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SOIL PHYSICS RESEARCH AND WATER MANAGEMENT

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Abstract

A major problem now facing a soil physicist is how to deal with water management on a large scale, to identify representative hydrodynamical soil properties and to investigate their variation; to determine average values of the different terms of the water balance; and to predict, based on those values, the possible effect of different treatments on water management. This paper focuses on the present stage of research in soil physics to answer these questions and to foresee lines of research to be developed in view of actual gaps.

The presentation centres on three main topics: (1) Identification of soil hydraulic parameters. Field measurement of the 'water content-water pressure-hydraulic conductivity' relations is of major importance in the soil-plant continuum and in simulating or predicting water flow. Simple, quasi-empirical techniques of identification, based on soil textural or structural parameters, could give a better spatial representation of these parameters than the techniques being used at present. (2) Averaging over a site. The recognition of the importance of spatial variability of soil parameters, vegetative systems, water fluxes and water input has been one of the major advances of research in soil physics and hydrology during the last decade. The impact on actual practice has, however, been small. Referring to the classical work of geostatisticians in the field of ore research, and more recently in soil physics this paper stresses the importance of a proper method of sampling. (3) Development of predictive models. Very accurate numerical models have recently been developed, even to represent complex flow situations. Such models represent powerful tools to extrapolate results, and to predict other situations. However, the extent to which a model represents the reality is limited by the reliability of its transfer coefficients. In this view, the development of deterministic models using stochastic parameters, initial and boundary conditions to account for spatial variability is probably the most important goal that we must attend. Much discussion is presently taking place on the practical way to treat the parameters; either with Monte-Carlo type of estimation, or simply through sensitivity analysis within the domain of variation.

This presentation will use both recently published examples from the literature and field results. Soil physics can, in the near future, make an important contribution to solve problems of water management only if researchers are willing to contribute more to the practical field experimentation.

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SOIL physics research has experienced a rapid growth over the last few decades. In a comprehensive report on the state of the art, Philip (1974) estimated that in 15 years the scientific activity in this field will be doubled. Although this figure is difficult to verify, it is backed up by the increase in the number of students, the creation of new research groups and the rate at which new journals have been appearing. This phenomenon is simply the result of the recognition of the important role of soil physics in today's high priority fields of agronomy, hydrology and the environmental sciences. Considering water resources management alone (quantitative and qualitative), it is now clear that an understanding of the physics of heat and mass transfer in the soil-plant-atmosphere continuum and the characterization of the functions that define the movement and retention of water in the soil are essential for any rational approach. Some important applications are plant-water consumption, groundwater recharge, soil drainage, losses of soil mineral constituents, transfer of pollutants, salinization hazards, etc.

This Conference offers a special opportunity to take stock of the present situation and to determine whether this abundance of research has solved the water resources problems, and if not, to identify the shortcomings and define the research areas that must be given priority.

I would first like to cite the conclusions of three earlier papers that have already considered this question.

Philip (1974) in his paper concluded:

"In these 50 years, soil physics has gained enormously in its self-confidence, its intellectual power, and its relevance to practical problems of the real world.... In the effort to meet the need (for an adequate quantitative predictive science), the soil physicist has become more prepared to avail himself of the general intellectual resources of physical sciences.... It must be conceded, however, that this maturity has not resolved all our difficulties. Our greatest success has been in the study of *local* processes . . . and it is clear that soil physicists must address themselves more to the question as how we may best use our understanding of small scale processes in attacking (the) larger problems."

Three years later, in his review entitled "Soil Physics — Reflections and Perspectives", presented at the Annual Meeting of the American Society of Agronomy, Swartzendruber (1977) concluded:

"Soil physics presently stands at a stage of great opportunity and challenge.... There is much unfinished business such as just getting under way with the crucial tasks of testing and validating mathematical and physical theories under field conditions...."

Finally, Moltz *et al.* (1979) in their paper presented at the Quadrennial Meeting of the International Geodesics and Geophysics Union (Canberra) observed as follows:

"During the past 4 years, notable advances were made in many areas of unsaturated zone hydrology.... In much of this work, it is evident that the most universal problem is the measurement and representation of soil hydraulics properties on a scale useful for application . . . while new instruments and methods have been devised for making *in situ* measurements of soil hydraulic properties, nothing approaching a breakthrough can be claimed.... Will increasingly sophisticated measurements enable us to gradually apply our conceptual models in a truly representative manner, or will the resulting measurements to be so ambiguous that they ultimately force the development of a radically different theory? This is the central question, and a great deal of future effort will be devoted to obtaining an answer."

Can we now, in 1982, offer the practitioner, whether field investigator, development engineer or planner, a more optimistic reply concerning our ability to solve the large-scale problems? Are the necessary tools available to close the gap between highly developed theoretical, numerical and experimental techniques at the laboratory level and the empirical handbook methods used by the practitioner in charge of development projects?

In an attempt to answer these questions, I offer this review of relevant publications illustrated by field results from work carried out in collaboration with the Institut Sénégalais de Recherches Agricoles (Centre de Recherches Agronomiques, Bambey, Sénégal), the Ministère de la Recherche of Ivory Coast, the Institut des Savannes, Centre de Recherches de Bouaké, Ivory Coast, and the Institut de Recherches Agronomiques Tropicales, Paris. This research is part of the coordinated program of the Joint "Food and Agriculture Organization/International Atomic Energy Agency" Division on "Isotope and Radiation Techniques for Efficient Water and Fertilizer Use in Semi-arid Regions." As a participant in these field programs, it seems to me that three essential questions might be posed by a soil physicist in charge of such development projects:

1. To what extent are the methods used in the measurement and characterization of the soil properties valid? Can they be simplified without a loss of information?
2. How should samples be taken on the site in order to obtain characteristic mean values for the test?
3. How can tne results be extrapolated, especially to predict the influence of the type of agronomic treatment, crop rotation, tillage, etc., taking into account the natural spatial variability of the site?

This paper was prepared with these three questions in mind.

Measurement Validity and Parameter Estimation

I would first like to state the following principle:

Only an approach based on a physical model can hope to give an objective solution useful for extrapolation.

As a result, all exploratory programs must involve the measurement of descriptive variables and the determination of transfer coefficients. This

section is limited to the purely hydraulic aspects (for solute flow, refer to Biggar *et al.* 1977).

The *descriptive variables* include the soil volumetric water content (θ in cm^3/cm^3), and the soil-water pressure (h in cm). Field measurements are now widely employed, using *neutron moisture probes* and *tensiometers* for repetitive and non-destructive measurements.

The transfer coeficients to be considered are the relationships between the water pressure and the water content (the $h(\theta)$ curve that determines the capacity of the soil to retain water), and between the hydraulic conductivity and the water content (the $K(\theta)$ curve that determines the capacity of the soil to transmit water). These two relationships, referred to as the *hydrodynamic characteristics* of a soil, can be determined in the field by well-known methods (Van Bavel *et al.* 1968; Vachaud *et al.* 1978).

Two important points that have received very little attention until today must be looked at closely:

1. What is the representativeness of the measurements of the descriptive variables (θ and h) in relation to the errors coming from the techniques used to determine them?
2. What is the uncertainty resulting from error propagation in the calculation of the hydrodynamic characteristics.

Water Content Measurement Error

The most extensive study appears to have been made by Sinclair and Williams (1979). From an analysis of variance of functions of stochastic variables (water content at a given depth, neutron count rate and probe position), the authors show that the neutron moisture meter readings involved three variances:

location error, including the spatial variability between different points on the site (see the next section).

instrument error, due to neutron scattering according to a Poisson distribution.

calibration error, due to uncertainty in the slope of the calibration curve.

This investigation is applied to two types of problems which will be encountered throughout this study, i.e., the estimation of *water content* at an elevation Z, time t and various locations p, or the estimation of the *change in water content* at elevation Z and position p over a time interval Δt.

The determining factor in the two cases is the *variance of the slope* of the calibration curve which can lead to an uncertainty greater than the spatial variability. The authors show that for any number of samples, a coefficient of variation of 15% for the slope leads to an estimation variance of up to 4 or 5%, while the location effect (spatial variability) decreases in $1/n$ where n is the number of measurements. The authors conclude that it is better to *limit* the number of measurements and concentrate on an *accurate* characterization of the calibration curve.

