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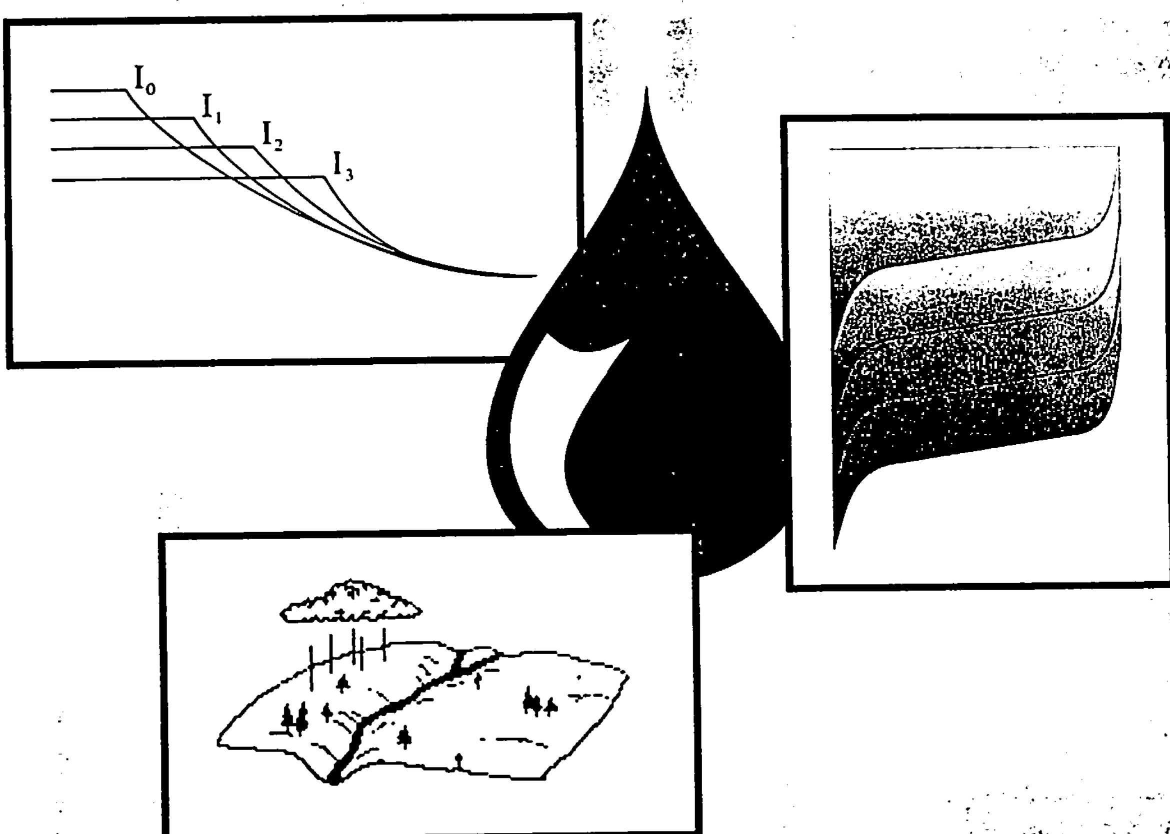

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Statistical and Geostatistical Characterization of Spatial Variability

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Introduction and Background

Classical statistical approaches assume field variability is purely random and observations of soil hydraulic properties are statistically independent regardless of their spatial position (Vauclin et al., 1983). Statistical characterizations include parameter estimates of a mean, variance and distribution. These estimates can be used as effective parameters or Monte Carlo simulations can be made using the distribution of the data. However, variations in soil properties tend to be vertically and horizontally correlated over space. Because of this, geostatistical characterizations are often used to incorporate the variability into infiltration and chemical transport modeling. Geostatistical methods facilitate the examination of spatial and temporal correlations in the data (Nielson et al., 1973). Tools such as kriging allow estimation of spatially correlated data using measurements taken in close proximity to the point at which the estimate is being made.

Statistical characterization of spatial structure of hydraulic characteristics is important for several forms of analyses (Unlu et al., 1990): such as 1) estimating point or spatially averaged values of soil hydraulic properties using kriging techniques, 2) designing sampling networks and improving their efficiency, and 3) stochastic modeling in order to understand the overall response of heterogeneous flow systems (Kitanidis and Lane, 1985).

