moving through the larger water-filled pore sequences. Warrick, et al. (1971) also observed that solute was not present in the advancing moisture front but lagged behind nearer the soil surface. This phenomenon has been observed elsewhere (e.g., see Andersen and Sevel, 1968). It is called the solute-lag effect (Gelhar, 1977). A probable explanation is that a pressure wave generates displacement of old capillary water at successively increasing depths. Simple calculations (Gelhar, 1977) show that the moisture front travels with speed $v_w \approx dK/d\theta$ while the solute travels with speed $v_s \approx K/\theta$. For typical soils, then $v_w >> v_s$ (e.g., $v_w = 20v_s$).

Foster (1975) and Oakes (1977) present another possible explanation of the solute-lag effect in the English Chalk, based on fracture flow with matrix diffusion. For the data of Young, et al. (1976), Oakes (1977) reports a dispersivity value of 0.2 m.

The paper of Andersen and Sevel (1968) is interesting since they made an effort to evaluate an effective dispersion coefficient on a large-scale (20 m deep) system. Environmental tritium data were taken in a group of four boreholes which had been augered at time intervals of about two years. Soil-moisture profiles had been measured regularly by the neutron method in the boreholes. A simple displacement with a dispersion model, assuming constant travel velocity, constant dispersion coefficient, and constant soil-moisture content throughout the profile was tested. A dispersion coefficient of $D_L = 10^{-7}\ m^2/sec$ yielded a good fit for the tritium profiles. This corresponds to a dispersivity $A_L = D_L/v = 70$ cm. Soil moisture profiles indicated a propagation velocity of the moisture front of about 3 to 3.5 m/month. The actual flow velocity estimated by environmental tritium profiles however, seems to be only 4.5 m/year. This solute-lag effect is similar to the one observed by Warrick, et al. (1971) and discussed above.

Jury, et al. (1982) describe a series of field experiments to test a transfer function model. This model may be criticized in that it does not use any physics about the processes involved; it is a black-box model. Its parameters do not correspond to any physical quantities and must be calibrated based on available data for the particular setting under consideration. Extrapolation of the predictions of such a model to depths, settings, or conditions other than the ones for which it was derived is not possible. The data presented in the paper to validate the model clearly demonstrate our point. For example, Figure 3-1 presents measured breakthrough curves and predicted ones, obtained from the transfer function model of Jury, et al. (1982). Comparison between these curves indicates that the model overestimates the solute concentrations near the surface, yields a good fit at 90 cm, and underestimates them at larger depths. It appears that the model was calibrated so that it fits the data at the 90 cm depth. The differences between theory and experiment are statistically significant; the model prediction falls above the 95 percent confidence interval of the data at 30 cm and below it at 180 cm.

The reason for this discrepancy, we believe, is that the transfer function model, or any other one-dimensional convective transport model, yields predictions that correspond to a linear increase of dispersivity with depth. It is possible that at small depths the dispersivity increases with the depth but it could approach a constant value as the depth increases. In that case the predicted breakthrough curves of Figure 3-1 would show less spread and greater maximum values at larger depths, i.e., they would trend toward agreement with the measured data.
We attempted to evaluate an effective longitudinal dispersivity from the data of Jury, et al. (1982), Figure 5, using spatial moment methods. Because the data were incomplete, it is impossible to evaluate the mean and the variance for I = 6 cm and I = 59 cm; only the curves at I = 15 cm and I = 26 cm were used. The spatial mean

