COMPARING MINI-DISK INFILTROMETER, BEST METHOD AND SOIL CORE ESTIMATES OF HYDRAULIC CONDUCTIVITY OF A SANDY-LOAM SOIL

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ABSTRACT

Saturated, K_s, and near-saturated, K, soil hydraulic conductivity control many hydrological processes but they are difficult to measure. Comparing methods to determine K_s and K is a means to establish how and why these soil hydrodynamic properties vary with the applied method. A comparison was established between the K_s and K values of a sandy-loam soil obtained, in the field, with the BEST (Beerkan Estimation of Soil Transfer parameters) method of soil hydraulic characterization and an unconfined MDI (mini-disk infiltrometer) experiment and, in the laboratory, with a confined MDI experiment and the CHP (constant-head permeameter) method. Using for the BEST calculations the soil porosity instead of the saturated soil water content yielded 1.4 to 1.1 times higher estimates of K_s and K, depending on the pressure head, and differences decreased in more unsaturated soil conditions. The confined MDI experiment yielded 22% - 77% higher K values than the unconfined MDI experiment, depending on the established pressure head, h_0, and differences were not significant for h_0 = -1 cm. In the close to saturation region, the soil hydraulic conductivity function predicted with BEST did not generally agree well with the K_s and K values obtained in the laboratory by a direct application of the Darcy's law. In particular, BEST yielded a 5.6 times smaller K_s value than the CHP method and up to an 8.1 times higher K value than the MDI. Overall, i) the two application methods of the MDI yielded relatively similar results, especially close to saturation, and ii) there was not a satisfactory agreement between the field (BEST) and the laboratory (MDI plus CHP) determination of soil hydraulic conductivity close to saturation, unless a comparison was made with the same soil water content. The detected differences were probably attributable to soil spatial variability, overestimation of K_s in the laboratory due to preferential flow phenomena, underestimation of K_s in the field due to air entrapment in the soil and infiltration surface disturbance, inability of BEST to describe the actual soil hydraulic conductivity function at the sampled field site. Testing BEST predictions of K_s and K in other soils appears advisable and combining the MDI and CHP methods appears a rather simple means to make these checks. These additional investigations could improve interpretation of the differences between methods, which is an important step for properly selecting a method yielding K_s and K data appropriate for an intended use.

Keywords: Soil hydraulic conductivity; Field methods; Laboratory methods; Mini-disk infiltrometer; Constant-head permeameter; BEST methods of soil hydraulic characterization.

INTRODUCTION

Knowledge of the hydrodynamic soil properties is essential for understanding water flow and solute transport processes in the soil-plant system (Autovino et al., 2018; Basile et al., 2020; Farzamian et al., 2021). Hydraulic conductivity of saturated, K_s, and near-saturated, K, soil are particularly important since flow and solute transport processes occur at the highest possible rates at or close to saturation. Saturated and near-saturated hydraulic conductivity can be determined by many methods (Carter and Gregorich, 2007; Dane et al., 2018; Angulo-Jaramillo et al., 2016), differing for various aspects such as the application ambit (laboratory, field), the established flow field (one-, 1D, or three-, 3D, dimensional), the information used to determine K_s or K (transient, steady-state, both transient and steady-state).

The choice of a method over another can be expected to have appreciable effects on determination of K_s and K, which raises the problem of choosing an appropriate measurement method in relation to the specific purpose of the sampling campaign and considering the advantages and disadvantages of the possibly usable methods (Alagna et al., 2016; Paige and Hillel, 1993; Reynolds et al., 2000). For example, laboratory methods are relatively comfortable to apply and they can guarantee highly controlled conditions which positively affect the reliability of the individual measurement. However, soil compaction and alteration of pore connectivity represent two of the possible adverse consequences of using a presumably undisturbed sample in the laboratory (Bouma, 1982; Lauren et al., 1988; Lee et al., 1985). Soil disturbance can be controlled to some extent when measurements are performed in the field, which also implies maintaining the connection between the sampled soil volume and the surrounding soil. However, field experiments are generally more tiring and also less accurate than the laboratory ones since they have to be performed in environmental conditions that are not always fully favorable for carrying out measurement activities. In addition, many potential limitations of several field methods can be identified such as relatively small sample size, possible edge flow along the ring wall, disturbance of the exposed soil surface due to water application, difficulty to guarantee a correspondence between theory and practice (Bagarello and David, 2020; Reynolds et al., 2008; Xu et al., 2012). In this complex context, one of the few sources of information from which practitioners can choose the appropriate methods for their specific circumstances is provided by comparisons between alternative methods for measuring K_s or K (Ghosh et al., 2019;

Reynolds et al., 2000). These comparisons are important also because, especially on large areas, K_s or K data can be collected by applying different methods. According to Braud et al. (2017), in this particular case it is necessary to develop appropriate methodologies for creating an equivalent set of hydraulic conductivity data from data obtained with different methods. Therefore, the comparison between K_s and K determination methods continues to remain a central aspect of soil hydrology research.

