APPLICATION AND EVALUATION OF TECHNIQUES WHICH DESCRIBE THE SPATIAL VARIATION OF SOIL-PHYSICAL AND HYDROLOGICAL VARIABLES

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Introduction

Soil properties vary in space. In the traditional statistical analyses, the spatial variability is expressed by a mean, variance and coefficient of variation of a sample population. Assuming that samples are stochastically independent, and that the central limit theorem applies one can determine the minimum number of samples required to estimate the mean within a certain predefined confidence interval.

It is expected, however, that closely spaced samples are more correlated to each other than samples farther apart. The traditional analysis of soil sample data takes no account of the location of a particular sampling site in the field or the spatial correlation of soil properties. A measure of such spatial dependence (a characteristic field length or correlation length) can be obtained from an autocorrelogram.

Autocorrelation coefficients can be calculated when the data are statistical homogeneous or second-order stationary. According to the theory of regionalized variables (Journel and Huybregts, 1978) the collection of all values of a characteristic over a region is the realization of a random function. The assumption of stationarity of order two is satisfied if the expected value of this random function is the same in each point of the field of interest and if the spatial covariance of the values in each pair of points in the region exists and is the same all over the field and depends only on their distance.

Russo and Bresler (1981a) defined the integral scale, which is the largest average distance for which the values of a property are correlated. Sisson and Wierenga (1981) introduced the autocorrelation length which is the distance at which the correlation coefficient between measured values is at most 1/e, while Gajem et al. (1981) defined the zone of influence as the greatest distance for which the autocorrelation coefficient is statistically significant greater than zero. Occurrence of repetitious or cyclic patterns can be identified from spectral analysis. A power spectrum partitions the sample variance around the mean into components by frequency (Sisson and Wierenga, 1981; Nielsen et al., 1983).

For situations where the second-order stationarity assumption does not apply (e.g., variance is infinite), the intrinsic hypothesis concept was introduced. Instead of the stationarity of the covariance, it requires the difference between the values in a pair of points to have a finite variance which value will also depend on the separation distance (lag) only. A plot of the variance of increments versus the lag is called a variogram. The shape of the variogram gives an indication of the spatial dependence of the soil property considered. The average distance over which points are significantly correlated is called the range.

The use of variograms or co-variograms to interpolate between an existing set of observations is called kriging. Owing to the fact that the variogram gives the expected relation between pairs of neighborhood points, different weights are given to the surrounding points, depending on their distance from the one to be interpolated. A strong advantage of kriging, over many other interpolation methods, is that the prediction variance can itself be estimated. Analysis of the variance makes it therefore possible to determine an optimum sample size (Vieira et al. 1981) and to verify the validity of kriging assumptions (Vauclin et al. 1983) by a jack-knife procedure. If a set of observations may prove not to be sampled sufficiently to yield interpolated values at other locations of acceptable accuracy, one may consider to include the spatial correlation between that variable and another more frequently observed variable. That procedure is called cokriging (Vauclin et al. 1983). It may also be the case that a drift exists within the field of interest. A procedure, which takes the drift into account, is called universal kriging (Delfiner, 1976; Delhomme, 1978).

Spatial variability of soil physical properties can partly be described by the similar media concept (Miller and Miller, 1955 and 1956). Similar media differ only in the scale of their internal microscopic geometries. The use of similar media concepts allows results, either experimental or computed, of soil-water behavior in one soil to be used to describe the behavior in another by employing reduced variables defined in terms of appropriate microscopic characteristics lengths. It can be deduced that, if λ is the microscopic characteristic length for a given soil, the soil-water potential (h) of similar soils may be scaled by multiplying by λ, and hydraulic conductivities (K) by dividing by λ^2. The purpose of the scaling method is to simplify the description of statistical variations of soil-water properties. By this simplicication, the pattern of spatial variability is described by a single scale factor α, that relates the soil hydraulic property of each location to a representative mean ($\alpha=\lambda/\lambda_m$). Youngs and Price (1981) extended the similarity concept to geometrically dissimilar soils. However, in that case scale factors calculated from the soil water characteristics (h-scale factors) are likely to be different from those calculated from the hydraulic conductivity data (K-scale factors). Various methods to determine the scale factors are described by Warrick et al. (1977a), Peck et al. (1977), and Ahuja et al. (1984a). Russo and Bresler (1981a) and Ahuja et al. (1984b) found a linear relation between the K-scale factors and h-scale factors.