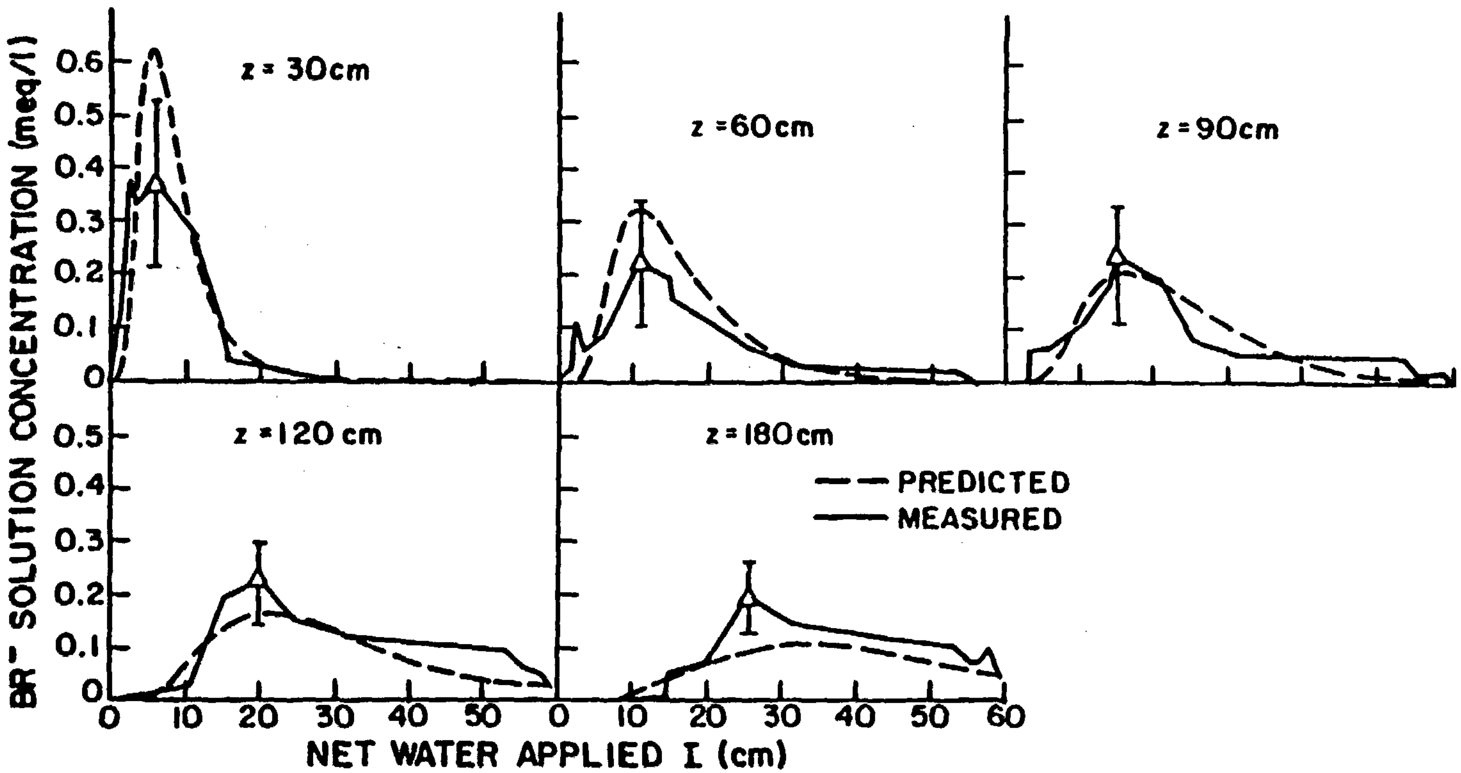

Figure 3-1. Predicted and measured average solute breakthrough curves at 30, 60, 90, 120, and 180 cm. Error bars represent 95 percent confidence limits of measured data.

Source: Jury, et al., 1982.

and variance of these curves are:

$$\bar{x}_1 = 0.76\ , \quad \bar{x}_2 = 1.17\ m \tag{3-1}$$

$$\sigma_1^2 = 0.044\ m^2,\ \ \sigma_2^2 = 0.1204\ m^2 \tag{3-2}$$

The rate of change of the spatial variance is related to the dispersion coefficient D by (see Equation 2-2)

$$\frac{d\sigma^2}{dt} = 2D = 2A_L v \tag{3-3}$$

where A_L is the longitudinal dispersivity and v is the mean velocity. With v approximately constant and $\bar{x} = vt$, we get

$$A_L = \frac{1}{2}\ \frac{d\sigma^2}{d\bar{x}} \tag{3-4}$$

And using the approximation $d\sigma^2/dx = (\sigma_2^2 - \sigma_1^2)/(\bar{x}_2 - \bar{x}_1)$, we get:

$$A_L = 9.45\ cm \tag{3-5}$$

Trautwein, et al. (1983) is also particularly interesting since measurements of the unsaturated flow system extend to a depth of 120 m below ground surface. The leakage of wastewater in the unsaturated zone beneath a waste disposal evaporation pond was studied. The soil formation consists of alternating layers of sand and clay. Borings in the vicinity of the pond revealed that 20 years after pond construction, pond water moved to a depth of 94 m. The authors attempted to model unsaturated flow beneath the pond using a one-dimensional finite element unsaturated flow model. Functional relationships were assumed for soil-moisture characteristics curves and hydraulic conductivity curves at each soil layer, depending on the soil type. These relationships were adjusted so that the results of the model would fit the mean soil-moisture data well. Comparison of the adjusted model values of saturated hydraulic conductivity, K_s, to measured laboratory values shows a large discrepancy. Field values are one or two orders of magnitude larger than laboratory values. This suggests that seepage from the pond occurred at a much faster rate than would have been predicted using the laboratory measurements of K_s. Using the laboratory measurements, the model would have predicted that the contaminant front would have moved only a few feet in 20 years, while in actuality the front moved approximately 100 m. The large differences between laboratory and field values of K_s are probably due to macroscopic features and natural soil heterogeneity not accounted for in laboratory experiments.

The authors have based their one-dimensional flow assumption on the fact that the dimensions of the pond (500 m x 1000 m) are much greater than the thickness of the unsaturated zone (approximately 100 m). However, field investigations of the site (Kent, et al., 1982), show an extremely large lateral spread of water in the unsaturated zone. The contamination plume extends laterally to a distance of about 2000 m around the pond! Some of the lateral spreading at this site may be due to the formation of saturated perched water zones above the water table. Obviously, the one-dimensional assumption does not seem justified. A three-dimensional model is required for a more realistic treatment of the flow in the unsaturated zone of this setting.

Immobile Water Models. As was discussed earlier, rapid movement of water is observed in preferential paths of structured soils, while water moves more slowly in the main soil matrix. In order to account for this and similar effects several models have been proposed in the literature. In the models proposed by Coats and Smith (1964), van Genuchten and Wierenga (1976), and Wierenga (1982), the pore water is partitioned into a mobile and immobile zone. Immobile water is held inside and between aggregates, in dead-end pores, or around soil particles. Solute movement into and out of this immobile water is assumed to be a relatively slow molecular diffusion process. It is assumed that the rate of solute transfer between the two zones is proportional to the concentration difference between the two zones. This model does not assign any specific geometry to the intra-aggregate pore-water region and does not consider solute concentration gradients within the aggregates. The transport in the mobile phase is described by the classic convection-dispersion equation. This simple model often describes experimental data better than a simple convective dispersion model at early breakthrough. Brissaud, et al. (1983) used this model to describe solute transport data in the field. They found that this model is somewhat better at describing the tails of the breakthrough curves, but it requires estimation of three model parameters instead of the one needed in a simple convection-dispersion model.

In another model proposed by Rao, et al. (1980) the porous medium is assumed to consist of porous spheres. The pore water consists of the inter-aggregate pore water (mobile water between the spheres) and the intra-aggregate pore water (stagnant water inside the pores of the spheres). It is again assumed that convective-dispersive solute transport occurs only in the inter-aggregate pore water, while the intra-aggregate pore water acts as a diffusion sink/source for the solutes. Fickian diffusion was used in order to describe the transport in each sphere. Similar simplistic models have been used to describe the diffusion of solutes from the macropore water into the soil matrix (e.g., Johnston, 1983).

The above models are quite interesting. However, they are only a very crude representation of the real system and require parameters that in most cases are difficult to determine. Their usefulness in real field situations has yet to be demonstrated.

Field-Scale Dispersivities. In the case of saturated flow it has been observed that dispersivity increases as the scale of the experiment increases. A theoretical explanation of this behavior is given in Gelhar, et al. (1979) and Gelhar and Axness, (1983). It is reasonable to expect a similar behavior for unsaturated flow. Therefore, it may prove useful to relate field-scale dispersivity for unsaturated flow to the scale of experiment.