Simple methods to determine soil hydraulic conductivity include BEST (Beerkan Estimation of Soil Transfer parameters) methods (Angulo-Jaramillo et al., 2019, 2016; Bagarello et al., 2014; Lassabatère et al., 2006; Yilmaz et al., 2010), the mini-disk infiltrometer (MDI; Dohnal et al., 2010; METER Group, 2021) and the constant-head permeameter (CHP) method (Reynolds and Elrick, 2002).

With BEST, a 3D field infiltration run under a nearly null ponded depth of water yields an estimate of K_s and the η parameter of the Brooks and Corey (1964) hydraulic conductivity function. Therefore, a given function is assumed to describe the relationship between K and the volumetric soil water content, θ, and the methodology produces an estimate of the parameters of this function.

The MDI is a miniaturized tension infiltrometer that allows simple and rapid determination of the soil hydraulic conductivity corresponding to fixed pressure head values, h, in the range from -0.5 cm to -6 cm. Typically, the device is used in the field to establish a 3D infiltration process at an established h value (Alagna et al., 2013; Fodor et al., 2011; Gonzalez-Sosa et al., 2010), but it was also applied to anthropic soils and green infrastructures substrates (Bondì et al., 2023; Gadi et al., 2017; Radinja et al., 2019). The corresponding K value is then obtained according to Zhang (1997) and Dohnal et al. (2010). However, the MDI has also been used in the laboratory on soil columns to establish 1D infiltration processes (Assouline and Narkis, 2011; Kargas et al., 2018). Using the unit hydraulic gradient (UHG) method (Klute and

Dirksen, 1986) with the MDI, a given sample can be equilibrated at fixed, and high (close to zero), *h* values to obtain points of the hydraulic conductivity curve close to saturation (Bagarello et al., 2007).

The CHP represents the standard method for determining K_s in the laboratory and it is often used as a benchmark for evaluating other methods (Reynolds et al., 2000). A 1D flow process is established under a constant hydraulic head on an initially saturated soil sample (Madsen et al., 2008). Knowledge of the hydraulic head gradient and measurement of volumetric flow rate yields the K_s value by direct application of the Darcy's law.

Considering BEST, the MDI and the CHP, alone or in some combination among them, as possible methods to obtain K_s or K data leads to recognize that there are at least three issues that require investigation and development.

According to the original BEST application procedure (Lassabatère et al., 2006), the saturated soil has to be sampled at the end of an infiltration run to determine the saturated gravimetric water content which is then transformed into a volumetric value, θ_s, by considering the dry soil bulk density. With another simpler and largely applied approach, θ_s is assumed to coincide with soil porosity, ϕ (Bagarello et al., 2011; Di Prima, 2015; Mubarak et al., 2009a; Xu et al., 2009; Yilmaz et al., 2010). The effect of the θ_s estimating method on the BEST prediction of soil hydraulic properties has been tested with somewhat contrasting results. For example, Alagna et al. (2016) suggested that sampling the soil confined by the ring at the end of the beerkan run to obtain an experimental value of θ_s could be expected to yield a more reliable estimation of soil hydraulic properties in comparison with that obtained by assuming a coincidence between θ_s and ϕ. Instead, the conclusion by Di Prima et al. (2017) was that the assumed coincidence between θ_s and ϕ as an alternative approach to the direct measurement of θ_s could be expected not to have a strong effect on estimation of K_s. Therefore, it does not seem clear to what extent ϕ represents a valid alternative to the direct measurement of θ_s.

