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Spatial Analysis of Soil Surface Hydraulic Properties: Is Infiltration Method-Dependent?

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ABSTRACT

The management of irrigated agricultural fields requires reliable information about soil hydraulic properties and their spatio-temporal variability. The spatial variability of saturated hydraulic conductivity, K_s and the alpha parameter $\alpha_{vG\text{-}2007}$ of van Genuchten equation has been revisited on an agricultural loamy soil after a 17-year period of repeated conventional agricultural practices for tillage and planting. The Beerkan infiltration method and its algorithm *BEST* were used to characterize the soil through the van Genuchten and Brooks and Corey equations. Forty field measurements were made at each node of a 6×7.5 m grid. The hydraulic properties and their spatial structure were compared to those performed in 1990 on the same field soil, through the exponential form of the hydraulic conductivity given by the Gardner equation, using the Guelph Pressure Infiltrometer technique. No significant differences in the results obtained in 1990 and 2007 were observed for both the particle-size distribution and dry bulk density. Also, both the K_s and $\alpha_{vG\text{-}2007}$ values were still better described by the lognormal distribution. The mean value of $\alpha_{vG\text{-}2007}$ was found to be identical to that of $\alpha_{G\text{-}1990}$, while that of $K_{s\text{-}2007}$ was significantly smaller than that of $K_{s\text{-}1990}$. In contrast to the Gardner equation, the van Genuchten/Brooks and Corey expression was found to better represent a possible gradual increase of water filling the porosity, which is more representative for a well-graded particle-size distribution of a loamy soil. The two parameters were autocorrelated up to about 30 m and 21 m, respectively, as well as spatially positively correlated together within a range of 30 m. Despite the difference in the mean values of K_s between the two studies, the spatial structures were similar to those found in the 1990 experiment except for the covariance sign. The similarity in autocorrelation ranges may indicate that the spatial analysis of soil hydraulic properties is independent of the infiltration methods (i.e., measurement of an infiltration flux) used in the two studies, while the difference in the covariance sign can be referred to the use of two different techniques of soil hydraulic parameterization. The spatial patterns of hydraulic parameter variations seem to be temporally stabilized, at least within the agro-pedo-climatic context of the study., This may be referred to the textural properties and to the soil structural properties. The latter are constantly renewed by the cyclic agricultural practices. However, further experiments are needed to strengthen this result.

Keywords: Soil hydraulic properties; Beerkan infiltration method; Loamy soil; Spatial variability; Time stability.

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INTRODUCTION

Irrigated agricultural soil management requires reliable information about soil hydraulic properties and their spatio-temporal variability. This variability is inherent in nature due to geologic and pedologic factors forming field soil. Within a particular agricultural field, changes in soil structure and consequently in soil hydraulic properties, may occur due to different tillage, irrigation, planting and harvest/residues managements (Mapa et al., 1986; Angulo-Jaramillo et al., 1997; Cameira et al., 2003; Iqbal et al., 2005; Mubarak et al., 2009). These changes also may occur during the cropping season and from year to year depending on climatic conditions (Strudley et al., 2008).

However, several studies which dealt with some soil hydraulic properties like soil water content and soil matrix potential and their spatio-temporal variability (Vachaud et al., 1985; Kachanoski et al., 1985; Kachanoski and de Jong, 1988; van Wesenbeeck and Kachanoski, 1988; Jaynes and Hunsaker, 1989; Goovaerts and Chiang, 1993; Chen et al., 1995; van Pelt and Wierenga, 2001) have shown that, although these properties change with time and with location in the field, the pattern of its spatial structure does not change with time when the observations are ranked or scaled against the field mean value. The implication is that the order of soil water content of different points will not change with time at a certain probability. This phenomenon has been called time stability or temporal stability (Vachaud et al., 1985), temporal persistence (Kachanoski and de Jong, 1988), or rank stability (Tallon and Si, 2003) in soil moisture spatial patterns. Most studies have attributed such phenomenon to soil texture and structure and topography. Although temporal stability of spatially measured soil water content and soil matrix potential probability density function has been demonstrated, it has not been shown for the spatial structure of variations in other soil hydraulic properties like saturated hydraulic conductivity.

Hydraulic characteristic curves of the soils, that are, the water retention, $h(\theta)$, and the hydraulic conductivity, $K(\theta)$ or $K(h)$ curves, where θ is the volumetric water content, can be represented and appropriately parameterized in analytical or numerical models. The most common expressions are the van Genuchten (1980), Brooks and Corey (1964), and Gardner (1958). Therefore, modelling soil water transfer has increased the demand for accurate measurements of soil physical and hydraulic properties. Several methods have been developed to determine soil hydraulic characteristic functions. Some of them require only readily available information, such as particle-size distribution or simple field measurements (Jarvis et al., 2002). Some others require a full experimental determination of the water retention curve $h(\theta)$ and the hydraulic conductivity $K(\theta)$ using laboratory apparatus (Raimbault, 1986; Mallants et al., 1997). Some methods have been based on field experiments such as infiltration tests (Simunek et al., 1998; Angulo-Jaramillo et al., 2000; Jacques et al., 2002). These are usually performed by imposing a given water pressure head through either single rings or disc infiltrometers, depending on the sign of the pressure head applied at the soil surface (Angulo-Jaramillo et al., 2000; Haverkamp et al., 2006). The field methods are generally less expensive and time-consuming and provide a better picture of soils than the laboratory methods. However, comparison of hydraulic properties determined by different measurement methods has revealed significant differences for fine-textured soils, although better agreement often has been found for relatively coarse-textured soils (Roulier et al., 1972). Mallants et al. (1997) derived a similar conclusion about both water retention and hydraulic conductivity functions using three different means of in-situ field and ex-situ laboratory techniques. Lassabatère et al. (2006) compared their fitting algorithm of infiltration data (BEST, Beerkan Estimation of Soil Transfer parameters through infiltration experiments) with four different fitting methods referred to as cumulative linearization (CL, Smiles and Knight, 1976), derivative linearization (DL), cumulative infiltration (CI) and infiltration flux