On this basis, Haverkamp (1981) recently estimated the *local* error (in one access tube) in the water content, including a probe positioning uncertainty terms (±2 cm). Fig. 1 shows the results for two types of soils: an aeolian sand in Senegal (DIOR soil, Bambey) and a clayey soil in the Ivory Coast (A6, Bouaké). The calibration curve and the measurement uncertainty (due to the instrument and the calibration) are given as a function of the measurement for each of the two soils. The Bouaké soil is more typical of agricultural soils with relatively low drainability. The slope variance is high (6×10^{-3} instead of 6×10^{-5}) as it is difficult to obtain a high variation of the θ range in the field. As a result, water content measurements with a relative accuracy of less than 10% cannot be hoped for at this site for a given tube.

This point is extremely important. Calibration techniques have generally fallen behind developments in instrumentation. It should be possible, following the method developed by Couchat *et al.* (1975), to offer the practitioner a less primitive method than those presently used (Williamson and Turner 1980) and thereby increase the confidence in the measurement of θ.

Pressure Measurement Error

A similar analysis can be made of tensiometer measurements. Although there is no calibration error in this case, instrument error (readings) and in particular position and interpolation errors can become preponderant, especially for low values of hydraulic head ($H = h - z$, z being the depth from the soil surface). Haverkamp (1981) also carried out this calculation for the two soils already described. Fig. 2 gives an estimation of the errors in h and θ for the measurement points of the $h(\theta)$ curves obtained in the field by using both tensiometers and a neutron probe at the same time.

Error in the Estimation of K(θ)

The most widely used method for the field estimation of $K(\theta)$ is the "Internal drainage" method (Hillel *et al.* 1972). The errors in the measurements of the descriptive variables (water content and pressure) are obviously propagated in the calculations of the *integrals* and derivatives with time (to obtain the flux) or with space (to calculate the hydraulic gradient) in order to arrive at Darcy's equation:

$$K = -\frac{W}{G} = \frac{-(d/dt)\int \theta \cdot dz}{dH/dz\,|_t}$$

The uncertainty in the experimental determination of the $K(\theta)$ curve was

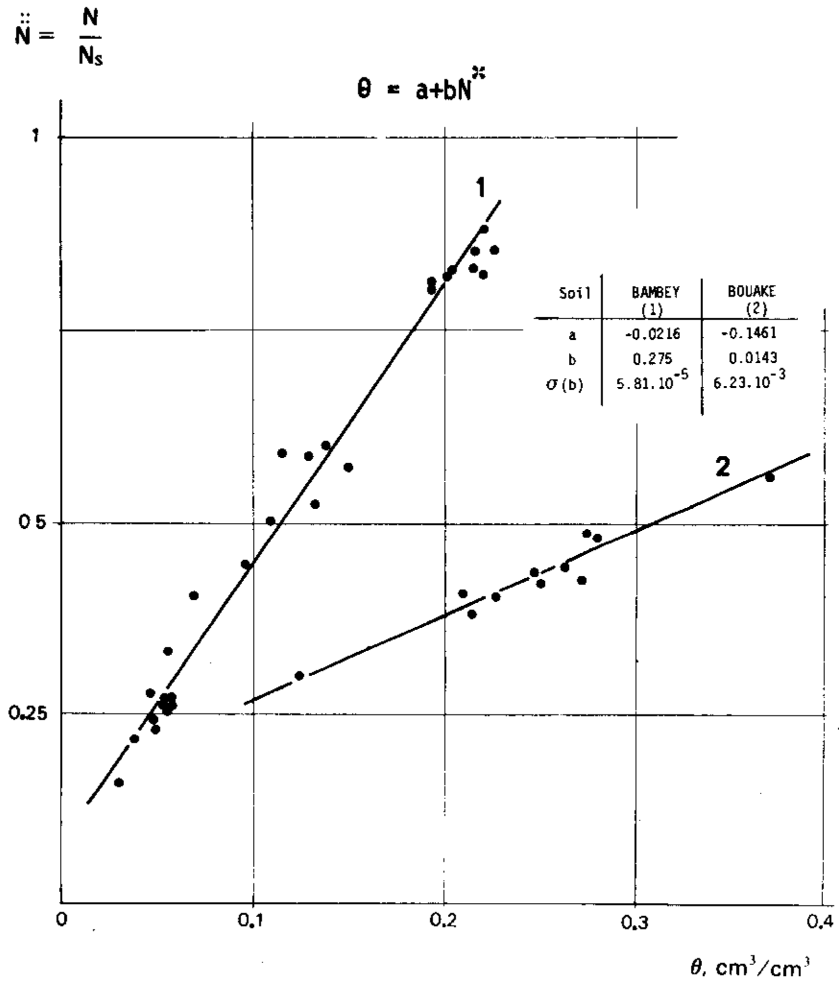


Fig. 1.a. Neutron moisture meter calibration curves obtained on the sites of Bambey (1), Senegal (Imbernon 1981) and Bouake (2), Ivory Coast, (Kalms *et al.* 1981).

not taken into account until the study carried out by Fluher *et al.* (1976) applying Taylor expansion to the study of the variance.

The importance of this investigation is twofold. First, it gives a relatively simple means of determining the measurement uncertainty: for a drainage test on sandy loam, the relative uncertainty (for a 68% confidence interval) varies from 50% for high water contents to more than 25% for dry soil (a result of

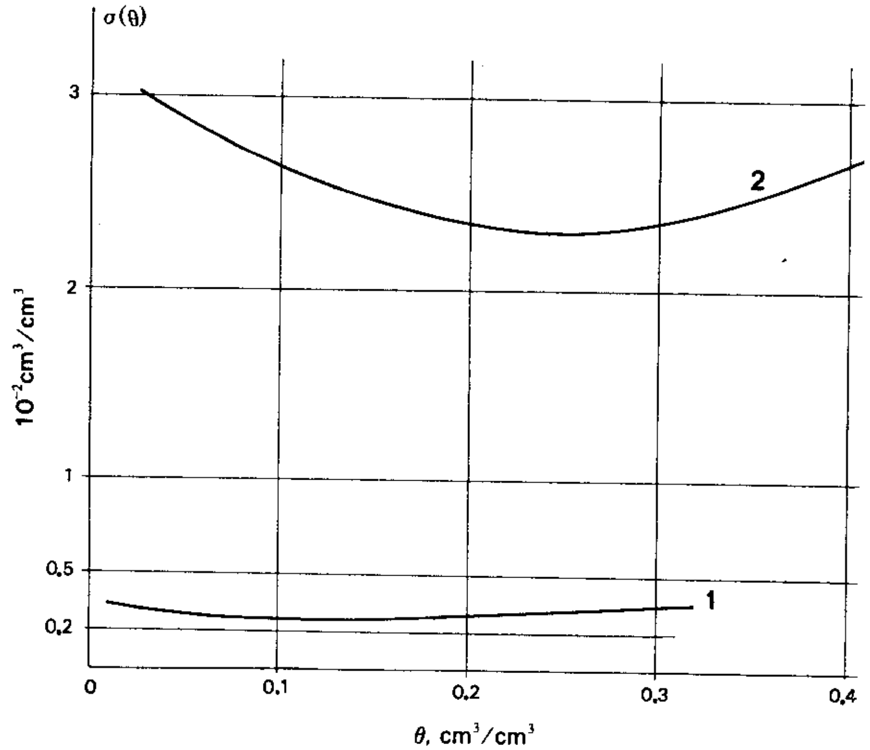


Fig. 1.b. Error in the determination of water content as a function of the count rate for the curves given in Fig. 1.a.

the very high error in the estimation of the flux when the variations in θ are low). Secondly, the authors present a measurement optimization strategy that involves giving the most importance to the water content readings at the beginning of drainage.