For a fixed pressure head, *h*, the mean *K* value of the undisturbed soil in an area of interest can be obtained by field or laboratory MDI runs. The dependence of *K* on the MDI application method is unknown since the MDI has rarely been applied to establish both 3D and 1D infiltration processes. An exception is the investigation by Kargas et al. (2018), whose objective however was to determine the infiltration shape parameter, γ, of the infiltration model by Haverkamp et al. (1994). Consequently, the unconfined 3D and confined 1D MDI runs were carried out in the laboratory on repacked soil samples. The outcome of a comparison between field and laboratory determinations of *K* on undisturbed soil is difficult to predict *a-priori*. On the one hand, a certain similarity can be expected since the size of the sample, that can be expected to affect *K* determination (Reynolds et al., 2000), does not vary much between a field and a laboratory experiment. On the other hand, however, several other factors could induce a difference between field and laboratory determination of *K* with the MDI. For example, differences in the established flow field (3D in the field and 1D in the laboratory) could determine differences in *K* estimation since flow is not forced to follow a pre-established direction only in the first case (Reynolds and Elrick, 1985). Soil spatial variability (Logsdon and Jaynes, 1996) can affect a comparison between 3D and 1D MDI experiments. A reason is that, in the laboratory, all *K* values can be determined on a single sample regardless of the established *h* values whereas, in the field, each *K(h)* data point is obtained at a different sampling point. Further, differences can also be expected as a consequence of the applied method to analyze the data, that is the Darcy law in the laboratory (Klute and Dirksen, 1986) and an infiltration model in the field (Dohnal et al., 2010; Zhang, 1997).

BEST (Lassabatère et al., 2006) allows a field determination of the unsaturated hydraulic conductivity function according to the Brooks and Corey (1964) model. An independent estimation of near-saturated *K* values can be obtained by collecting an undisturbed soil sample in the field and using it in the laboratory for a sequence of 1D MDI runs followed by a CHP run. In the former case, the experiment is very simple but it is assumed that *K* decreases for smaller (more negative) pressure heads according to a pre-established law. In the latter case, the experiment is longer and it only yields some discrete *K* values, depending on the number of applied pressure heads. Comparing these two methods to obtain hydraulic conductivity is advisable. A reason is that BEST methods have been largely applied in many circumstances (Angulo-Jaramillo et al., 2019) but their predictions have been compared with soil hydraulic properties obtained with independent methods only in an overall limited number of investigations, not always with unequivocal results (Alagna et al., 2016; Bagarello and Iovino, 2012; Castellini et al., 2018). Supporting the usability of BEST methods could open new perspectives of practical interest. For example, BEST can potentially be used to also obtain an estimate of the macroscopic capillary length (Di Prima et al., 2020; White and Sully, 1987) and this soil parameter allows estimating bulb geometric variables for both buried and surface infiltration point sources (Baiamonte et al., 2024; Philip, 1984). Therefore, BEST experiments could be suggested for designing point irrigation systems taking into account spatial variability of soil hydraulic properties, with implications in terms of water saving and efficiency of water distribution.

The general objective of this investigation was to compare saturated and near-saturated hydraulic conductivity of a sandy-loam soil obtained with BEST, MDI and CHP methods. The specific objectives were to: i) compare soil hydraulic conductivity obtained with BEST when two different methods for estimating the saturated volumetric water content are used; ii) compare, for fixed pressure head values, the soil hydraulic conductivity obtained with field and laboratory application of the MDI; and iii) compare the soil hydraulic conductivity relationship obtained with BEST with that determined by using 1D MDI and CHP runs for the close to saturation region.

MATERIALS AND METHODS

Field site

The experimental site is located near Palermo (Western Sicily, Italy), in the area formerly known as Conca d'Oro (Golden Basin), where the soil is fertile and there is a good supply of freshwater (38°04'53.1" N and 13°25'08.4" E). In the field, there is a 30-year-old mandarin orchard planted with a spacing of 5 m × 5 m. The soil is not tilled and only mechanical weed control is performed. The altitude is 35 m a.s.l. and the surface is flat. The soil is a typic Rhodoxeralf with a depth of nearly 1 m and a moderate gravel content. According to the USDA classification, the soil texture, determined on two replicate soil samples, is sandy-loam with percentages of clay, cl = 16.6%, silt, si = 20.2% and sand, sa = 63.2%. The sampled area was of about 50 m^2.

At the beginning of the field campaign, disturbed soil samples were collected at three randomly chosen points within the sampling area and at two depths (0-5 and 5-10 cm) for each point. This soil was used to determine the gravimetric soil water content, w (kg/kg). The mean of these six w values was assumed to represent the antecedent gravimetric soil water content, w_i (kg/kg), at the field site since the sampled area was small and all field runs were carried out in a short time, that is in two consecutive days.