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(IF) (Vandervaere et al., 2000). They found that for the same experimental data, BEST provided acceptable estimations of hydraulic parameters leading to a complete characterization of hydraulic characteristic curves, and the other methods (CL, DL, CI, and IF) differed in the estimations of sorptivity and hydraulic conductivity, as shown by the high coefficients of variation. The coefficients of variation were higher for hydraulic conductivity than for sorptivity showing that the hydraulic conductivity estimation is method dependent. Because the unsaturated soil hydraulic functions are necessary inputs for numerical simulation models used to evaluate alternative soil and water management practices, there is a need to continuously assess the accuracies and limitations of measurement techniques and fitting methods to determine the soil-hydraulic properties.

Vauclin et al. (1994) examined the spatial variability of saturated hydraulic conductivity, K_s and the alpha parameter, α_G, of the exponential form of the Gardner relationship. The field experiment was performed on a bare agricultural soil with a loamy texture, by carrying out surface infiltration measurements with the Guelph Pressure Infiltrometer technique (Reynolds and Elrick, 1990) in July 1990. Thirty two measurements were made at each node of a 4×8 m grid. They showed that geometric mean values of both K_s and α_G were found to be 2.95×10^{-5} m s^{-1} and 11.7 m^{-1} with coefficients of variation (CV) of 36% and 48%, respectively. In addition, they found that these two parameters were autocorrelated up to about 25 m and to 20 m, respectively and spatially correlated together within a distance of 24 m. As repeated conventional agricultural practices for tillage and planting have been applied after that field experiment, there is a need for examining the effects of this medium-term agricultural management on soil hydraulic properties and their spatial structures to know if a temporal stability of these spatial structures can be demonstrated. The present study proposes the use of the Beerkan infiltration method (Haverkamp et al., 1996) to provide the soil hydraulic properties using a simple *in situ* single ring infiltration test. The originality of the Beerkan method lies in its inverse procedure, which uses a specific identification algorithm based on soil physical constraints. The algorithm noted as *BEST* takes into account the van Genuchten (1980) equation with the Burdine (1953) condition for $h(\theta)$ and the Brooks and Corey (1964) equation for $K(\theta)$ (Lassabatère et al., 2006).

The objectives of the present field study are twofold: (i) to examine the effects of 17 years of repeated agricultural practices on soil hydraulic properties and their spatial variability at the site studied by Vauclin et al. (1994), using the Beerkan infiltration method, and (ii) to determine whether the infiltration tests are sensitive to the methods used to describe the spatial and temporal variability of the agricultural field.

(1) analysis of the spatial variability
(2) comparison with the results derived from a spatial analysis performed 17 years ago using another infiltration method to verify whether or not the infiltration tests are sensitive to the methods used to describe the spatial and temporal variability of the agricultural field.

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MATERIAL AND METHODS

Studied Site

As cited in Vauclin et al., 1994, the field is located at the Domain of Lavalette at the Cemagref Experimental Station in Montpellier, France (43°40'N, 3°50'E, altitude 30 m). The average annual rainfall is 790 mm year^{-1} (1991-2006). Repeated conventional agricultural practices i.e., annual soil tillage in fall, soil preparations in spring during which the soil was ploughed and then the 8 cm topsoil was grinded to very small size aggregates (less than 2 mm size) and related management practices (seeding, fertilization, pesticides) are mechanically applied since 1990. The rotation of crops cultivated in the field was: sorghum (*Sorghum bicolor*) and soybean (*Glycine max*) (1990-1993), sunflower (*Helianthus annus*) (1994-1995), winter wheat (*Triticum aestivum*) (1996), maize (*Zea mays*) (1997-2002), winter wheat (*Triticum aestivum*) (2003-2006) and finally maize (*Zea mays*) (2007).

Infiltration Measurements

Five measurements were carried out at 6 m intervals in each of eight rows, 7.5 m apart, giving a total of forty measurement sites, taken during five days in May 2007 without irrigation and rainfall. At each site, infiltration test was conducted using a 65 mm-radius cylinder driven approximately 1 cm into the soil to avoid lateral water losses (Fig. 1). A fixed volume of water (100 ml) was poured into the cylinder at time zero, and the time needed for the infiltration of this known water volume was recorded. When the first volume was completely infiltrated, a second known volume of water was added to the cylinder, and the time required for it to infiltrate is added to the previous time. The procedure was repeated for a series of several volumes until apparent steady-state flow regime was reached, i.e. until three consecutive infiltration times were identical, and cumulative infiltration was recorded (Haverkamp et al., 1996; Lassabatère et al., 2006). A soil sample of about 200 cm^3 using a core sampler with core dimensions of 5 cm height by 7.3 cm diameter, was collected at each site to determine both the soil dry bulk density (ρ_d) (Grossman and Reinsch, 2002) and the initial soil gravimetric water content. The latter was converted into volumetric soil water content (θ_0) through the dry bulk density. Another soil sample was collected for particle-size analysis, which was determined for the fine soil fraction (< 2 mm). Particle-size distribution was analyzed by means of the sedimentation method on eight composed soil samples (Gee and Or, 2002). Saturated volumetric water content (θ_s) was calculated as the total soil porosity considering that the density of the solid particles is 2.65 g cm^{-3}.