Field Observations

Despite its utility, the application of geostatistics to modeling of unsaturated zone processes is a relatively new field, limited by a lack of high quality field data. Nielsen et al. (1973) performed one of the first extensive field experiments quantifying the variability of hydraulic properties of a Panoche clay loam over a field area.. Significant variability was found in the particle composition, soil-water pressure, bulk density, and saturated and unsaturated hydraulic conductivity in the soil profiles. Nielsen et al. (1973) reported an average and a standard deviation (sd) of the saturated hydraulic conductivity (Ks) in the vertical direction of 1.5 cm hr^{-1} and 0.05 cm hr^{-1} respectively. Byers and Stephens (1983) obtained core samples in horizontal and vertical transects in order to study the statistical and stochastic properties of particle-size parameters and Ks. Laboratory measured Ks was found to be log-normally distributed, with a mean and a sd of 61.2 cm hr^{-1} and 36 cm hr^{-1}. Hopmans et al. (1988) examined data collected in horizontal and vertical transects over a 650 ha watershed. The mean ln Ks value reported was 1.7 with a sd of 0.6 (Ks in cm day^{-1}). Brace (1980, 1984) and Clauser (1992) found permeabilities tend to grow with the characteristic scale of measurement for both sedimentary and crystalline rocks.

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Carvallo et al. (1976) measured unsaturated hydraulic conductivity ($K(\psi)$) versus depth in five infiltration plots within a 0.01 ha area in a Maddock sandy loam with an average particle composition of 80 % sand, 11 % silt and 9 % clay. Tensiometers were installed at seven depth intervals down to a maximum of 1.52 m. Soil-water characteristic data determined from soil samples were used in conjunction with tensiometer measurements to compute $K(\psi)$. Significant variability in $K(\psi)$ was found both between plots and over vertical profiles. Standard deviations of $K(\psi)$ ranged from 0.3 $cm\,hr^{-1}$ at saturation to 0.003 $cm\,hr^{-1}$ at the residual water content.

In addition to these values, correlation scales for various soil characteristics have been reported in the literature. Some of the reported values for ln Ks are presented in Table 1. These data include values for vertical Ks along horizontal transects only. As the table shows, considerable variability exists between the observed correlation scales and also with respect to the material.

Russo and Jury (1988) found the sample network configuration had a distinct effect on estimation of covariance function parameters. They compared the effect of two different sampling networks for estimating the covariance function on the predicted head variance. It was found that using a modified sampling network with irregular spacing was superior to use of a systematic sampling network (regular grid) for estimation of parameters of the covariance function. This was particularly true for fine textured soils when a three-dimensional treatment of hydraulic conductivity variations was employed and when relatively small correlation scales were present.

Saturated and residual water contents have been measured extensively and representative statistical characteristics for these parameters are available in literature. Some values are presented in Table 2. An excellent review of similar data is presented in Jury (1985).

Relationship to Scale

Examination of the data in Table 1 reveals a close relationship between the overall scale at which the measurements were taken and the correlation scale of the ln Ks data (Fig. 1). A very high correlation coefficient (0.93) exists between these two variables. For a more general case, data assembled by Jury (1985) also indicate a significant positive correlation between the correlation length of several measured soil physical properties and the sample grid spacing used to develop the variogram or correlogram. The implication of this is that correlation length is not a property of the measured soil, but rather a property of the measurement methods and scale. When working with soil pH data, Gajem et al. (1981) found an apparent correlation length of 1.5 m when a 0.2 m measurement grid spacing was used, a correlation length of 21.6 m when a 2.0 m spacing was used, and a correlation length of 130 m when a spacing of 20 m was used.

Table 1. Observed standard deviations and correlation scales for hydraulic conductivity measured along horizontal transects.

Source	Material	Parameter	Standard Deviation	Correlation Scale (m)	Overall Scale (m)
Delhomme (1979)	limestone aquifer	ln Ks	2.3	6,300	30,000
Binsariti (1980)	basin fill aquifer	ln Ks	1.0	800	20,000
Russo and Bressler (1981)	Hamra Red Mediterranean soil	ln Ks	0.4	1	100
Luxmoore et al. (1981)	weathered shale subsoil	ln Ks	0.8	1	14
Sisson and Wierenga (1981)	silty clay loam soil (alluvial)	ln Ks	0.6	0.1	6
Viera et al. (1981)	Yolo soil (alluvial fan)	ln Ks	0.9	15	100
Devary and Doctor (1982)	alluvial aquifer (flood gravels)	ln Ks	0.8	820	5,000
Byers and Stephens (1983)	fluvial sand	ln Ks	0.5	1	15
Hoeksema and Kitanidis (1985)	sandstone aquifer	ln Ks	0.6	45,000	50,000
Unlu et al. (1990)	Panoche soil, 75 cm depth	ln Ks	0.6	8	80
Unlu et al. (1990)	Panoche soil, 105 cm depth	ln Ks	1.0	9	80
Anderson and Cassel (1986)	Portsmouth sandy loam, A horizon	ln Ks	722	1.5	500
Anderson and Cassel (1986)	Portsmouth sandy loam, A horizon	ln Ks	5.5×10^4	2.5	500
Anderson and Cassel (1986)	Portsmouth sandy loam, A horizon	ln Ks	2.2×10^3	na	500
Hopmans et al. (1988)	sand, BC horizon	ln Ks	0.8	0.75	12.5

Table 2. Representative saturated and residual moisture content and standard deviations for various soil textures.