Unfortunately, very few field-scale dispersivities have been reported for unsaturated flow systems. Most were observed in small-scale settings and the methods used in their determination vary and are often questionable. Table 3-4 gives some values of longitudinal dispersivity ($A_L = D_L/v_s$) for several laboratory and field experiments. Some of the A_L values of Table 3-4 were actually reported in the publications, while others we evaluated using simple calculations. The longitudinal dispersivities A_L were plotted as a function of the scale of the experiment and are shown in Figure 3-2. This figure shows a lack of reported values of A_L for scales larger than 2 m. Nevertheless, the few existing data points show an increase of A_L with the scale of experiment, analagous to the saturated-flow case. The longitudinal dispersivity may depend on other factors, such as soil type, soil heterogeneity and moisture content; however, it is not possible to recognize any dependence of A_L on these factors, due to the dearth of data. A theoretical stochastic approach is being developed at MIT in order to derive the relationship between unsaturated flow dispersivities and the properties of the system.

Table 3-4

FIELD DISPERSIVITIES, A_L

Author	Type of Experiment	Vertical scales of experiment (m)	Longitudinal dispersivity A_L(m)
Yule and Gardner, (1978)	Laboratory	0.23	0.0022
Hildebrand and Himmelblau, (1977)	Laboratory	0.79	0.0018
Kirda, et al. (1973)	Laboratory	0.60	0.004
Gaudet, et al. (1977)	Laboratory	0.94	0.01
Brissaud, et al. (1983)	Field	1.00	0.0011, 0.002
Warrick, et al. (1971)	Field	1.20	0.027
Van de Pol, et al. (1977)	Field	1.50	0.0941
Biggar and Nielsen, (1976)	Field	1.83	0.05
Kies, (1981)	Field	2.00	0.168
Jury, et al. (1982)	Field	2.00	0.0945
Andersen, et al. (1968)	Field	20.00	0.70
Oakes, (1977)	Field	20.00	0.20