Experimental methods

Beerkan infiltration experiments

Small diameter (D = 0.08 m) rings were inserted on the soil surface to a depth of 0.01 m for the beerkan infiltration runs (Lassabatère et al., 2006). Rings were inserted manually or by gently using a rubber hammer and ensuring that the upper rim of the ring remained horizontal during insertion. The rings were relatively small since the soil surface layer was moderately stony and the small size of the ring simplified finding appropriate surfaces for soil sampling (**Fig. 1a**). A total of 15 infiltration runs were carried out in a single day at randomly selected locations. For each run, 20 water volumes, each of 57 mL, were successively poured in nearly 3 s for each volume on the confined infiltration surface. Therefore, with each volume, the initial ponded depth of water was equal to 11.3 mm. For each water volume (1^{st}, 2^{nd}, …, 20^{th}), the infiltration time was measured from water application to disappearance of all water, when the subsequent water volume was poured on the infiltration surface (Bagarello et al., 2021; Lassabatère et al., 2006). Energy of the applied water was dissipated on the fingers of the hand in an attempt to minimize soil disturbance due to water application. A cumulative infiltration, I (mm), vs. time, t (h), curve, comprising 20 data points, was therefore obtained at each sampling point.

Mini-disk field infiltration experiments

Mini-disk infiltrometers (MDI; METER Group, 2021), having a disk diameter of 4.5 cm, were used to perform three-dimensional (3D) infiltration experiments in the field. In particular, the MDI was used to obtain 3D infiltration data for established pressure heads, h_0, equal to -6, -3 and -1 cm. Each individual infiltration process for an established h_0 value was carried out at a different sampling point, so that all infiltration curves were obtained under similar initial soil water content conditions. For a given h_0 value, the experiment was replicated at 15 different, randomly chosen, sampling points, for a total of 45 MDI infiltration runs in the field. The soil surface at a sampling point was gently leveled with a trowel and small amounts of loose soil were used when necessary to improve the contact between the device and the infiltration surface (**Fig. 1b)**. The MDI was fixed to a support to keep it still during the run and infiltration was measured until the reservoir emptied. Readings were taken visually at ≤1 min, during the first stage of the run, to 5 min intervals in the most advanced stages of the infiltration process. A cumulative infiltration curve, $I(t)$, was obtained at each sampling point.

Mini-disk laboratory infiltration experiments

Mini-disk infiltrometers were also used to perform 1D infiltration experiments in the laboratory. At this aim, undisturbed soil cores were collected at 15 randomly chosen points of the soil surface, after scraping out the first cm of soil, in 5-cm-inner diameter by 10-cm-high stainless-steel cylinders to determine unsaturated soil hydraulic conductivity, K (mm/h), in the laboratory. Each cylinder was inserted vertically into the soil by hammering gently on the top of the cylinder with a rubber hammer and progressively removing the surrounding soil up to the established depth to reduce disturbance during sampling. A nylon guard cloth was attached to the base of the core to prevent soil loss from the bottom of the sample. To improve the contact between the MDI and the soil, its surface was gently leveled in the laboratory with a sharp knife and, when necessary, small amounts of loose soil were applied to the top of the sample.

To establish a given h_0 value at the base of a soil core, a plastic box of 38 (length) × 17 (width) × 13 (height) cm^3 was filled with a bed of sand and several small holes (diameter = 1 cm) were made on the walls of the box at a downward distance $h^* = h_0$ (L) from the surface of the sand bed (**Fig. 1c**). A metal net was glued to each hole to prevent sand from escaping. Water was added to the box to form a saturated zone below the holes and an unsaturated zone above them. At hydrostatic equilibrium, the soil water pressure head at the surface of the sand bed was assumed to be equal to h_0. Different boxes were prepared, depending on the considered h_0 value (h_0 = -6, -3, -1 cm). The first established pressure head was h_0 = -6 cm.

The soil cores were placed on the surface of the sand box and left to equilibrate for 24-48 hours. During this period, small volumes of water were periodically added to the box to maintain a constant h_0 value at the surface of the sand bed. Then, the MDI, set at this h_0 value, was placed on the soil surface by gently pressing the device and using a support to maintain it in place (**Fig. 1c**) and infiltration was measured. The small space between the walls of the cylinder and the porous plate of the device (5 mm) was not sealed to enable air to freely escape from the soil core during the run. After the run, the soil sample was left to freely drain for nearly 12 hours. Then, the same core was equilibrated for 48 hours at -3 cm and the MDI run at h_0 = -3 cm was carried out. Finally, after another 12 hours of free drainage, the core was equilibrated for 48 hours at -1 cm and the last MDI run at h_0 = -1 cm was performed. Generally, each individual run continued until the reservoir of the device emptied. For particularly slow runs, infiltration was stopped after 6 hours. Readings were taken visually at 1 to 5 min time intervals, with shorter intervals in the early stages of the run and for the higher (closer to zero) pressure heads. A run with a given h_0 value was replicated on 14 soil cores since a soil core broke during laboratory treatment. Immediately after each run with a given h_0 value, the soil columns were weighed to later determine the final gravimetric, *w* (kg/kg), soil water content.