In the study of Vauclin et al. (1994), the surface infiltration measurements were performed with the Guelph Pressure Infiltrometer technique (Reynolds and Elrick, 1990) with thirty two measurements made at each node of a 4×8 m grid, taken during a four-day period in July 1990 without irrigation and rainfall too. The pressure infiltrometer connected to a water-filled reservoir was driven approximately 5 cm into the soil using a wooden hammer. The infiltration measurements were carried out for three values of ponded water pressure heads. The first positive head, H_1, was set at 6 cm and early-time readings were made at 20 s and at each 10 s afterwards for about 2 min and then at 20 s intervals thereafter to about 4 min. Readings of the water level in the reservoir were then continued at 1 min intervals for about 20 min in order to obtain the steady intake rate, Q_1, at H_1=6 cm. The air tube was then raised and H_2 set at 16 cm. A smaller time interval of about 5 to 7 min was needed in order to establish the steady state intake rate, Q_2. Similarly for Q_3 at H_3=25.5 cm. A small soil sample was taken at each measurement site to obtain the initial soil gravimetric water content. Also, another soil sample was taken posterior each infiltration test from within the ring area for determining the field saturated gravimetric water content. Both initial and saturated gravimetric water contents were converted into volumetric ones through the dry bulk density.

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Soil Hydraulic Characterization

In the present work, the BEST algorithm was used for determining soil hydraulic properties (Lassabatère et al., 2006) through the van Genuchten equation for the water retention curve, $h(\theta)$, (Eq. 1a) with the Burdine condition (Eq. 1b) and the Brooks and Corey relation (Eq. 2) for the hydraulic conductivity curve, $K(h)$, (Burdine, 1953; Brooks and Corey, 1964; van Genuchten, 1980) :

$$m = 1 - \frac{2}{n} \qquad [1a]$$

$$[1b]$$

$$K(h) = K_s \left(\frac{\theta}{\theta_s} \right)^{\eta} \qquad [2]$$

$$\theta(h) = \theta_s \left(1 + (\alpha_{vG} |h|)^n \right)^{-m}$$

where n and m are the dimensionless shape parameters of the water retention curve and α_{vG} is simply called the alpha parameter of van Genuchten model (m^{-1}). K_s is the saturated hydraulic conductivity ($m\ s^{-1}$) and η (-) is the shape parameter of the hydraulic conductivity relationship. Consequently, the representation of the hydraulic properties makes use of five parameters: θ_s, n, α_{vG}, K_s, η. Following Haverkamp et al. (1996), n, m and η are assumed to be dominantly related to the soil texture, while the others are supposed to mainly depend on the soil structure.

The BEST algorithm estimates the shape parameters from particle-size distribution by classical pedotransfer functions. The other scale parameters are derived by modelling the 3D infiltration data. Following Haverkamp et al. (1994), the 3D cumulative infiltration $I(t)$ and the infiltration rate $q(t)$ can be described by the explicit transient (Eq. 3a and b) and steady-state (Eq. 3c and d) equations:

$$I(t) = S\sqrt{t} + \left[A\, S^2 + B\, K_s \right] t \qquad [3a]$$

$$q(t) = \frac{S}{2\sqrt{t}} + \left[A\, S^2 + B\, K_s \right] \qquad [3b]$$

$$I_{+\infty}(t) = \left[A\, S^2 + K_s \right] t + C\, \frac{S^2}{K_s} \qquad [3c]$$

$$q_{+\infty}(t) = q_{+\infty} = A\, S^2 + K_s \qquad [3d]$$

where S is the sorptivity ($m\ s^{-1/2}$), A, B and C are related to the Brooks and Corey parameters and to the initial volumetric water content (see Lassabatère et al., 2006 for more details). The BEST code uses equivalent equations obtained by the replacement of the saturated hydraulic conductivity K_s by its function of sorptivity S and steady-state infiltration rate $q_{+\infty}$ (Eq. 3d) into equations 3a and 3b:

$$I(t) = S\sqrt{t} + \left[A(1-B)\, S^2 + B\, q_{+\infty} \right] t \qquad [4a]$$

$$[4b]$$

$$q(t) = \frac{S}{2\sqrt{t}} + \left[A(1-B)\, S^2 + B\, q_{+\infty} \right]$$

The saturated hydraulic conductivity is calculated from the steady-state infiltration rate and prior estimation of sorptivity by fitting transient infiltration data on the two-term infiltration equations (Eq. 4) using a data subset for which the transient two-term infiltration equations are valid. The maximum time t_{max} for which transient expressions can be considered valid is defined as follows:

$$[5]$$

$$t_{\max} = \frac{1}{4(1-B)^2} t_{grav}$$

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where $t_{grav} = (S/K_s)^2$ is the gravity time defined by Philip (1969).