This method was recently improved by Haverkamp (1981), who showed the advantage of using an objective *smoothing function* on the data. Fig. 3 shows the experimental values and the uncertainty band for the characterization of the K(θ) relationship of the A6 soil from Bouaké, obtained either by direct application of the Fluher method or by imposing the following smoothing constraints:

The variation of the water content with time at a given elevation is exponential (Nielsen *et al.* 1973).

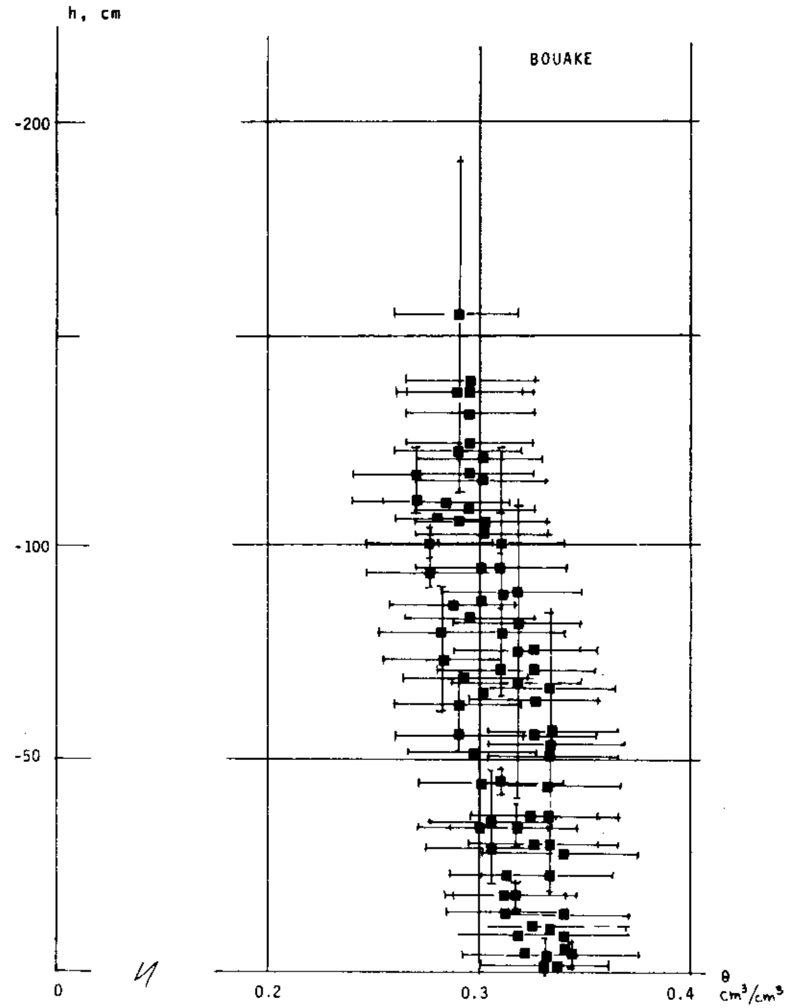


Fig. 2. Uncertainties in the characterization of the soil water pressure-water content relationship obtained at the site of Bouake, Ivory Coast, during an internal drainage experiment. (from Haverkamp, 1981).

The hydraulic head profile in space corresponds to a third degree polynomial with a zero slope at the soil surface.

Considering the experimental constraints, many articles have been

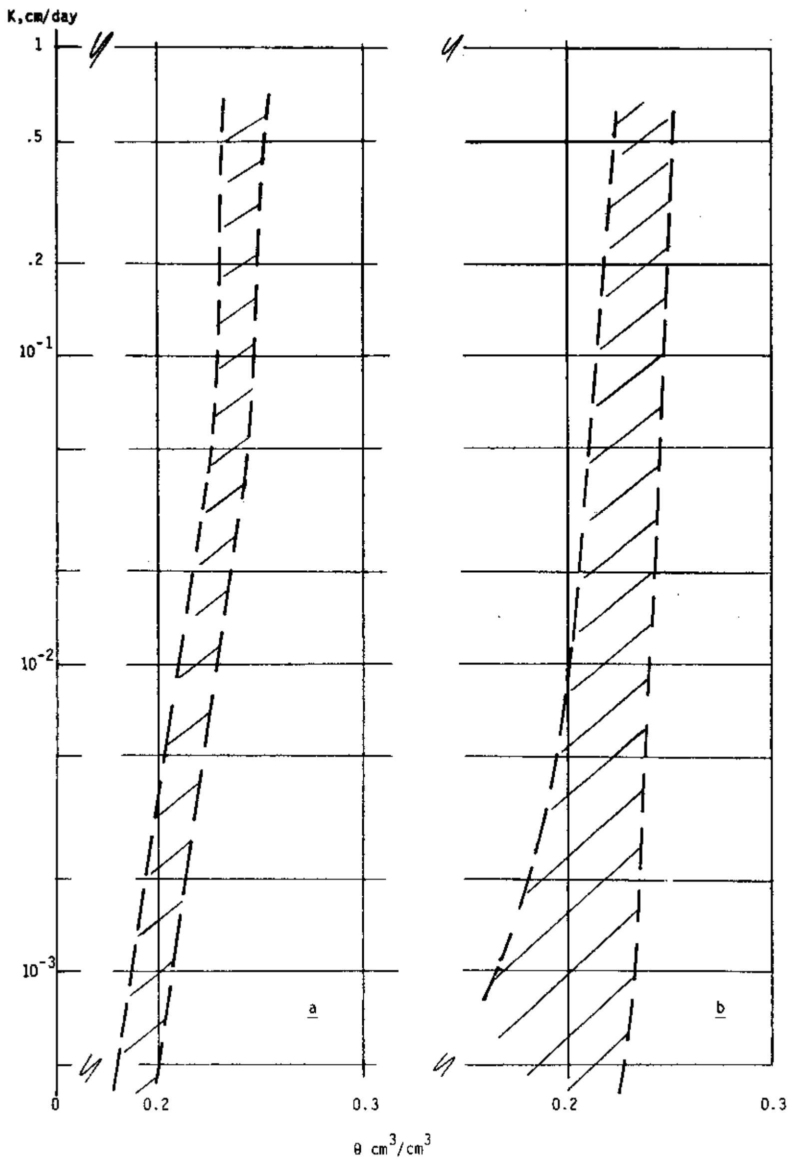


Fig. 3. Estimation of error propagation on the determination of K(θ) at Bouake using:
— the smoothing constraints of Haverkamp, 1981 (Fig.3.a)
— or the method of Fluher, (1976) (Fig.3.b)

published on the estimation of the K(θ) relationship as a function of h(θ). A review was published by Mualem (1978). These estimation methods led to many test programs aimed at establishing their validity with respect to test measurements (Brust *et al.* 1968; Green and Corey 1971; Rogowski 1971; Roullier *et al.* 1972; Nielsen *et al.* 1973; Campbell 1974; Clapp and Hornberger 1978; Parkes and Waters 1980). However, these comparisons are on the whole biased as they include no estimation of the measurement uncertainty band.

Methods for Estimating the Hydrodynamic Characteristics

Given the magnitude of the experimental uncertainty bands, it appears worthwhile to continue along the lines of the previous attempts and look for direct approximations of the functions h(θ) and K(θ).

In my opinion, all *a priori* attempts at estimating h(θ) from pore interconnection theories and capillary models are doomed to fail as they depend on arbitrary parameters for the transfer from the microscopic to the macroscopic scale. Furthermore, the uncertainty in h(θ) is much less than in K(θ), it can be easily measured and represents the intrinsic characteristics of the poral interspace.

On the other hand, a physically based approximation of the K(θ) relationship would appear to be as good as a measurement. At present, two parallel approaches can be used:

> A capillary type model based on an estimation of K(θ) from the h(θ) relationship. The model which appears to have the most well-founded physical basis is that of Mualem, later used by Van Genuchten (1979). The Millington-Quirk model (see Nielsen 1973) has also been widely used.
>
> An empirical model (an exponential or power law for the K(θ) variation correlated with soil texture parameters (Clapp and Hornberger 1978).