Laboratory constant-head permeameter experiments

After the 1D MDI experiment at h_0 = -1 cm, the soil cores were left exposed to the air for a few days and then they were used for measuring the saturated soil hydraulic conductivity, K_s (mm/h), with the constant-head laboratory permeameter (CHP) method. Preliminarily, these cores were saturated from the bottom according to Booltink and Bouma (2002). The saturation procedure lasted a total of 24 hours, during which, specifically, the soil cores were placed inside a plastic box and the water level was raised five times in 2 cm increments. A funnel was used to support the sample and to collect percolating water (**Fig. 1d**). A Mariotte bottle was used to establish the constant head of about 1 cm above the soil surface. The amount of water passing through the sample was measured by weighing the collected water volume at fixed intervals of 1 min. A Scout Pro portable electronic balance 4000 g connected to a CR1000 datalogger was used to automate the measurements. The average value of the flux density was calculated for a fixed time interval. The experiment was considered concluded when the flow approached a steady-state condition characterized by a near constant steady-state value, i.e. when the flux density appear nearly stable in 10 consecutive time intervals. At the end of the run, the soil was dried in an oven at 105° C for 48 h and the dry soil bulk density, ρ_b (g/cm^3), was determined from the oven-dry soil mass and the bulk soil volume.

Calculations

An estimate of the antecedent volumetric soil water content, θ_i (m^3/m^3), that is, the soil water content at the beginning of the field campaign of measurements and sampling, was obtained by the product between w_i and the mean of the 14 ρ_b values obtained after the CHP experiment.

The volumetric soil water content, θ (m^3/m^3), at the end of a 1D run with the MDI set at a given pressure head (h_0 = -6, -3 or -1 cm) was calculated from the gravimetric soil water content and the dry soil bulk density of the considered soil core.

The BEST-steady algorithm (Bagarello et al., 2014) was applied to estimate the soil water retention curve (van Genuchten, 1980) and the hydraulic conductivity function (Brooks and Corey, 1964) from the experimentally determined intercept, b_s (mm), and slope, i_s (mm/h), of the straight line fitted to the last three (I, t) data points describing steady-state conditions on the cumulative infiltration plot. The shape parameters of the retention and hydraulic conductivity curves were obtained from the particle-size distribution data (Lassabatère et al., 2006). BEST-steady requires antecedent, θ_i (m^3/m^3), and saturated, θ_s (m^3/m^3), volumetric soil water content that, in this investigation, were assumed not to change from point to point, since the sampled area was small and the beerkan runs were carried out in a single day. The value of θ_i was obtained from the w_i determination. To determine θ_s, it was considered that field-saturated soil water content is generally lower than porosity, ϕ (m^3/m^3), due to the presence of entrapped air (Reynolds and Elrick, 2002; Reynolds and Topp, 2008). According to several investigations, θ_s/ϕ can vary from 0.70 to 0.95 (Alagna et al., 2016; Dane and Hopmans, 2002; Gonzalez-Sosa et al., 2010; Mubarak et al., 2009a; Somaratne and Smettem, 1993; Verbist et al., 2013). Therefore, in this investigation, θ_s was estimated from the fitted line to the $\theta(h)$ values that were determined after the 1D MDI runs. The results of BEST application obtained with this estimate of θ_s were indicated as BEST_R (R indicating the use of water retention data) calculations. However, in practical application of BEST methods, θ_s is often set equal to the porosity, that is easily determined from the ρ_b measurements (Auteri et al., 2020; Bagarello et al., 2023; Mubarak et al., 2009b). Therefore, this approach was also applied for comparative purposes and these results were indicated as BEST_P (P indicating porosity).

For each MDI infiltration run in the field, the two-parameter infiltration model (Philip, 1957) was fitted to the cumulative infiltration, I (mm), vs. time, t (h), data by minimizing the sum of the squared residuals between the measured and the predicted I values (Lassabatère et al., 2006) to simultaneously estimate the C_1 ($mm/h^{1/2}$) and C_2 (mm/h) parameters of the model. The soil hydraulic conductivity value corresponding to a given pressure head was then calculated according to Zhang (1997) and Dohnal et al. (2010). The required soil water retention parameters for calculating K were taken from Castellini et al. (2018), who applied the evaporation method to characterize this field site in a previous investigation.

The 1D MDI measurements were made applying the same pressure head at the two ends of the soil core. Therefore, the assumption was that, at steady-state, a unit hydraulic gradient was established, i.e. the pressure head was the same throughout the soil core. Consequently, steady-state flux density was equivalent to the unsaturated soil hydraulic conductivity corresponding to the imposed pressure head, i.e. K_{-6}, K_{-3} or K_{-1} (mm/h), depending on the run. Finally, the Darcy's law was applied to calculate K_s from the CHP laboratory data.