The α_{vG} is then estimated from the sorptivity S and the saturated hydraulic conductivity K_s by the following relation:

$$\alpha_{vG}^{-1} = \frac{S^2(\theta_0, \theta_s)}{c_{p(vG)}(\theta_s - \theta_0)(K_s - K_0)} \quad [6]$$

where $c_{p(vG)}$ is a function of the shape parameters for the van Genuchten (1980) water retention equation (see Haverkamp et al., 2006 and Lassabatère et al., 2006). $K_0 = K(\theta_0)$ is the initial hydraulic conductivity calculated by Eq. (2). The reader is referred to the study of Lassabatère et al. (2006) that described the main characteristics of BEST algorithm coded with MathCAD 11 (Mathsoft Engineering and Education, 2002).

While in the 1990 experiment, a detailed description of the soil hydraulic parameterization can be found in Vauclin et al. (1994), only a brief presentation is given herein. The empirical representation of $K(h)$ relationship as proposed by Gardner (1958) was used:

$$K(h) = K_s \exp(\alpha_G\ h) \quad \text{with } 0 < \alpha < \infty \text{ and } h \le 0 \quad [7]$$

where α_G (m^{-1}) is simply called the alpha parameter calculated as:

$$\alpha_G = \frac{K_s - K_0}{\Phi_m} = |h_f|^{-1} = \lambda_c^{-1} \quad [8]$$

where Φ_m is the matrix flux potential ($m^2\ s^{-1}$). $K_0 = K(h_0)$ was considered small enough to be neglected relative to K_s. λ_c is the capillary length (Philip 1985) and $|h_f|$ is the effective wetting front potential for Green and Ampt (1911) infiltration. Thus, the three parameters α_G^{-1}, λ_c and $|h_f|$ are equivalent and represent a single parameter.

Both K_s and Φ_m that together determine the steady flow, also define the early-transient cumulative infiltration $I(t)$ (m) under a ponded head H:

$$I(t) = S_H.t^{0.5} \quad [9]$$

where S_H is the "ponded" sorptivity which is related (Philip 1985) to the zero-ponded sorptivity S_m through the following relationship:

$$S_H = S_m(1 + 2b.\alpha_G.H)^{0.5} \quad [10]$$

Following White and Sully (1987), S_m is calculated as:

$$S_m = \left(\frac{\Delta\theta.\Phi_m}{b}\right)^{0.5} \quad [11]$$

where $\Delta\theta$ is the difference between the saturated and initial volumetric water contents. The constant b can be assigned a value of 0.55 (White and Sully, 1987).

Eq. (9) was used to obtain S_H from the early-time of the experimental data. A linear regression of the first 10 to 15 points, chosen from a graph $I(t)$ vs $t^{0.5}$, was used to calculate the slope and thus S_H. Eq. (10) was then used to derive S_m. Steady-state flow rates Q ($m^3\ s^{-1}$) for the pressure infiltrometer are given by Reynolds and Elrick (1990):

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$$Q = \pi r^2 K_s + \left(\frac{r}{G}\right)\left(K_s H + \Phi_m\right) \qquad [12]$$

with G=0.316 d/r + 0.184 and where d is the depth of insertion (0.05 m) of the infiltrometer ring into the soil and r=0.05 m is the radius of the ring. K_s, Φ_m and α_G were calculated by Eq. (12) using non-linear regression on the three steady-state flow measurements.

Philip (1985) defined the capillary length [m] noted herein as:

$$\lambda_c = \frac{1}{K_s - K_0} \int_{h_0}^{0} K(h)\, dh$$

[13]

$$= \frac{1}{K_s - K_0} \int_{\theta_0}^{\theta_s} D(\theta)\, d\theta$$

where $D(\theta) = K(\theta)dh/d\theta$ is the diffusivity [$m^2\ s^{-1}$]. He found that the capillary length is equal to α_G^{-1} for the Gardner soil as noted above (Eq. 8). White and Sully (1987) found that the capillary length is expressed as a function of the sorptivity with b= 0.55 for the Gardner soil:

$$\lambda_c = \frac{b.S^2(\theta_0, \theta_s)}{(\theta_s - \theta_0)(K_s - K_0)} \qquad [14]$$

Haverkamp et al. (2006) presented a generalized form of the capillary length with proportionality constant c_p which replaced the term (1/b) as:

$$\alpha_h = \frac{S^2(\theta_0, \theta_s)}{c_p(\theta_s - \theta_0)(K_s - K_0)} \qquad [15]$$

c_p depends on the functional relationships chosen to describe the soil hydraulic characteristics. It should be mentioned that Eq. (15) can apply for any soil hydraulic functional relationship.

When describing unsaturated water flow subject to a given set of initial and boundary conditions, the water flow behavior of the soil should be independent of the choice of the soil hydraulic functional relationships. This can be guaranteed by applying the sorptivity criterion as in Haverkamp et al. (2006) to calculate the correct c_p value. So, when dealing with field soils, c_p is entirely defined by textural soil parameters (Haverkamp et al., 2006). By taking into account the value of c_p calculated for the van Genuchten equation as given in Lassabatère et al. (2006), Eq. (3) leads to consider that α_{vG} represents the reciprocal of the capillary length for the van Genuchten soil. This concept of capillary length led us to compare the alpha parameter of the Gardner equation, α_G, estimated in 1990 to the alpha parameter of the van Genuchten relationship, α_{vG}, obtained in 2007.