Source	Parameter	Soil Texture	Soil Horizon or Orientation	Mean	Standard Deviation
Hopmans and Stricker (1989)	θsat	sand	A	0.406	0.032
Hopmans and Stricker (1989)	θsat	sand	BC	0.391	0.045
Hopmans and Stricker (1989)	θsat	sand	D	0.437	0.055
Wierenga et al. (1989)	θsat	gravelly sandy loam	vertical	0.321	0.032
Wierenga et al. (1989)	θr	gravelly sandy loam	vertical	0.086	0.020
Burden and Selim (1989)	θsat	silt loam	horizontal	0.54	0.045
Burden and Selim (1989)	θr	silt loam	horizontal	0.14	0.095
Nielsen et al. (1973)	θsat	clay loam	vertical	0.454	0.045
Cameron (1978)	θsat		vertical	0.470	0.045
Anderson and Cassel (1986)	θsat	sandy loam	A	0.457	0.071
Anderson and Cassel (1986)	θsat	sandy loam	Btg	0.391	0.100
Anderson and Cassel (1986)	θsat	sandy loam	Bg	0.310	0.071
Anderson and Cassel (1986)	θr	sandy loam	A	0.096	0.032
Anderson and Cassel (1986)	θr	sandy loam	Btg	0.075	0.032
Anderson and Cassel (1986)	θr	sandy loam	Bg	0.044	0.032
Carvallo et al. (1976)	θsat	sandy loam	vertical	0.393	0.014
Russo and Bressler (1981)	θsat		0-0.9 m	0.367	0.045
Russo and Bressler (1981)	θr		0-0.9 m	0.078	0.045
Greminger et al. (1985)	θsat	Yolo loam	0.3 m	0.459	0.010
Greminger et al. (1985)	θsat	Yolo loam	0.6 m	0.486	0.014

Russo and Jury (1987) studied the uncertainty of the estimation of correlation length scales for stationary fields using 100 computer generated realizations of a two-dimensional isotropic second-order stationary field with a given correlation length. They found both the accuracy of correlation scale estimates and the fitted variogram increase with the number of sample points and as the correlation scale of the underlying process increased. To obtain reasonable estimates of the correlation scale, the distance between sampling points must be smaller than half the range of the underlying process. However, data from a one-dimensional transect may result in underestimation of the correlation scale of the underlying process by a factor of two or more.