Data analysis

Three sets of K_s values (BEST_P, BEST_R, CHP method) and, for each negative pressure head, four sets of K values (BEST_P, BEST_R, 1D MDI, 3D MDI) were overall obtained. For each of these 15 datasets, the hypothesis that the data were normally distributed was never rejected according to the Lilliefors (1967) test at $P = 0.05$. Consequently, the arithmetic mean and the associated coefficient of variation, CV (%), were used to summarize each dataset.

Three different comparisons were carried out.

A comparison was established between the soil hydraulic conductivity values obtained with BEST_R and BEST_P. At this aim, a two-tailed paired t test at $P = 0.05$ was applied for K_s, K_{-1}, K_{-3} and K_{-6}. This comparison was made since i) field-saturated soil water content can be expected to be smaller than soil porosity (e.g., Reynolds and Elrick, 2002), ii) the BEST protocol assumes that θ_s is directly measured (Lassabatère et al., 2006), but iii) in many practical applications of BEST, θ_s is assumed to coincide with soil porosity (Bagarello et al., 2011; Mubarak et al., 2009b; Xu et al., 2009; Yilmaz et al., 2010).

The K values obtained with the 1D and 3D MDI experiments were compared with each other. An F test and a two-tailed t test at $P = 0.05$ were applied for each h_0 value and hence for K_{-1}, K_{-3} and K_{-6}. Moreover, the cumulative empirical frequency distributions (CFDs) of the laboratory and field data were compared to succinctly visualize all the data for a given pressure head. This comparison was made because, to our knowledge, little is known about what to expect when exactly the same device, that is the MDI, is used to determine the unsaturated hydraulic conductivity of an undisturbed soil with methods which differ by the established infiltration process (3D, 1D) and the method of data analysis for determining *K* (fitting an infiltration model to the data, directly using the Darcy's law).

Finally, the *K* values obtained with BEST-steady were compared with those obtained by the CHP method (K_s) and the 1D MDI experiments (K_{-1}, K_{-3} and K_{-6}). Also in this case, F and two-tailed t tests at $P = 0.05$ were applied for each considered variable. The objective of this comparison was to evaluate differences between a direct measurement of hydraulic conductivity values at and near saturation using undisturbed soil samples and the Darcy's law with a field estimate of these values. This comparison was made since there is much practical interest for BEST methods (Angulo-Jaramillo et al., 2019) but the predicted soil hydraulic properties with this methodology have been compared with those obtained with other methods only in a few investigations.

RESULTS

Dry soil bulk density and soil water content

The mean dry soil bulk density, ρ_b, was equal to 1.329 g/cm^3 (coefficient of variation, CV = 4.0%; sample size, N = 14) and the mean antecedent gravimetric soil water content, w_i, was equal to 0.094 kg/kg (CV = 11.0%; N = 6). Consequently, the mean antecedent volumetric soil water content, θ_i, was of 0.125 m^3/m^3.

The relationship between θ and *h* was approximately linear in the $-6 \leq h \leq -1$ cm range with a coefficient of determination, R^2, of 0.69 (**Fig. 2**). The θ vs. *h* data were described a little better ($R^2 = 0.73$) by a power relationship but this relationship was not considered since it was not usable to estimate the θ value corresponding to a null pressure head. According to the fitted linear regression line, $\theta(0) = \theta_s$ was equal to 0.382 m^3/m^3. Therefore, θ_s = 0.382 m^3/m^3 was considered for the BEST_R calculations. The soil porosity, ϕ, obtained from the mean ρ_b value and assuming a soil particle density of 2.65 g/cm^3, was equal to 0.499 m^3/m^3. Therefore, θ_s = 0.499 m^3/m^3 was considered for the BEST_P calculations. The θ_s/ϕ ratio, equal to 0.77, was rather small but it appear plausible since it fell in the range of the θ_s/ϕ values found in the literature (Alagna et al., 2016; Dane and Hopmans, 2002; Gonzalez-Sosa et al., 2010; Mubarak et al., 2009a; Somaratne and Smettem, 1993; Verbist et al., 2013).

The θ_i/θ_s and θ_i/ϕ ratios were equal to 0.33 and 0.25, respectively. The θ_i/θ_s ratio was a little higher than the threshold of 0.25 recommended by Lassabatère et al. (2006) for applying BEST. However, wetter conditions can occur in practice (Xu et al., 2012) and Di Prima et al. (2016) showed that, with BEST-steady, the estimated soil hydrodynamic parameters can be expected to be accurate even in initially relatively wet conditions for many soils. Therefore, a θ_i/θ_s ratio of 0.33 was considered not to impede application of BEST in this investigation.