Statistical and Geostatistical Analysis

Measured variables, namely saturated hydraulic conductivity and the alpha parameter, were analyzed by standard statistics to obtain their mean and coefficient of variation values. The normality of data frequency distribution was tested using both the Kolmogorov–Smirnov test and the values of the skewness (g_1) and kurtosis (g_2) coefficients. Following Vauclin et al. (1982), the latter coefficients are expressed as:

$$g_1 = \frac{n}{(n-1)(n-2)} \cdot \frac{m_3}{(m_2)^{3/2}} \qquad [16a]$$

$$g_2 = \frac{n(n+1)}{(n-1)(n-2)(n-3)} \cdot \frac{m_4}{(m_2)^2} \qquad [16b]$$

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where m_2, m_3 and m_4 are the 2nd, 3rd and 4th moments of the distributions, n being the number of measurements.

The Student's t-variables corresponding to g_1 and g_2 have also been calculated as follows:

$$t_1 = \frac{g_1}{s_{g_1}} \quad \text{and} \quad t_2 = \frac{(g_2 - 3)}{s_{g_2}} \qquad [17]$$

where s_{g1} and s_{g2} are estimated by the expressions:

$$s_{g_1}^2 = \frac{6n(n-1)}{(n-2)(n+1)(n+3)} \qquad [18a]$$

$$s_{g_2}^2 = \frac{24n(n-1)^2}{(n-2)(n-2)(n+3)(n+5)} \qquad [18b]$$

The analysis of significance using the Student's t-test at the 95% level of probability was also used to compare the soil hydraulic parameter K_s and the alpha parameter, obtained in the present work with those found in the 1990 experiment.

Geostatistics was used to quantify the spatial dependence and spatial structure of the two parameters K_s and α_{vG} for comparison purposes with the 1990 results. The spatial structure of these two parameters was identified by the semivariogram using the Variowin program Model (Pannatier, 1996). The experimental semivariogram $\gamma(l)$ was estimated as:

$$\gamma(l) = \frac{1}{2N(l)} \sum_{i=1}^{N(l)} [z(r_i) - z(r_i + l)]^2 \qquad [19]$$

where $N(l)$ is the number of pairs separated by lag distance, l; $z(r_i)$ and $z(r_i+l)$ are measured values at locations r_i and r_{i+l}, respectively. Experimental semivariograms were normalized by dividing each semivariance by the experimental variance value (Vieira and Gonzalez, 2003).

The cross-semivariogram was also considered in order to investigate the spatial correlation between the two parameters K_s and α_{vG} and it was calculated as (Vauclin et al., 1983):

$$\gamma_{12}(l) = \frac{1}{2N(l)} \sum_{i=1}^{N(l)} [z_1(r_i) - z_1(r_i + l)][z_2(r_i) - z_2(r_i + l)] \qquad [20]$$

where $z_1(r_i)$ and $z_2(r_i)$ are the values of the Ln K_s and Ln α_{vG} at location r_i. The corresponding cross semivariogram was normalized by the product of the experimental standard deviations of the two parameters.

As in the 1990 field campaign, both the normalized experimental values of semivariogram and cross-semivariogram were fitted on the theoretical spherical model:

$$\begin{aligned} \gamma^*(l) &= C_0 + C_1\left(1.5\frac{l}{a} - 0.5\left(\frac{l}{a}\right)^3\right) && \text{for } l \le a \\ &= C_0 + C_1 = C_2 && \text{for } l > a \end{aligned} \qquad [21]$$

where C_0 is the nugget effect, a is the range and C_2 is the sill. Eq. (21) was fitted to the experimental data by using least squares minimization with respect to a maximum separation distance which was restricted to less than one-half of the largest dimension of the field (Mulla and McBratney, 2002).

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RESULTS AND DISCUSSION

Analysis of Methods

Particle-size distribution analysis showed that the soil is classified as a loamy soil containing on average 43% sand, 40% silt and 17% clay. The values of coefficients of variation (CV) range from little (5.5%) to high (39.0%) according to the classification described by Wilding (1985) (Table 1). There was a slight increase in the sand percentage and a small decrease in both the silt and clay percentages compared to those found in the 1990 field study (35% sand, 45% silt and 20% clay). The mean value of soil dry bulk density (ρ_d) of the topsoil was estimated at 1.37 g cm^{-3} with little value of CV (4.8%) (Table 1) and was similar to that estimated in the 1990 experiment (1.4 g cm^{-3} with a CV value of 1.4%). The small mean value of ρ_d can be referred to the fragile structural porosity created by soil preparations in spring. The little value of CV can be due to the annual soil tillage and soil preparation which makes the surface soil layer homogeneous.

The analysis of particle-size distribution and the 3D infiltration modeling led to the full determination of the hydraulic shape and scale parameters using the Beerkan method and its fitting technique (BEST). Table 2 summarizes the statistical analysis of the hydraulic parameters.

The shape parameters of $h(\theta)$ and $K(\theta)$ were hardly variable at the field scale. This very low variability is consistent with the assumption that the shape parameters are dominantly related to the soil texture (Haverkamp et al., 1996). These estimated values are in agreement with common values published for a loamy soil (Haverkamp et al., 1997; Wosten et al., 1999; Lassabatère et al., 2006).