In both cases the practitioner is at best confronted with a model with *two parameters* that must be determined:

(i) A Calibration Parameter

The most important measurement to calibrate the model is the hydraulic conductivity *at saturation*. Note that measurement of this fundamental value still includes very high uncertainties and more accurate methods should be found as soon as possible. Infiltration tests require large areas to obtain negligible lateral seepage effects (Imbernon 1981). Well permeameter tests (Talsma and Hallam 1980) give highly local measurements and involve an identification technique which is not always valid. Crust method tests (Baker and Bouma 1976) and air-entry permeameter tests are extremely delicate and very local.

(ii) A Shape Parameter

The shape parameter is either fixed arbitrarily or by calibration using an experimental curve.

In our work, we have applied two procedures:

A characterization of the hydraulic conductivity at saturation and of the shape parameter by numerical simulation of an infiltration test with an iterative least squares fitting to the infiltration curve (Haverkamp *et al.* 1979).

A correlation between the conductivity, the shape parameter (both determined by least squares method using an experimental $K(\theta)$ relationship) and a soil texture parameter (Fig. 4) (Vachaud *et al.* 1982).

The first method represents a lighter procedure for the characterization of $K(\theta)$ while the second can be used to *extrapolate* point measurements. Both methods have proved to be highly effective.

As a conclusion for the first part of this report, the following two points must be emphasized:

Although the first point might seem trivial, it concerns the fact that *measurements are not free of errors*, which becomes important when the practitioner is interested in differences either at a point (measurement of water consumption or of flux) or between a number of points (cultivar comparisons or averaging techniques).

The second point concerns the use of information coming from the knowledge of the $K(\theta)$ relationship. In spite of the high uncertainty, this relationship gives order of magnitude approximations of the possible fluxes. Numerical simulation should be able to give the limiting values of probable flows by means of *sensitivity analysis*. Nevertheless, the use of $K(\theta)$ relationship to make an a *posteriori* determination of the flux values using Darcy's equation in the low water content range is doubtful as the uncertainty in $K(\theta)$ is added to that of the gradients (Souza *et al.* 1979). Simple estimations based on the conservation equation should be more representative in any case and much easier, especially for arid conditions, even if flow through the base of the soil profile is not completely zero (Vachaud *et al.* 1982).

Dealing with Spatial Problems

Up to this point we have been concerned with the local scale. Since the fundamental study by Nielsen *et al.* (1973), soil physicists have all realized that it is impossible to characterize behaviour over a parcel by means of a local measurement. The cumulative probability curve in Fig. 5 shows tensiometric measurements obtained the same day and at the same depth (60 cm) by a set of 22 tensiometers distributed over an 1800 m^2 parcel of land with a uniform

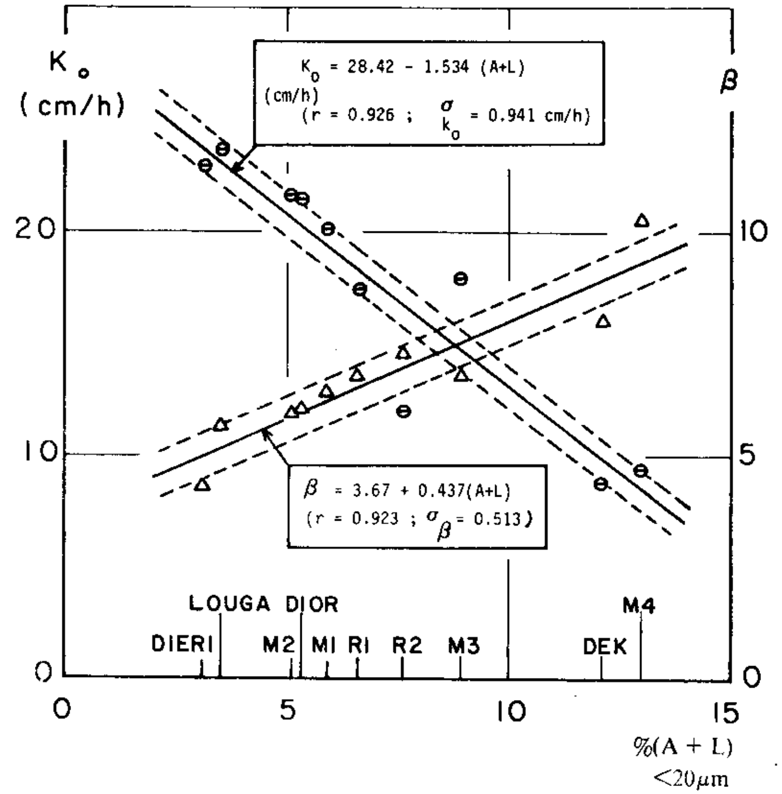


Fig. 4a. Empirical correlation between the saturated hydraulic conductivity (K_o) and the power exponent β of the $K(\theta)$ relationship written as $K(\theta) = K_o (\theta/\theta_o)^\beta$ as a function of the silt + clay content (A+L, <20μ) for 10 sites in the sandy soil of the North-Central part of Senegal.

cultivar of rainfed rice in Bouaké. The $\pm V$ band corresponds to hydraulic head measurements from −214 cm to −597 cm (mean, −405 cm).

This type of variation for water content or flux measurements is characteristic of the spatial heterogeneity of the soil-water properties and has been reported by many investigators. In addition to the reference article by Nielsen *et al.* (1973), other papers which should be cited include Rogowski (1972), Biggar and Nielsen (1976), Carvallo *et al.* (1976), Sharma *et al.* (1980), and Imbernon (1981). Note that by spatial heterogeneity we refer to a *stochastic* heterogeneity in an apparently uniform domain, which is stationary (independent of position and time) (Philip 1980).

In spite of the rather pessimistic viewpoint of Philip (1980) that "no

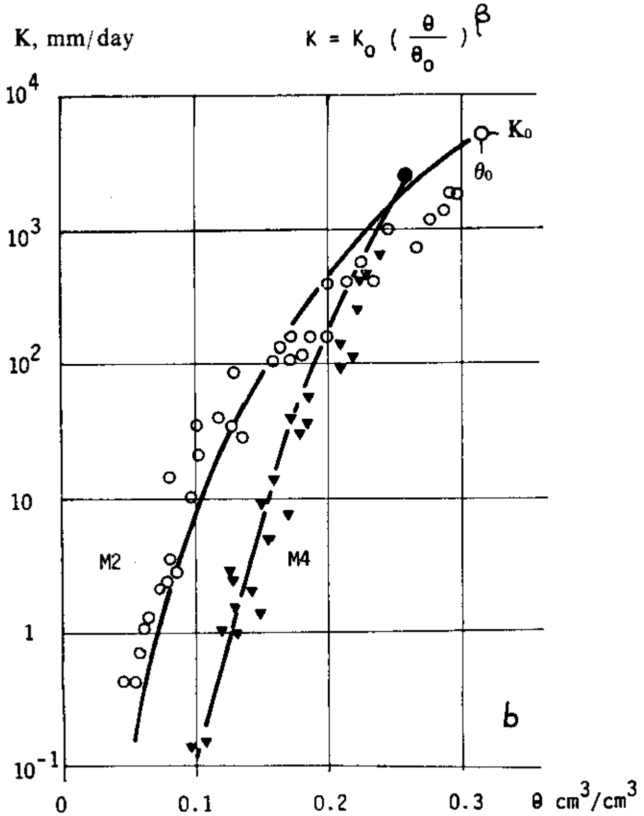


Fig. 4b. Application of fig.4a for two sites of characterization in Bambey, M2, with a mean silt content of 5.9%, M4, with a mean silt content of 15.8%. Figures are experimental points. Continuous line is obtained from Fig. 4.a.

conceivable process of averaging sample values of parameters can give a useful estimate of the flow behaviour of the total system," all of today's research work tends towards a methodology that can be used to estimate the mean behaviour of a homogeneous media equivalent to the stochastic system, and to define the uncertainty band for the results. This approach requires knowledge of the heterogeneity *structure* defined by the *frequency distribution* of the variable (Fig. 5) and its spatial or temporal *autocorrelation* function.