Knowledge of a representative soil water characteristic and hydraulic conductivity function are necessary for modeling soil water flow in a deterministic way. Inherent to the use of spatially dependent hydraulic properties, however, is the use of stochastic models. One may e.g. determine experimentally the soil hydraulic properties at many locations in a field and simulate water flow for each location. This will result in a distribution of hydraulic functions and output variables (soil water potential or volumetric water content for unsaturated flow). This approach was followed by Russo and Bresler (1981a). One may, on the other hand, use Monte Carlo simulations to generate many possible different hydraulic functions. This is especially attractive when the variability of these functions can be represented by a single stochastic scale factor α for which its distribution is known (Warrick and Amoozegar-Fard, 1979; Warrick et al. 1977b; Freeze, 1975). In general the distributions of the scale factors are obtained from experimental data. Tests for normality or lognormality include: X^2-test, Kolmogorov Statistic (Rao et al. 1978), and fractile diagrams. A priori knowledge of the α-distribution was assumed by Peck et al. (1977), who studied the effect of spatial variability of soil hydraulic properties on water budget modeling in a watershed. In a similar way, Clapp et al. (1983) showed that heterogeneity in soil hydraulic properties may account for approximately 75% of the observed standard deviation in water content.

Freeze (1975), in simulating one-dimensional saturated water flow, compares the results, obtained from many possible realizations of a flow parameter, with the results when average values for the flow parameters are used. The definition of an 'equivalent' uniform porous medium (average parameter values for whole profile) underlies deterministic modeling. Freeze concludes that the use of an equivalent porous medium concept is not justified in transient groundwater modeling. Also Bresler and Dagan (1983) concluded that the traditional deterministic approach, thereby using effective soil properties, may not be meaningful for unsaturated water flow in spatial variable fields. The study by Freeze (1975), as well as most other studies related to the stochastic nature of unsaturated flow assume the flow parameters to be statistically independent. Smith and Freeze (1979a and b) utilized a first-order nearest-neighbor model to generate conductivity sequences with a known spatial structure. The one- or two-dimensional domain was divided into discrete blocks, and conductivity values were generated that allowed neighboring conductivity values to be autocorrelated. Smith and Freeze concluded that the standard deviation in water potential values was approximately halved when the one-dimensional saturated water flow problem was extended to its two-dimensional analog. In addition, their results indicated that the uncertainty in the output variable was strongly dependent on the strength of the correlation structure between neighboring conductivity values. Similar conclusions were drawn by Bakr et al. (1978) and Gutjahr et al. (1978) who used spectral analysis to solve perturbed forms of the groundwater flow equation, describing flow through a porous medium with a spatially dependent hydraulic conductivity. This spectral approach has the advantage of yielding a closed form solution for the statistical properties of hydraulic head.
Anderson and Shapiro (1983) found good agreement between the technique employed by Smith and Freeze and the spectral analysis technique, when applied to one-dimensional unsaturated water flow.

Delhomme (1979) and Smith and Schwarz (1981) introduced the concept of conditional simulations to analyze the uncertainty resulting from the spatial variability of transmissivity. A regular grid was superimposed on the flow field and from sample values a kriging estimate of the transmissivity value was computed at each node. Through conditional simulations other possible realizations of the transmissivity field were generated which had the same variability as the unconditional kriging estimates and which preserved the sample values at the measurement points. After numerical solution of the partial differential equation for each conditional simulation, one can then statistically analyze the resulting probability distributions of the model output.