The Kolmogorov–Smirnov (K-S) test on original datasets of K_s and α_{vG} produced p-values of <0.001 and 0.03, respectively, and when the datasets were log-transformed, the p-values changed to 0.04 and 0.19, respectively. As the log-transformed data passed the K–S normality test, the null hypothesis (H0: the data are drawn from a normally distributed population) should be rejected. Moreover, based on student tests at a significance level of 0.05, the value of the skewness (g_1) indicated that the normal distribution (g_1=0) should be rejected for both K_s and α_{vG} (Table 2). This was not so evident following the kurtosis (g_2) coefficient. This can be referred to the limited number of data points used in the present work (40 measurements). Following Rao et al. (1979) several hundreds of observations would be needed to really decide an appropriate distribution function. Based on the Kolmogorov-Smirnov test and the values of the skewness coefficient, it seems that the lognormal distribution appears acceptable for the K_s and α_{vG} values. This is also in agreement with the 1990 experimental results. Similar results have been found by White and Sully (1992), Russo and Bouton (1992) and Mulla and McBratney (2002).

The geometric mean value of α_{vG} was estimated at 12 m^{-1} with moderate CV value (22%). For K_s, the geometric mean value was 8.4×10^{-6} m s^{-1} with high CV value (76%). Similar values of CV were reported in the literature (Vauclin, 1982). Saturated hydraulic conductivity and their related quantities such as infiltration rates were found to have a high statistical variability. Mulla and McBratney (2002) reported range of coefficients of variation of K_s from 48% to 352%. In the present study, the majority of physical and hydraulic parameters showed small and moderate CV values (<35%) according to the classification described by Wilding (1985). This could be due to the homogeneity in the soil preparations.

On the basis of the Student's *t*-test, the mean values of K_s estimated in the two studies were significantly different. The mean value found in the 1990 experiment ($K_{s\text{-}1990}$ = 2.95×10^{-5} m s^{-1}) was about four times higher than that obtained in the present study. While the mean

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value of the alpha parameter found in the 1990 experiment ($\alpha_{G\text{-}1990}$ = 11.7 m^{-1}) was similar to that of the present study (Table 2).

Analysis of the hydraulic conductivity curves *K(h)* which have been described by two different *in situ* infiltration methods with two different techniques of soil hydraulic parameterization, facilitates understanding the soil hydraulic behavior and comparing the results of the two experiments. Figure 2 shows two *K(h)* curves: (i) solid line obtained by using the Gardner equation with the two parameters estimated in the 1990 experiment and (ii) dashed line as generated by the two parameters of the 2007 study. The curve *K(h)* of Gardner showed a sudden change of unsaturated hydraulic conductivity close to a given value of water pressure head (about 0.4 m). Below this value, hydraulic conductivity sharply decreases. This hydraulic behavior corresponds to a coarse soil with a more open pore structure. The *K(h)* curve, plotted according to the parameter values found in the present study, is quite different. When the water pressure increases, hydraulic conductivity increases gradually. The van Genuchten/Brooks and Corey expression better represents a possible gradual increase of water filling the porosity, which is more representative for a well-graded particle-size distribution of a loamy soil (Table 1).

The difference in results of the hydraulic parameters may find its origin in the analysis of estimation methods used in both studies. In Vauclin et al. (1994) the first 10 to 15 points chosen from the infiltration curves, i.e. I vs. $t^{1/2}$ as the early-time experiment data, were employed to calculate the slope and thus the sorptivity using a linear regression. The subjective choice of the early-time and thus the complementary steady-state experimental data influenced the estimate of the alpha parameter (α_G). The estimated value of α_G was used in their study to calculate the value of K_s. As they have noted, the error in this calculation depends in large part on the error in the estimate of that parameter.

These problems were, in part, solved by the Beerkan method and its algorithm *BEST* that associates the analysis of both the transient and asymptotic regimes by using the very accurate explicit transient two-term and steady-state expansions given by Haverkamp et al. (1994). Indeed, *BEST* calculates in an iterative process the time for which transient expressions are valid as defined by Eq. 5. This improved the robustness of the estimation of the sorptivity (S) and the saturated hydraulic conductivity (K_s) as well as the subsequent estimate of the alpha parameter (α_{vG}). In addition, the difference may result from the total period of infiltration test. As a matter of fact, each test was operated for a period sufficiently long (about 2 hr) in the 2007 study compared to 1 hr in the 1990 experiment. This led to gain in precision in determining soil hydraulic parameters.

Temporal Stability of Geostatistical Parameters

With the overall number of observations (n=40), all the number of couples *N(l)* for each lag in both the semivariogram and cross semivariogram are much larger than the minimum (about 30 pairs of points per lag) required in the literature (Mulla and McBratney, 2002) as shown in Figs. 3, 4 and 5. This led the estimates of the experimental data of semi-variances to be confident.

The results of spatial variability analysis for Ln K_s and Ln α_{vG} indicate the existence of a spatial structure across the field (Figs. 3 and 4). The experimental values were fitted to a spherical model using Eq. (21). The corresponding values are reported in Table 3 against those found in the 1990 field campaign. The following comments can be made:

1) Semivariograms show small nugget effect, C_0, of about 0.13 and 0.03 for Ln K_s and Ln α_{vG}, respectively. Generally, the nugget parameter is a measure of the amount of variance due to errors in sampling, measurement, and other unexplained sources of variance and corresponds to the spatial variation occurring within distances shorter than the measurement test interval (6 m × 7.5 m) (Mulla and McBratney, 2002). However, the smallness in the C_0

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values found in the present study indicates the precision in the infiltration method and the relative homogeneity of the surface soil layer due to the annual soil tillage and soil preparation.