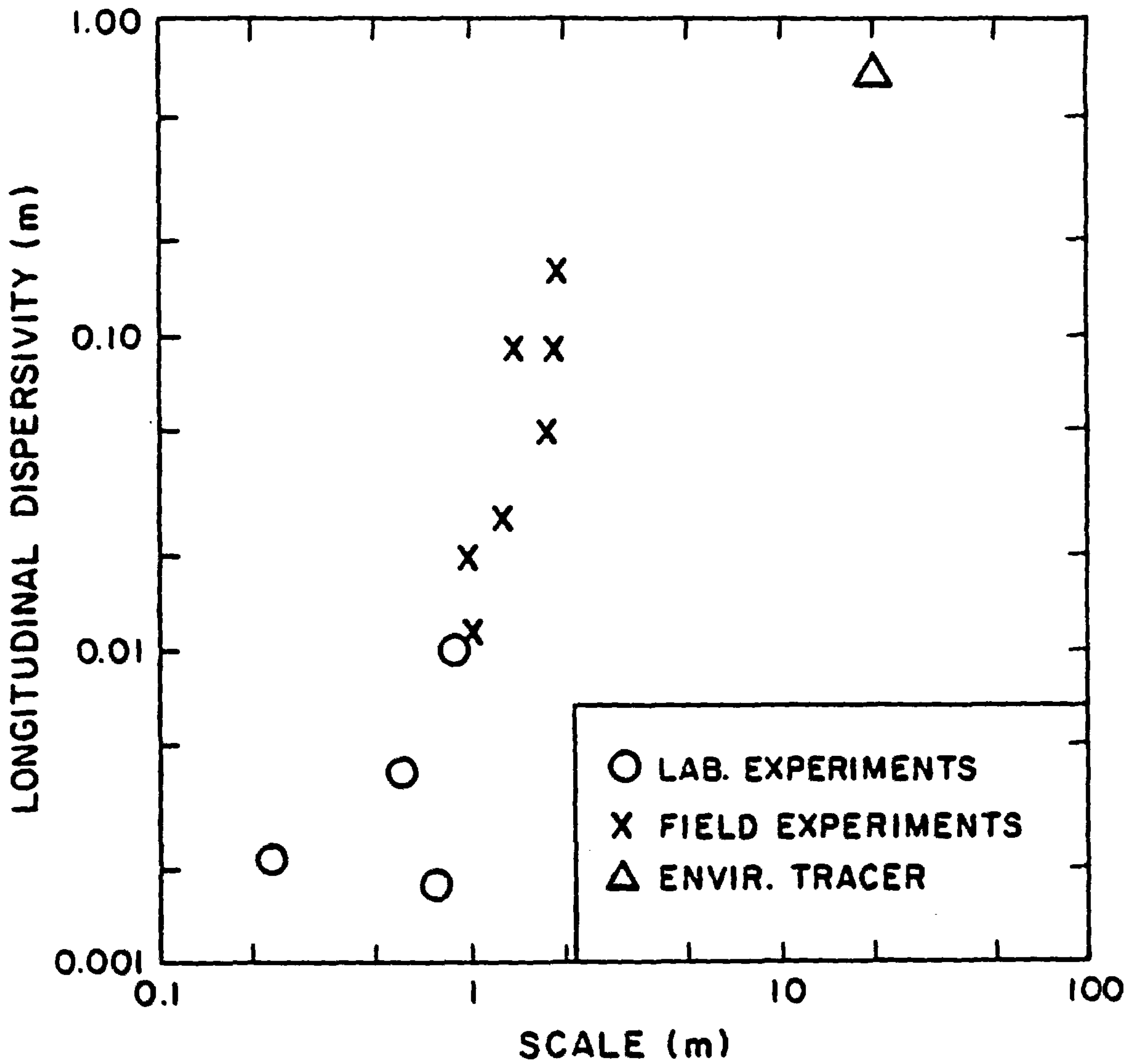

Figure 3-2. Scale of Observation Versus Longitudinal Dispersivity for the Unsaturated Zone

CONCLUSIONS AND RECOMMENDATIONS

Conclusions

A number of observations of field-scale unsaturated flow and transport have been discussed. The objective was to identify mechanisms of flow and transport that are important in field-scale problems. The discussion has shown that flow and transport are much more complicated and behave differently in the field than in the laboratory. This is probably due to the three-dimensionality of flow and the natural heterogeneity observed in the field. Unfortunately, the available field-scale observations are very few compared to the complexity of the problem. Therefore, our conclusions are mainly qualitative. Nevertheless, even qualitative conclusions are important since they relate to the design of field experiments.

Our main conclusion is that heterogeneity is of paramount importance in field-scale flow and transport. Due to natural heterogeneity, flow in the field is, in general, a three-dimensional process, even in what appears to be relatively homogeneous soil. Depending on the nature of spatial heterogeneity and water availability, certain types of flow are more pronounced. Two types of flow which relate to waste disposal control practice and modelling were observed. The first is lateral spread of moisture in stratified soils due to capillary forces. The second is fast vertical flow in fingers or channels due to gravity.

The field observations of Routson, et al. (1979), Price, et al. (1979), Crosby, et al. (1968, 1971), Prill (1977), Knoll and Nelson (1962) and Palmquist, et al. (1962) have indicated conditions for which lateral spread of soil moisture is important. The conclusion is that lateral movement is enhanced if the soil is stratified, the initial soil moisture is low, the size of the application area is small relative to the size of the unsaturated zone, and the water application rate is low. It was observed that, at high moisture tension, lateral movement is more pronounced in finer soil layers than in coarser layers. The dependence of lateral movement on moisture content suggests a soil-moisture dependence of the anisotropy of hydraulic conductivity. This is in accordance with the theoretical results of Yeh and Gelhar (1983).

This lateral flow process is pertinent to waste disposal in the thick unsaturated sediments which are common in arid settings. In line with the previous discussion, it can be expected that the waste liquid will spread laterally while vertical movement will be relatively limited. As a result, contamination may arrive at the water table much later than predicted by a classical one-dimensional model, and areally extensive contamination may result.