Infiltration and soil hydrodynamic parameters

Generally, the established beerkan infiltration processes appeared consistent with theory since the concavity of the *I* vs. *t* curves was faced downwards, denoting that the infiltration rates initially decreased during the run, and the *I* vs. *t* relationship assumed a nearly linear shape at longer times. Consequently, soil hydraulic conductivity data were obtained at each sampling point with both BEST_R and BEST_P (**Table 1**).

Also in the case of the field MDI experiments, infiltration appeared consistent with theory since infiltration rates initially decreased during the run and they stabilized at longer times. The method by Zhang (1997) and Dohnal et al. (2010) was successfully applied to calculate K_{-1}, K_{-3} and K_{-6} at each sampling point (**Table 1**).

For the majority of the 1D laboratory runs with the MDI, the *I* vs. *t* relationship was nearly linear from the early stages of the infiltration process or it appeared to some extent concave upwards (**Fig. 3**). Therefore, infiltrations rates, *ir* (L/T), were nearly stable during the run or they stabilized after a phase of increasing values. For each run, the stabilized *ir* value obtained by linear regression of the last *I* vs. *t* data points, ir_s, was compared with the slope of the linear regression line fitted to all *I* vs. *t* data and forced to pass through the origin of the axes, *sl*. A similarity between ir_s and *sl* was expected when the *I* vs. *t* relationship was nearly linear for the entire infiltration process whereas $ir_s > sl$ denoted an upwards concavity of this relationship (Bondì et al., 2023). On average, ir_s and *sl* differed by 1.12, 1.22 and 1.27 times for h_0 = -1, -3 and -6 cm respectively. Moreover, a linear regression analysis between *sl* and ir_s was carried out (**Fig. 4**). According to the calculated 95% confidence intervals for the intercept and the slope, the linear regression line between these two variables did not differ from the identity line by jointly considering all pressure heads (-7.42 – 0.42 and 0.98 – 1.07, respectively) and also for h_0 = -1 cm (-27.5 – 13.2 and 0.91 – 1.19). However, it differed from the identity line for both h_0 = -3 cm (-1.10 – 2.70 and 0.69 – 0.87) and h_0 = -6 cm (0.06 – 1.53 and 0.57 – 0.77). Therefore, a smaller h_0 value produced a greater upwards concavity. However, taking into account that the estimate of the steady-state infiltration rate coincides with the estimate of *K* for the 1D MDI experiment and that an error of 25% can be considered negligible even in more stringent conditions, that is when the data are free of any experimental error (Reynolds, 2013), the detected concavity did not introduce any relevant uncertainty on estimation of *K*. Therefore, all infiltration runs were included in the developed *K* dataset for each established pressure head and, for each run, *K* was assumed to coincide with ir_s (**Table 1**).

All CHP experiments were successful and they yielded a K_s value for each sampling point (**Table 1**).

Comparisons

BEST-R vs. BEST-P

Regardless of h, higher K values were obtained with BEST_P as compared with BEST_R (**Table 1**) but differences decreased monotonically as the pressure head became smaller (more negative). In particular, two corresponding estimates of K differed by 1.37, 1.36, 1.30 and 1.08 times for h = 0, -1, -3 and -6 cm, respectively. Relative variability of the K values was nearly independent of the saturated soil water content estimation approach.

In the following, the BEST_R calculations were considered for the comparison with other estimates of K for the following reasons: i) air is usually entrapped in a porous medium when it is saturated by downward infiltrating water under ponded conditions (e.g., Reynolds and Elrick, 2002), such as in the case of the beerkan infiltration runs, ii) use of θ_s is conceptually more consistent with the original BEST method (Lassabatère et al., 2006), and iii) using BEST with a saturated soil water content smaller than ϕ can be expected to represent the best choice to satisfactorily reproduce laboratory measured soil water retention values (Alagna et al., 2016).

Field vs. laboratory MDI experiments

The 1D MDI experiment yielded higher K values than the 3D one by 22% for h_0 = -1 cm, 35% for h_0 = -3 cm and 77% for h_0 = -6 cm (**Table 1**). Differences between two corresponding datasets were statistically significant for the two lowest pressure heads but not for the highest h_0 value. In the field, the highest relative variability of the data was detected for the smallest pressure head. Instead, in the laboratory, relative variability was highest for the highest pressure head. Consequently, the 1D MDI K values were 1.8 times more variable than the 3D MDI ones for h_0 = -1 cm (high, for the 1D data, vs. medium, for the 3D data, variation) (Warrick, 1998)Fare clic o toccare qui per immettere il testo. and 1.8 times less variable for h_0 = -6 cm (medium vs. high variation). Relative variability of the two estimates of *K* was similar, and medium, for h_0 = -3 cm (*CV* = 32-37%). For both h_0 = -6 and -3 cm, the cumulative empirical frequency distribution (CFD) of the field *K* values was entirely located to the left of the one corresponding to the laboratory values (**Fig. 5**). For h_0 = -1 cm, the two CFDs intersected with each other since the laboratory method yielded both the highest and the lowest of the overall determined *K* values.