2) The ranges in lag distance of Ln K_s and Ln α_{vG} are about 30 m and 21 m, respectively. Within the range, the measurements of variable are correlated with each other.

3) The sill values, C_2, of Ln K_s and Ln α_{vG} are close to the unity as theoretically expected. The stability of semi-variances beyond the range highlights the lack of drift. This indicates that the dimensions of the field seem to be sufficient to describe the whole spatial structure of the two parameters.

4) K_s is strongly correlated to α_{vG} up to a distance of about 30 m (Fig. 5). This value is indeed comparable to the ranges found for the two parameters (Table 3). The nugget effect (C_0) of the cross semivariogram is close to zero. This is due to the fact that the two parameters are related to the same soil pore space. It should also be noticed that the sill C_2 (0.86) is close to the correlation coefficient value i.e., the normalized covariance value (0.62) as theoretically expected. The positive cross-semivariance value signifies a positive relationship between the two variables.

In the 1990 experiment, K_s and α_G were found to be autocorrelated up to about 25 m and to 20 m, respectively (Vauclin et al., 1994). Semivariograms in that study showed a relatively large nugget effect, C_0. The semivariogram of α_G showed a C_0 value larger than that of K_s (Table 3 and Figs. 3 and 4). This indicates that α_G is more fluctuating at short distances than K_s. The authors in that study referred that to a greater influence of macropores on α_G as compared to K_s. They found also that these two parameters were spatially negatively correlated together within a distance of 24 m.

Although the mean value of $K_{s\text{-}2007}$ is significantly different from the $K_{s\text{-}1990}$ results, we can compare the spatial structure found in the two studies because the geostatistical analysis deals with the variance between observations. The values of $C_{0\text{-}1990}$ were very larger than those of $C_{0\text{-}2007}$ for both the semivariogram and cross-semivariogram. This difference could be due to the use of two different calculation algorithms with two different models of hydraulic characteristic curves of the soil. Also, in contrast to the results found in 2007, those of the 1990 experiment showed that both the saturated hydraulic conductivity and the alpha parameter were spatially negatively correlated together (Fig. 5). The negative values of the cross semivariogram indicated that the value of α_G will be higher as much as the value of K_s will be smaller. As α_G^{-1} or α_{vG}^{-1} is a measure of the importance of capillary forces relative to gravity for water movement, the positive spatial correlation between saturated hydraulic conductivity and the alpha parameter observed herein seems to be in agreement with the physics of water infiltration in soils (see Eqs. 6 and 8).

This difference could be due the two different *in situ* infiltration methods with two different techniques of soil hydraulic parameterization used in the two studies. The infiltrometer used in the 1990 experiment was driven approximately 5 cm into the soil. It is expected that this large depth induced the water flow to be divided into two parts: a 1D flow for the early-time of the experiment and a 3D one for the rest of the time. This was not considered in the soil hydraulic parameterization. In addition, each infiltration test was operated for a relatively short period (about 1 hr). This short time with the subjective choice of the early-time and thus the complementary steady-state experimental data influenced the estimate of the alpha parameter (α_G) and consequently the calculation of the value of K_s.

For its part, the Beerkan method which was used for longer periods of infiltration time (about 2 hr per test), was performed with a simple ring inserted into soil to a depth of 1 cm neglecting the 1D water flow regime. Also, its algorithm *BEST* associated the analysis of the transient and asymptotic states by calculating the time for which transient expressions are

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valid. This probably improved the robustness of the estimation of sorptivity (*S*) and saturated hydraulic conductivity (K_s) and the subsequent estimation of the alpha parameter (α_{vG}).

However, the spatial structure of the two parameters shows no significant change over time. The range values of the 2007 field study for both the semivariogram and cross semivariogram are very close to those found in the 1990 experiment. This indicates that the spatial pattern of hydraulic parameter variations seems to be temporally stabilized under the agro-pedo-climatic context of the field study. Although there are only two dates of measurements, we can refer this time stability to a possible existence of a deterministic factor imposed by the dry bulk density and particle-size distribution which were found of the same order of magnitude in both experiments. Also, it may be explained by the structural properties of the soil which are constantly renewed by the cyclical agronomic practices.

Indeed, Vachaud et al. (1985) introduced the concept of "time stability' of soil moisture patterns as temporal invariance in the relationship between spatial location and statistical measures of soil moisture and water pressure head. Time stability can be also be viewed as the temporal persistence of spatial distribution patterns (Kachanoski and de Jong, 1988). For instance, in three different agro-pedo-climatic contexts of studying soil moisture, Vachaud et al. (1985) tested the time stability concept and showed that particular sites within a field always displayed mean behavior while others always represented extreme values. Munoz-Pardo et al. (1990) also analyzed the spatial variation of gravimetric water content at three dates of sampling, textural components and yield components of two rainfed crops cultivated on the same field. They found that the dryest and wettest locations at one sampling date trend to remain the dryest and wettest ones at the other dates. In that case, they explained this time stability by a determinism which was mainly imposed by the spatial distribution of silt plus clay content of the soil.