In this field, the recent contribution of geostatistics (David 1977) has been of prime importance in the development of analysis techniques, especially concerning the desophistication in the use of variograms or correlograms to define the limiting distance after which two points are no longer

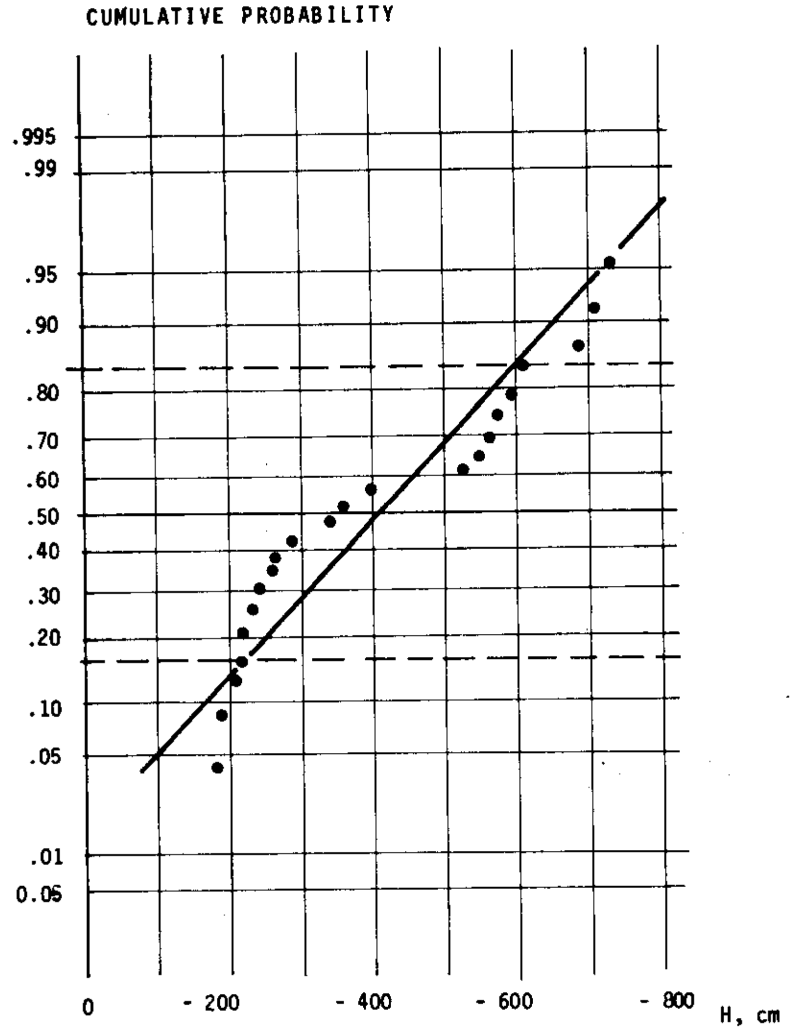


Fig. 5. Cumulative distribution of probability for 22 hydraulic head measurements (tensiometer reading) obtained the same day (Nov. 27, 1980), same depth, under a uniform cultivar of rainfed rice at Bouake, Ivory Coast. The continuous line corresponds to a normal distribution with a mean of −405.6 cm and a standard deviation of ±191.7 cm. (from Kalms *et al.* 1981)

autocorrelated. The knowledge of this "range" is fundamental, as the classical statistical analysis methods (variance, standard deviation, test of validity) can only be applied to observed values beyond this range. Inside this distance, weighting techniques must be used to account for autocorrelation between measurements, which has in particular led to the use of the kriging technique (Delhomme 1979; Hajrasuliha *et al.* 1980). For example, Fig. 6 shows three types of spatial correlograms obtained at Bambey in Senegal (Imbernon 1981) concerning pairs of samples at different lag distances for the silt and clay content, the sorptivity and the water content measured at a given depth and a given day. The range for this site is about 25 m. Hajrasuliha *et al.* (1980) found an autocorrelation range of about 400 m for soil salinity measurements at one of the test sites. In another area, all the measures were autocorrelated up to a distance of 800 m, the variograms showing a significant drift.

Knowledge of the autocorrelation function is also very important in the design of the *sampling procedure*, as measurements spaced at a distance less than the range are of no use if classical analysis techniques are to be used. Note, however, that the basic concept of autocorrelation range may be biased. A recent study by Gajem *et al.* (1981) shows that the range of the domain of influence can depend on the spacing between measurements. The range went from 57.5 cm for the measurement of the 15 bar water content, with a spacing of 20 cm between measurements, to 4600 cm for the water content in the field with a spacing of 200 cm during an exploratory program carried out at Tucson, whereas the two quantities are highly correlated.

For independent measurements, the *frequency distribution* must be known in order to determine the mean and the standard deviation. Extensive work carried out under the direction of D.R. Nielsen has made us all aware of the fact that the retention terms (water content and storage) are generally normally *distributed* while the transfer terms (flux and conductivity) are *log-normally distributed.* Fig. 7 shows the histogram of water flux measurements obtained by Imbernon (1981) in Bambey (Senegal) at a depth of 110 cm, 5 days after the end of infiltration tests carried out under identical conditions at 22 points over an area of 1 ha. The parameters of the log-normal distribution take on the following values when fitted to the measurement points:

—estimation of the mean: 14.4 mm/day
—68% confidence interval: 5.25 to 39.4 mm/day

An estimation based on a normal distribution would have given a mean of 26.2 and a 68% confidence interval from 10.5 to 42 mm/day, corresponding to an error of 100% for the mean.

Vachaud *et al.* (1982) give another example obtained on the same site, concerning the estimation of nitrogen losses for a peanut crop. For six measurements taken over one hectare, the cumulative loss over the growing season varied from 2 to 70 kg/ha for the different measurement points, with a mean of 14.6 kg/ha (Fig. 8).

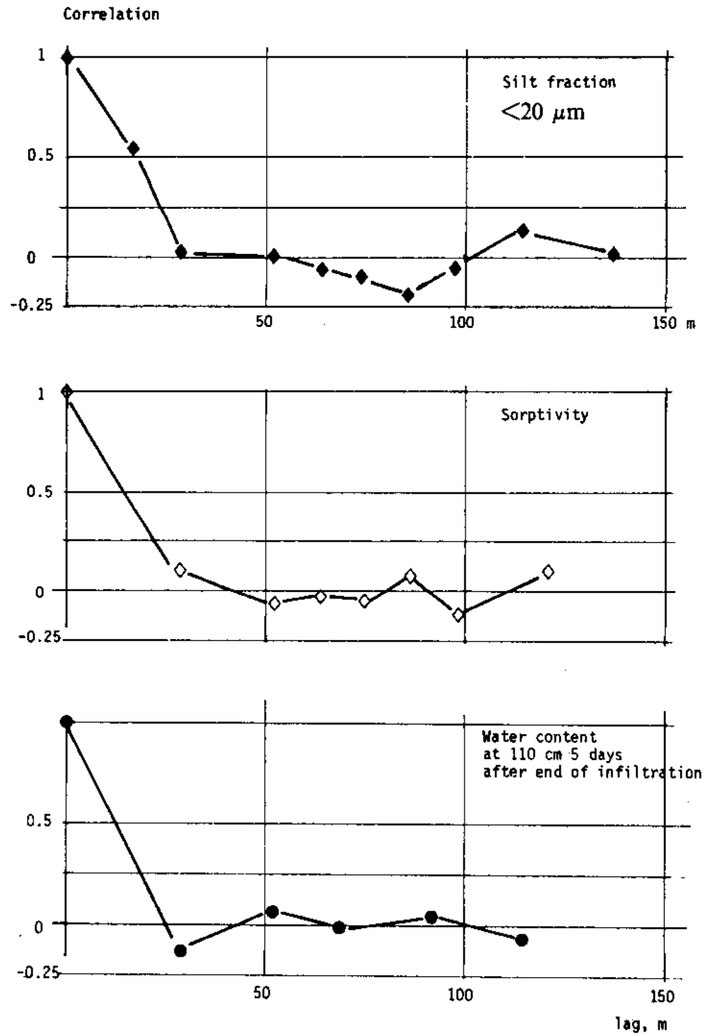
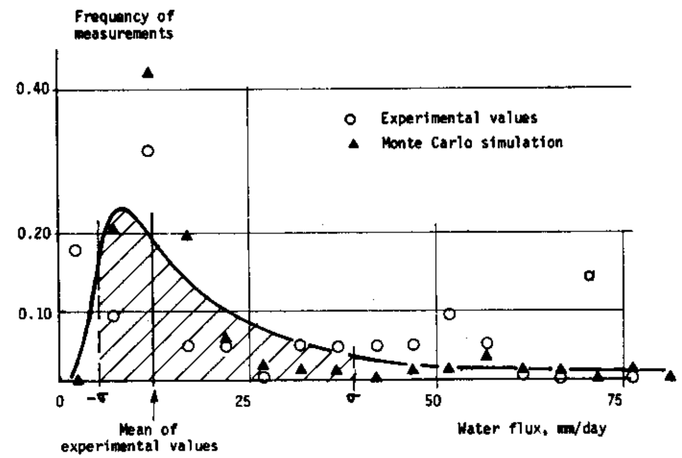