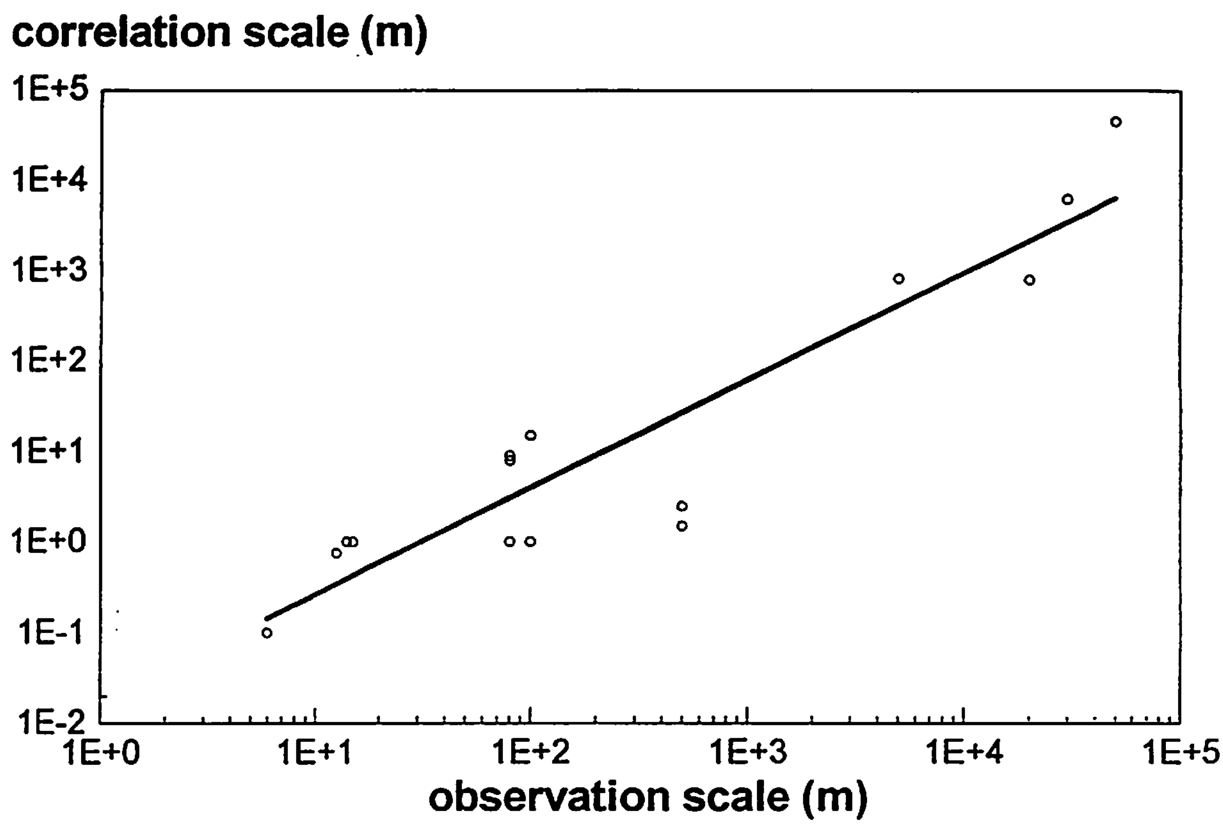

Fig. 1. Relationship between observation scale of the measurements and the calculated correlation scale for the Ks data of Table 1.

Neuman (1994) also observed an increase in estimated correlation scale with increasing measurement scale in computed aquifer permeabilities and dispersivities. He reviewed the data summarized by Gelhar (1993) and plotted the correlations scales of hydraulic conductivity and transmissivity data versus the characteristic lengths of the corresponding fields sites. The data indicated a consistent increase in reported correlation scale with field length over a range of correlation scales from 10 cm to 45 km. In general it appeared the correlation scale was typically 10 % of the field scale. For figure 1, the correlation scale was approximately 8 % of the observation scale.

To explain the phenomena observed by Neuman (1994), he noted that if hydrogeologic media is viewed as a nested sequence of distinct hydrogeologic units with a discrete hierarchy of scales then "one obtains a semivariogram function that increases with the mean tracer travel distance(s) in a stepwise rather than gradual fashion" as the semivariograms of the discrete units are superimposed. This would correspond to crossing from one soil type or geologic material to another. Further, each step in such an echelon represents a natural correlation scale at which the log permeability is statistically homogeneous or nearly so; while other scales are locally either inactive or suppressed (Neuman, 1994). The change in dominant natural scales from one setting to another can result in an infinite variety of possible semivariograms. To resolve this variation Neuman (1994) stated that data from a large number of settings would be required so that enough scales are sampled to ascertain the "underlying commonality of these semivariograms in the form of a generalized power law."

Interpretation and semivariogram analysis assumes a natural scale over which the property in question is statistically homogeneous. As Neuman (1994) points out, increasing correlation scales with increasing field scale implies statistical homogeneity of log permeabilities is at best a local phenomenon occurring intermittently over narrow bands of the scale spectrum. Hence one must question the utility of routinely associating geologic medium properties with REV's as has been the custom for several decades.

Using theoretical simulations, Starks and Fang (1982) showed that a nonlinear drift in the mean value of the parameter, if not filtered out before analysis, would produce an apparent increase in correlation length as the sample density was decreased. In the theory of regionalized variables from which geostatistics is derived, each measured parameter is considered to be a single realization taken from a single probability distribution. In order to apply this theory it is necessary to assume the property is spatially stationary so that each location is described by the same probability distribution, and spatial covariances depend only on the separation between measurements and not on the absolute location (Journel and Huijbregts, 1978). Thus, a drift, or a change in the mean of the data would be a violation of the underlying assumptions. Data sets collected along long transects crossing over several different soil types may contain this nonlinear drift component.

Summary

As this review has shown, considerable variability exists in natural porous media. This variability precludes precise characterization of the hydraulic parameters. Experience has shown soil parameters vary widely and at best we can determine the statistical and spatial characteristics of this variability. It appears that geostatistics has considerable applicability to infiltration science. However, the relationship between correlation scale and measurement scale is disturbing. The correlation scale is not solely a function of the parameter being measured, but also a function of the measurement scale. While interesting from an observational standpoint, it is not clear whether the relationship is useful in the predictive sense. On the average, it appears the correlation scale is approximately 10 % of the observation scale. However, from site to site this may vary considerably. Application of this relationship in scaling-up would not account for the site specific differences. Thus, care must be taken in study design, data analysis and data interpretation.

One possible avenue to approach the problem centers around soil classification. Studies which sample across two different soil types break the underlying assumptions behind geostatistical analysis. Characteristics of different soil types are expected to have different means and distributions. Crossing over from one soil type into another, as is frequently done in these studies, violates the assumption of spatial stationarity. Our efforts may be better directed toward characterizing either single soil types or examining relationships between soil types lying in similar landscape positions.

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