Data collection

A total of approximately 1100 borings were made in a 650 ha study area. Profile descriptions for each boring, containing information on the root-zone, groundwater and horizon characteristics, were used to establish a soilmap. In addition, soil hydraulic properties for each horizon were determined at three different scales of observation. In the first sampling scheme 7 profiles across the study area were examined. These 7 sites were chosen in such a way that together they were representative for the whole area, and therefore included all horizons present in the study area. In the second sampling scheme, an area of 0.5 ha was chosen such that the 7 sites within this area were all from the same soil map unit. Finally, the highest sampling density was achieved by sampling at 6 sites within a 2 m^2 area.

Soil water characteristics were obtained in the laboratory by the suction method and in situ by simultaneous measurement of soil-water tension and volumetric water content by tensiometers and neutron probe, respectively. Hydraulic conductivity functions were determined by the crust method (Bouma, 1977), the hot air method (Arya et al., 1975) and/or the sorptivity method (Dirksen, 1979).

Data analysis and modeling (suggestions)

Soil hydraulic properties (soil water characteristics and hydraulic conductivity functions) were not determined and/or analyzed in the same way for the three different sampling schemes. It seems that statistical comparison of the hydraulic properties requires equal treatment. For example, calculation of the hydraulic conductivity function from the soil water diffusivity as determined by the sorptivity method should be the same for all available data.

No tabular data are required for the numerical solution of the water flow equation if the hydraulic properties are represented by analytical expressions. Van Genuchten (1978) presented a closed-form analytical model that makes it possible to represent both the soil water characteristic and the hydraulic conductivity function with the same parameters. If use of measured saturated hydraulic conductivity values results in deviation between measured and calculated hydraulic conductivity values, an intermediate measured K-value should be used to match theoretical functions with experimental data.

It can also be useful for modeling that the distribution of hydraulic properties is known. Therefore, means and variances must be calculated and a X^2-test can be used to test for (log) normal distributions. From these distributions one can calculate how many samples are needed to obtain satisfactory average estimates of soil physical variables for a given area, or obtain different realizations of the hydraulic functions by Monte Carlo simulations (Russo and Bresler, 1981a). It might be possible that hydraulic properties can be estimated from easy to obtain parameters, e.g., particle size data, bulk density or porosity (Puckett et al., 1985).

The scaling method is proposed, as it combines data on soil hydraulic properties from many locations and describes the pattern of spatial variability. Scaling parameters calculated from soil water characteristics should be compared with those calculated from hydraulic conductivity functions, and if possible be related with each other (Warrick et al., 1977a). The scaling technique can be applied among and within the 3 sampling schemes. Comparison of scaled and measured hydraulic properties (regression analysis) will indicate which soils should be scaled independently. The range of scale-factor values at different depths indicate which horizon has the largest spa-

tial variation.
Correlation lengths can be calculated from variograms and correlograms of soil physical properties (Russo and Bresler, 1981b; Peck, 1980). A relatively rapid rate of lateral variation of soil properties induces lateral water movement and affects therefore the one-dimensional vertical flow assumption.

Initially one may solve the unsaturated one-dimensional water flow equation for many possible realizations of the hydraulic function (Russo and Bresler, 1981a; Freeze, 1975). This will result in distributions of output variables (e.g. volumetric water content and actual evapotranspiration), of which the mean can be compared with the results when scaled mean soil hydraulic properties are used or with actual measured data. Another possibility is to determine the distributions of depth of rootzone and groundwater table, and to add their variation together with the varying soil physical properties to the flow model. The effects of lateral water movement in the unsaturated and saturated zone, caused by spatial variable soil hydraulic properties, can be analyzed by numerical solving two one-dimensional flow systems simultaneously. Simulations of saturated water flow indicated that an increase in the dimension of the flow domain decreases the spatial variation of output variables.

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