Fig. 6. Spatial correlograms obtained in a 1 ha site of sandy soil deposit at Bambey, Senegal, and relative to the study of:
— the mean silt and clay fraction on 0-110 cm
— the sorptivity (from 22 double ring infiltrometer tests)
— the water content at 110 cm, 5 days after the end of infiltration
(from Imbernon, 1981).



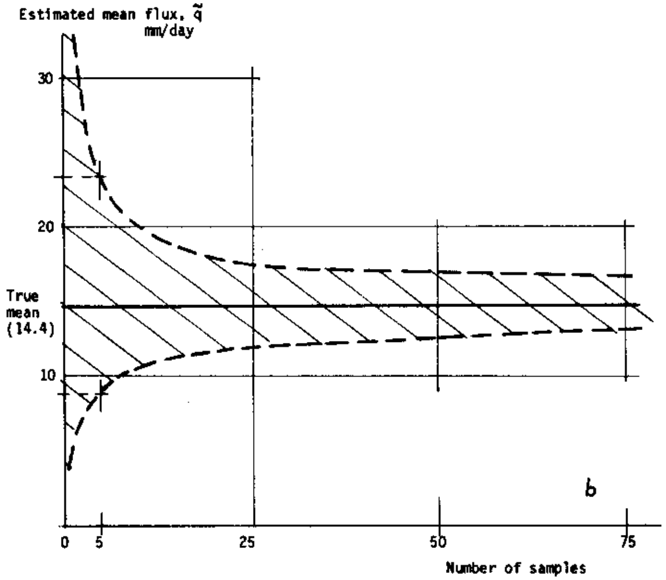


Fig. 7a. Frequency distribution of water flux measured at 110 cm, 5 days after the end of an infiltration in the site of Bambey, Senegal (24 experiments on 1 ha). The plain line corresponds to the log normal distribution. The hatched area is the domain of uncertainty on the experimental values with p = 0.68.

Fig. 7b. Evaluation of the estimation of the mean water flux q (corresponding to Fig. 7.a) as a function of the number of sampling, assuming a probability of 0.68.

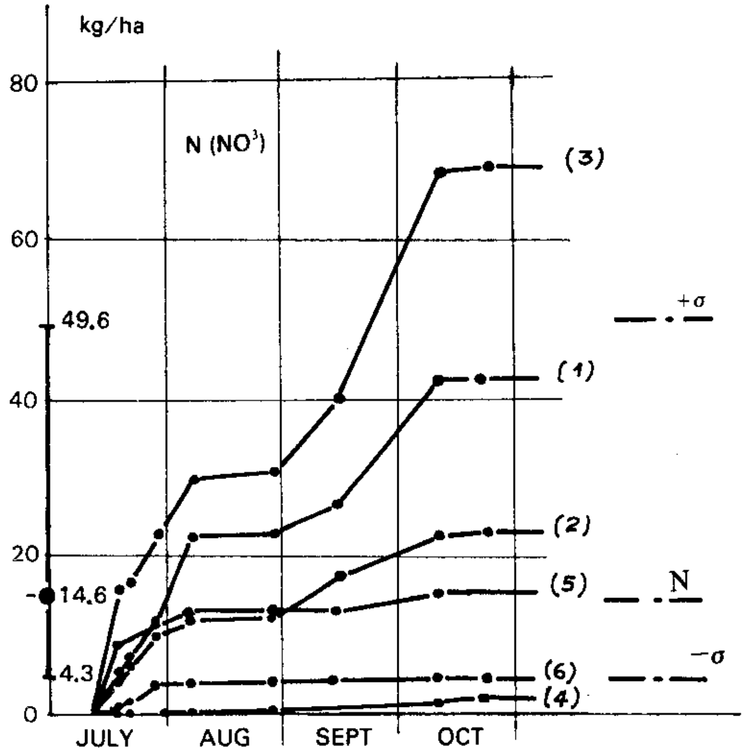


Fig. 8. Estimation of cumulative nitrogen losses under peanut from 6 indepent sites of measurements (neutron moisture meter, tensiometers solute concentration cells) at a depth of 110 cm under a one hectare field of sandy soil at Bambey, Senegal, (from Vachaud *et al.* 1982).

A final example is taken from tests carried out at Bouaké (Ivory Coast) involving a comparative study of water consumption for two varieties of rainfed rice on two 1800 m² areas with the same soil and with six neutron moisture probe measurements for each variety (Kalms *et al.* 1981). Fig. 9 shows the cumulative probability distribution for water storage from 0 to 90 cm and the storage variation over a period of 2 weeks. Three points can be noted: these measurements follow a normal distribution, the variability in the soil completely covers any possible differences between varieties, and the variance in the storage variation (consumption) is much less than the variance in the storage. The last point can only be explained by the existence of a very strong covariance between measurement sites, probably due to a smoothing of the spatial heterogeneity by root extraction.

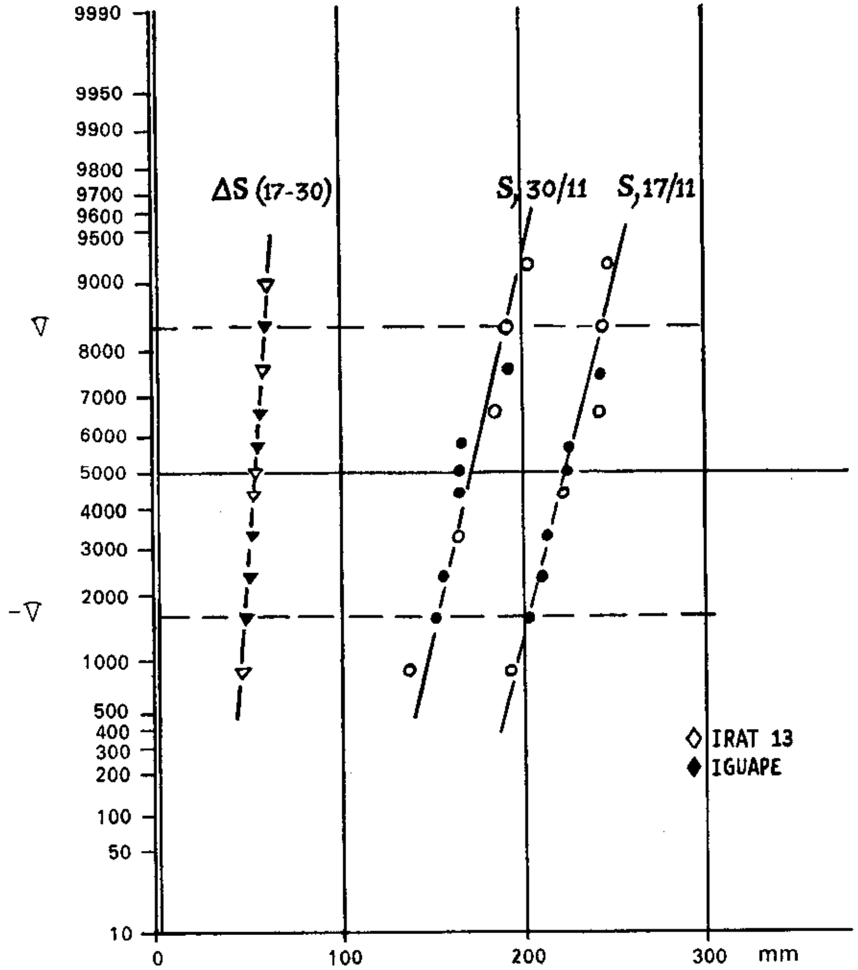


Fig. 9. Cumulative probability distribution of the values of the total soil water storage in the first metre measured at two dates (Nov. 17 and Nov. 30, 1980), and of the total change of water-storage during the interval, obtained at Bouake, Ivory Coast, for two cultivars of rainfed rice (IRAT 13, and Iguape CATETO) from 11 neutron access tubes (from Kalms *et al.* 1981).