Starr, et al. (1978), Quisenberry and Phillips (1976), Johnston, et al. (1983), Thomas and Phillips (1979), Bouma (1981), and Beven and Germann (1982) discussed several cases of fast gravity flow. A particular case is that of fast gravity flow occurring in discrete fingers in a coarse-textured soil layer, if this layer is overlain by a saturated fine texture soil layer (Starr, et al., 1978; Palmquist, et al. 1962). Another type of fast gravity flow occurs in macroporic features such as cracks, root holes, and fissures. These structures are more important if they are open to more permeable surface materials or the ground surface. In addition there should be a significant source of water, especially for the larger pores, since these pores will be water filled only at tensions close to zero (i.e., near saturation). The significance of these pores also depends on the initial soil moisture, the relative size and concentrations of pores, and the soil structure.

A consequence of such fast flow in solute transport is that salts in the soil surface will be moved to a much greater depth by rain or irrigation than would be predicted by displacement type models. On the other hand, it is probable that much of the salt will be bypassed and remain near the soil surface. In arid regions, with low rates of liquid additions, which is the case in many waste disposal control problems, fast gravity flow may not be important.

Some implications of the above flow mechanisms for the practical problem of modelling large-scale unsaturated flow and transport systems will now be discussed. In the case of dry, stratified soils the fact that lateral flow can be significant should not be ignored. This implies that three-dimensional models are required. As was discussed earlier, the anisotropy of the effective unsaturated hydraulic conductivity for such soils is dependent on the moisture content and this should be taken into account in the modelling process. No sufficient information exists on longitudinal and transverse dispersion coefficients for such systems, but it is expected that they will be larger than laboratory values due to natural heterogeneity, which causes spatial variability of velocity.

It seems improbable that the classical flow and transport models (Darcian, Fickian) will be capable of predicting flow and transport in cases of fast gravity flow in macrostructures or fingers. Nevertheless, it may be possible (but not probable) that such models can be used if appropriate effective model parameters are selected. Effective hydraulic conductivities, for example, may be one or two orders of magnitude larger than the local values. (Bouma, 1982; Trautwein, et al. 1983).

A general conclusion of the review is that flow in the field is a three-dimensional process and velocity is highly variable in space, even in what appear to be relatively homogeneous soils. It is expected that such velocity variability will tend to increase the field-scale dispersion coefficients. It is also expected that field-scale dispersion coefficients increase with the size of the experiment. Some evidence indicating such dependence was given in Figure 3-2.

Recommendations

A clear conclusion of this review is that existing field observations of unsaturated flow and transport are very limited, considering the complexity and importance of the problem. Natural heterogeneity plays an important role in field-scale vadose-zone flow and transport. Small laboratory models canno realistically produce the natural heterogeneity exhibited in the field, so it is difficult to draw conclusions from such experiments for the processes occurring in the field. In order to better understand the mechanisms of field-scale flow and transport, obtain field-scale parameters and test theories and models, carefully designed large-scale field experiments are definitely needed.

In particular, it is important to understand how lateral flow and fast gravity flow depend on soil type, stratification, water application rate, size of water application area, and initial moisture content. A series of experiments in various soil settings is necessary to develop quantitative understanding of these mechanisms. In order to produce realistic results for waste disposal applications, the experiments should be carried out at pertinent vertical scales of up to several tens of meters. Traceable solutes should be introduced over areas of comparable horizontal extent. Because of its critical role in field-scale transport, three-dimensional spatial variability of soil properties should be measured at the experimental sites.

Modelling of field-scale unsaturated flow and transport requires knowledge of how the large-scale effective model parameters depend on spatial variability, scale, moisture content and other factors. The proposed experiments will yield the large-scale effective parameters. Using the stochastic theory, these effective parameters can be predicted from statistical parameters determined by measuring the spatial variability of soil. The stochastic approach thus provides a systematic method of predicting large-scale parameters from small-scale data. One goal of the proposed experiments is to test the validity and practicability of this approach.

Section 4

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