CONCLUSIONS

In the present study, spatial analysis of soil hydraulic properties was revisited after 17 consecutive years of repeated agronomic practices on the same field. Surface infiltration tests were performed using the Beerkan infiltration method and its algorithm *BEST* to characterize the soil through the van Genuchten and Brooks and Corey equations. The hydraulic properties and their spatial structures were compared to those performed in 1990 on the same field soil, through the exponential form of the hydraulic conductivity given by the Gardner equation, using the Guelph Pressure Infiltrometer technique.

Our field results suggested that both the K_s and $\alpha_{vG\text{-}2007}$ values were still better described by the lognormal distribution. In contrast to the Gardner equation, the van Genuchten/Brooks and Corey expression was found to better represent a possible gradual increase of water filling the porosity, which is more representative for a well-graded particle-size distribution of a loamy soil. The mean value of $\alpha_{vG\text{-}2007}$ was found to be identical to that of $\alpha_{G\text{-}1990}$, while that of $K_{s\text{-}2007}$ was significantly smaller than that of $K_{s\text{-}1990}$. The geostatistical analysis performed on 40 field estimates of both $K_{s\text{-}2007}$ and $\alpha_{vG\text{-}2007}$ and made at each node of a 6×7.5 m grid, showed that these two parameters were autocorrelated up to about 30 m and to 21 m, respectively, as well as spatially positively correlated together within a range of 30 m. Despite the difference in the mean values of K_s between the two studies, the spatial structures of both the K_s and alpha parameter were similar to those found in the 1990 field experiment except for the covariance sign. Knowing that the geostatistical analysis deals with the variance between observations, the similarity in autocorrelation ranges may indicate that the spatial analysis of soil hydraulic properties is independent of the infiltration methods (i.e., measurement of an infiltration flux) used in the two studies. However, the difference observed in the covariance sign can be referred to the use of two different methods of soil hydraulic parameterization in the two studies. The parameter estimation techniques used in the 1990 field study are based on subjective choice of the early-time and thus the complementary

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steady-state experimental data that influenced the estimate of the alpha parameter and consequently the calculation of the K_s value. The positive values of covariance found in the 2007 field study indicate that the value of $\alpha_{vG\text{-}2007}$ will be higher as much as the value of $K_{s\text{-}2007}$ will be higher. As α_G^{-1} or α_{vG}^{-1} is a measure of the importance of capillary forces relative to gravity for water movement, the positive spatial correlation between saturated hydraulic conductivity and the alpha parameter seems to be in agreement with the physics of water infiltration process in unsaturated soils. Therefore, robustness in the parameter estimation method to accurately estimate sorptivity and hydraulic conductivity from infiltration modeling data is needed to measure the spatial correlation between hydraulic properties of a field soil.

Although there are only two distant dates of measurements (i.e., 17-year interval), semivariograms showed that the spatial patterns of hydraulic parameter variations seem to be temporally stabilized, at least within the agro-pedo-climatic context of the study. This time stability may be referred to a possible existence of a deterministic factor imposed by the pore network, dry bulk density and particle-size distribution which remain constant in time. It may also be explained by the soil structural properties which are constantly renewed by the cyclic agricultural practices.

The number of data points used in the current study seems to be limited. Several hundred of observations are suggested in literature (Rao et al., 1979) to really decide on the appropriate distribution function of soil properties. Also, Oliver and Webster (1991) recommend at least 100 to 200 sampling locations for accurate estimation of a semivariogram. Such number was not available in most of reported studies. However, our number of measurements is in agreement with the findings of some other authors who recommend that the number of data values must be greater than 30 (Journel and Huijbregts, 1978). Therefore, in the future studies, increasing the number of measured data would be recommended if the measuring and sampling methods permit it.

The experimental findings of the present work raise the importance of taking into account the new values of hydraulic parameters of the loamy soil studied herein and their spatial variability to effectively apply many agronomic and environmental research and management. The Beerkan infiltration method and its algorithm BEST appear to be a promising, easy, robust, and inexpensive way of characterizing the hydraulic behavior of soil and its spatial and temporal variability. Similar systematic studies in the same field are suggested in future in order to test more completely the concept of long time stability. It is hoped that this work will contribute to an increased understanding of water transfer at an agricultural field for enhancing the sustainable use and management of soil and water resources.

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The effect of cultivation on the spatial variability of soil properties and their relationships to projected past and present microtopography will also be examined.

The effects of long term land management on observed spatial variance relationships has not been studied previously.

van Pelt and Wierenga (2001) came to a similar conclusion about soil matric potential.

The normality of data frequency distribution was tested using both the Kolmogorov–Smirnov test and the values of the skewness (g_1) and kurtosis (g_2) coefficients. The K-S test on original dataset of α_{vG} produced a p-value of 0.03 and when the dataset was log-transformed, the p-value changed to 0.19. As the log-transformed data passed the K–S normality test at a significance level of higher than 0.05, the null hypothesis (H0: the data are drawn from a normally distributed population) should be rejected. Moreover, based on student tests at the significance level of 0.05, the value of the skewness (g_1) indicated that the normal distribution (g_1=0) should be rejected (Table 2). K_s values cannot pass either the normal or the lognormal tests. The p-values produced for both original and log-transformed datasets of K_s, were <0.001. However, on the basis of student tests at the significance level of 0.05, the normal distribution (g_1=0) should be rejected whereas the lognormal one is more acceptable for K_s (Table 2).