An important problem for the practitioner is the determination of the number of measurements required for the estimation of the mean within a given uncertainty band around its real value.

For a normal distribution, the problem is easily solved as the variance of the estimation of the mean $\sigma(\bar{X})$ is related to the variance of the distribution

by $\sigma(\bar{X}) = \sigma(X)/\sqrt{n}$, where n is the number of measurements. Knowing $\sigma(X)$, it is therefore easy to determine the n necessary to obtain a certain value of $\sigma(\bar{X})$. $\sigma(\bar{X})$ obviously approaches 0 as n approaches infinity.

For a log-normal distribution, the problem is much more difficult as the geometric mean and the standard deviation are *non-consistent* (functions of the shape of the distribution, via its dissymetry) and biased (for σ, which is also a function of n). Landwehr (1978) presented an in-depth study of the properties of this type of distribution, especially the influence of n on the estimation of the mean and the errors that result if a normal distribution is used. The influence of the dissymetry factor is very clear. The tables made by Sichel (1966) and presented by David (1977, pp. 36, 38) are perhaps of greater practical value and can be used to determine estimation factors for a given population and variance to obtain a mean within 90% of its true value at probability levels of 90 and 95%.

The problem can also be dealt with by a purely numerical approach using the generation of a large number of samples by Monte-Carlo simulation taken at random from the distribution function, and studying the statistical properties of the solution according to its size. Fig. 7b shows the results of such a calculation presented by Imbernon (1981) for drainage flux values obtained at Bambey. It can be seen from this figure that at least 5 measurements are necessary (corresponding to a mean flux estimation between 9 and 23 mm/day for a 68% confidence interval) and it is of no use to increase the sample size above 20 measurements (corresponding to a mean flux estimation between 12 and 18 mm/day).

Hajrasuliha *et al.* (1980) used the same type of analysis to show that about 50 measurements were necessary to estimate the mean salinity to within ±20% (with a probability of 68%) for a 150 ha site in southwestern Iran, while the initial exploratory program involved 232 samples. Warrick and Amoozegar-Fard (1979) have presented results from another such Monte-Carlo analysis which clearly show that for a given sample size (100) during a drainage test, the estimation of the mean varies with time (and thus with the flux value). The estimation varies from ±50% to ±14% (for a 95% probability) while the mean flux varies from 32 to 0.44 mm/day with a variance going from 6.27 to 0.073.

The main disadvantage of the Monte-Carlo type methods is the difficulty in generalizing, as each result represents a particular distribution. This type of analysis is, however, extremely useful in preliminary explorations and defining the sample size that meets the needs of the planner. For a given development project it is in my opinion essential to determine either the acceptable uncertainty band for the estimation of the mean, or the maximum number of measurements that is practically possible and then determine the feasibility of the project.

Thus, the work carried out over the last 4 or 5 years clearly shows that a major effort must be made before any project to characterize a site is undertaken. A single field measurement is never representative of the average phenomenon. Owing to the strong log-normal character of the distributions that define the structure of certain parameters, a relatively high sample number is required (order of 10). Several points should be emphasized:

1. In order to simplify the procedures, it is possible to look for correlations between certain hydraulic parameters and the easy-to-measure structural properties (porosity) or textural properties (pore size) (Imbernon 1981). Initial tests involving samples for pore size analysis and determination of water content per weight along one or two diagonal transepts, taken at regular intervals of about 10 metres, can be used for a simple definition of the autocorrelation range and the essential distribution functions.

2. Results from Senegal and the Ivory Coast indicate that in certain cases a point may exist that characterizes the mean behaviour of the parcel (Vachaud *et al.* 1982). The location of this point comes from a systematic study of the cumulative probability diagrams. Such behaviour can greatly simplify the procedures and serve as a guide for the exploratory program.

3. Preliminary testing can be used to determine in advance the chances of success for an exploratory program, especially for comparative tests. The natural variance of water content can be such that possible differences in water consumption for different cultivars are less than the significant change of soil water storage (Kalms *et al.* 1981). In any case, classical Fisher test type comparative studies with n treatments (n approximately 6) and n repetitions, are in my opinion doomed to fail in advance, at least concerning the measurements in the soil. This is due to the sample size necessary for each treatment in order to obtain representative estimates of the mean value of the uncertainty bands. For the same number of measurements, a very large gain in information will be obtained by decreasing the number of treatments and increasing accordingly the number of repetitions far enough to avoid autocorrelation.

Extrapolation by Simulation

Simulation modelling can be considered as the preferred tool in development project work as it can be used to simulate different strategies (concerning treatments), leading to a fast determination of their influence without heavy exploratory programs. This is, however, only possible if the model is an accurate representation of the physical processes involved and of the characteristics of the media.

If we limit ourselves to the basic physical phenomena of water transfer in porous media (infiltration, drainage, groundwater recharge and, under certain conditions, solute flow), numerical models which represent well the physical processes are now available. This is probably the soil physics field which has witnessed the most progress over recent years. The review report by Molz *et al.* (1979) fully describes the state of the art.

Attempts to apply these models to field conditions have, however, up to present been almost total failures. This is due essentially to two reasons:

Difficulties in taking into account the stochastic characteristic of the transfer coefficients, and of the initial and boundary conditions in a deterministic model.

Difficulties in simulating the complex behaviour of the soil-plant-atmosphere system in the presence of growing vegetation.

I will not however finish on a pessimistic note as there are at present several lines of research in soil physics, hydrology and geohydrology which have not yet led to field applications but which have been the subject of methodological studies which may soon be exploitable.

One approach, introduced by Freeze (1975), uses a Monte-Carlo method to take into account the effect of spatial heterogeneity by random sampling of the variables from a known distribution function to obtain a mean value from all the partial solutions. This type of approach was successfully applied to the study of dispersion in aquifers (Tang and Pinder 1979). In hydrology, Smith and Hebbert (1979) studied the effect of the stochastic input (rainfall) and the influence of the spatial heterogeneity of the soil on runoff. More recently, this technique was used by Amoozegar-Fard *et al.* (1981) to study solute transport in a porous medium as part of a sensitivity study on the uncertainty in values of water velocity and of the diffusivity coefficient for the prediction of solute transport. The use of a stochastic velocity distribution led to a solution which was very different from the one obtained using a deterministic value, especially for the time for the arrival of solute at a given depth.

Another approach, which is in my opinion more easily applicable in the field, especially in sensitivity studies, is based on the *scaling factor* theory and on the coalescence of the variability of the $h(\theta)$ and $K(\theta)$ relationships towards a mean relationship characteristic of the site. As opposed to the preceding method, the simulation model is based on the solution of a purely deterministic scaled variable equation. The distribution function of the scaling factor is used to obtain the real characteristic value (pressure h) from the scaled mean characteristic value (pressure $\bar{h}$) for each measurement point. This method was developed by D.R. Nielsen, A.T. Warrick and their colleagues, and is presented in detail in a paper by Simmons *et al.* (1979). It represents one of this decade's most important contributions to soil physics. As an illustration, this technique was used on the 22 infiltration tests carried out on the Bambey site in Senegal (Imbernon 1981) and Fig. 10 shows both the infiltration curves obtained at each basic site and the scaled infiltration curve for the site.

Examples showing the possibilities of this method were recently presented by Warrick and Amoozegar-Fard (1979) for the calculation of simulated infiltration components, for a 150-ha site already characterized by Nielsen *et al.* (1973). Two important results should be pointed out:

For a constant head infiltration condition ($h = 0$) on the entire site,

Fig. 11a shows the infiltration rate and the uncertainty band for 25% (lower curve) and 75% of the probable values.

For gravity drainage at the same site and under the same conditions, Fig. 11b shows the drainage outflow curves (mean, 25% and 75%) simulated for a depth of 180 cm.

The same type of calculation was also presented by Sharma and Luxmoore (1979); however, their objective was more ambitious, involving a simulation of the entire hydrologic cycle (including plants), with an artificial value of the scale factor distribution function. The results can only be considered in relative terms but nevertheless show the high sensitivity of the water balance terms to the coefficient of variation of the scaling factor distribution. All else being equal (mean values, initial and boundary conditions) the runoff can vary from 1 to 6 when the coefficient of variation of the scale factor on a site goes from 0.3 to 1.5.

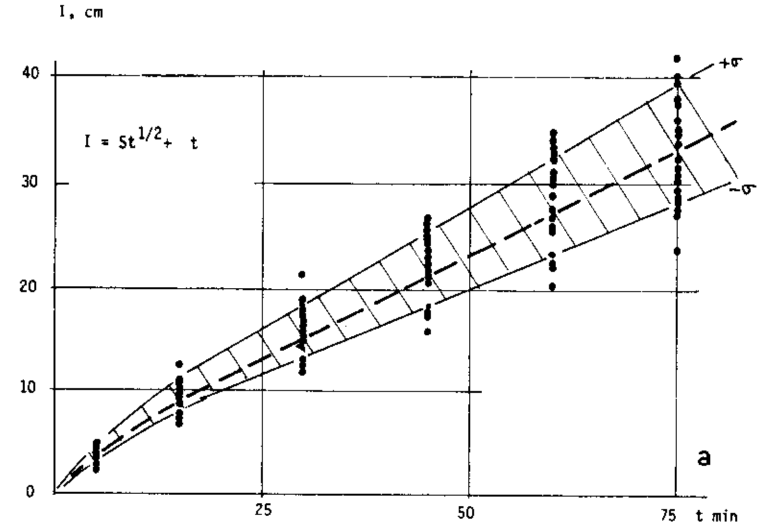
Such results clearly show that numerical modelling and the characterization of parameters in the field represent two indispensable and highly related tools. Although the models presently available could be further simplified, they can play a major role in project dimensioning as long as realistic parameters are used. By means of sensitivity studies, these models can also be used to determine parameters to be measured along with their number and the importance of the quality of the measurements.

In my opinion, validation studies are urgently required and should be carried out as soon as possible on test zones in order to determine the possibilities and the limits of these methods.

In conclusion, soil physics research in the 1980s has entered a phase where we can be optimistic about the possibilities of finding practical solutions to water resource development problems.

Any fieldman is much more aware of the complexity of the physical medium than the soil physicist. Confronted by the impossibility of proposing a solution for these complex phenomena, the physicists at first devoted themselves, in the laboratory, to understanding the deterministic aspects of the transfers and to developing simplified but accurate models, while the practitioners continued to use empirical methods which were sufficient for a particular site but difficult to extrapolate. The contribution of geostatistics was an essential catalyst in the evolution of soil physics. I have tried to show in this report that, thanks to these contributions, we can now quantify the complexity of the natural environment, characterize the independent variables, determine the validity of the measurements, extend the results from point values to mean values, optimize sampling and finally make models capable of simulating the processes involved.

A great deal of research is still required at the laboratory level, concerning the study of the soil-plant-atmosphere relationships, the processes involved in the decomposition of organic matter and denitrification, the kinetics of



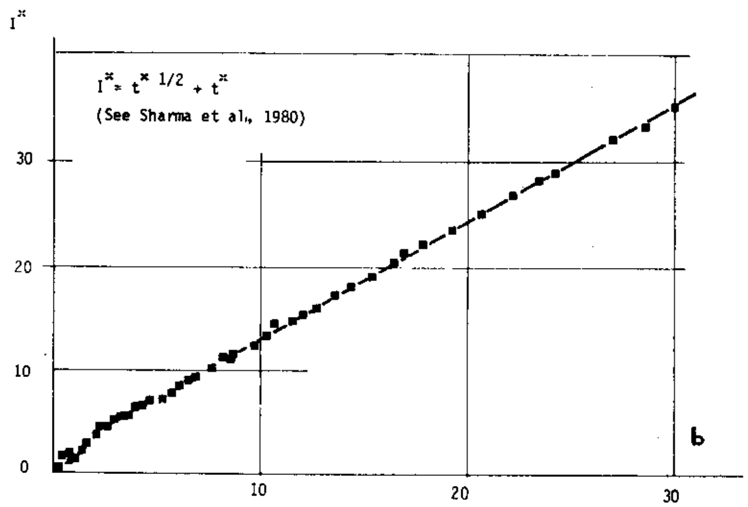


Fig. 10a. Infiltration measurements obtained in the site of Bambey from 24 double ring infiltration tests on 1 ha, together with the mean value, and the domain of uncertainty with $p = 0.68$.

Fig. 10b. Reduced infiltration curve from Fig. 9.a, using the model of Sharma *et al.* 1980. (from Imbernon, 1981)

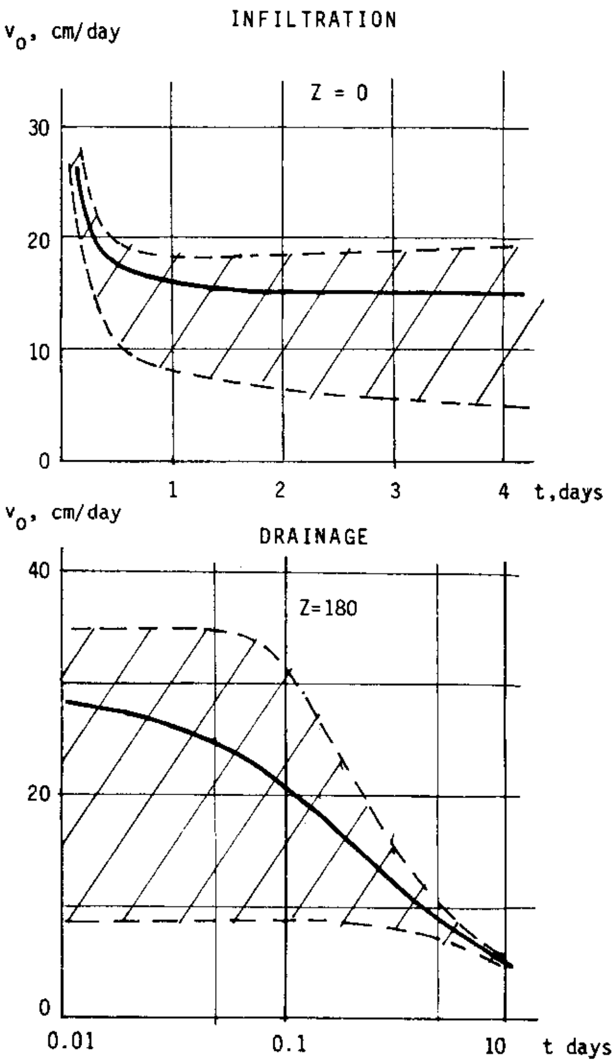


Fig. 11. Estimation of mean values of velocity, and of the 25-75 percentile range computed from the spatial characteristics of the Panoche clay loam under:
— constant flux infiltration at Z = 0
— gravity drainage at Z = 180 cm
by courtesy of A.T. Warrick, 1979

exchanges between soil and solution, etc. In the quantitative domain of water resources, research can be made operational in a very short time if enough measurements are available to reinforce or invalidate present theories. For this, it is imperative that researchers leave the laboratories and put their potential to work for the practitioners and planners by attacking the context of the field reality.

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