BEST_R vs. CHP and 1D MDI experiments

Regardless of the considered variable (K_s, K_{-1}, K_{-3}, K_{-6}), the differences between the field (BEST_R) and the laboratory (CHP and 1D MDI) estimates of *K* were statistically significant (**Table 1**). The CHP method yielded a 5.6 times higher K_s value than BEST_R and the lowest laboratory value of K_s was 1.4 times larger than the highest K_s value obtained in the field (**Fig. 5**). The 1D MDI estimates of K_{-1} were 1.7 times larger than those obtained with BEST_R. For K_{-3} and K_{-6}, BEST_R yielded higher estimates than the 1D MDI by 3.4 and 8.1 times, respectively. In both cases, *K* increased monotonically as the pressure head became less negative (**Table 1**), but at a very different rate for the two tested methods (**Fig. 6**). In particular, the K_s/K_{-6} ratio was equal to 1.23 with BEST_R and 55.4 with the laboratory soil cores. Therefore, BEST_R predicted a nearly flat soil hydraulic conductivity curve for $h \geq -6$ cm. This shape was not consistent with the *K* data obtained in the laboratory that instead suggested that even a small variation of *h* induced large changes of *K*. The BEST_R $K(h)$ curves intersected the data obtained in the laboratory (**Fig. 6**). In particular, the former curves were positioned below the laboratory *K* data for *h* = 0 and above them for $h \leq -3$ cm. A certain overlap between the curves obtained with BEST_R and the *K* values measured in the laboratory was detected for h = -1 cm. Relative variability of K_s did not differ appreciably between BEST_R and the CHP method and it was medium in both cases (**Table 1**). In unsaturated conditions, BEST_R predicted nearly constant *CV* values that instead decreased, although not exactly monotonically, from high to low pressure heads with the 1D MDI experiments. According to Warrick (1998), relative variability of the unsaturated K values was generally medium, with two exceptions (BEST_R, h = -6 cm; 1D MDI, h = -1 cm), in which variation was high. Even in this case, however, the *CV* values were not appreciably greater than the value that discriminates between a medium and a high variation (*CV* = 50%).

DISCUSSION

Estimating saturated soil water content for the BEST calculations

The choice of the θ_s value to be used for the BEST calculations is expected to be important to recognize a correspondence between estimated and independently measured water retention values (Alagna et al., 2016) but it seems that the same cannot be said with reference to soil hydraulic conductivity. The reason is that the estimates of K_s with BEST_P and BEST_R differed by 1.4 times (**Table 1**) which can be considered a rather small difference (Elrick and Reynolds, 1992). Moreover, for another sandy-loam soil, Di Prima et al. (2017) reported that changing the estimate of θ_s (porosity instead of the field measured value) implied that the K_s estimates differed by only 1.2 times. Probably, the effect was smaller than that detected in this investigation since our θ_s/ϕ value was equal to 0.77 whereas it was 0.85 (i.e., closer to one) in the study by Di Prima et al. (2017).

This investigation also demonstrated that the effect of the used approach for estimating θ_s decreased for more negative pressure heads. To explain this last result, that apparently was not discussed so far, it has to be noted that, with the applied BEST-steady algorithm, K_s and K are given by (Bagarello et al., 2014):

$$K_s = \frac{C\, i_s}{\frac{\gamma\, b_s}{r(\theta_s - \theta_i)} + C} \qquad (1a)$$

$$K(\theta) = K_s \left(\frac{\theta}{\theta_s}\right)^{\eta} \qquad (1b)$$

where i_s (L/T) and b_s (L) are the slope and the intercept, respectively, of the linear regression line fitted to last data points that describe steady-state conditions on the I vs. t plot, C is a constant that, for an initial soil hydraulic conductivity, $K_i << K_s$, does not depend on the soil water content, θ, and is equal to 0.639, γ is a parameter for geometrical correction of the infiltration front shape, usually assumed to be equal to 0.75, r (L) is the radius of the source and η is a shape parameter. An increase of θ_s (e.g., using ϕ instead of $\theta_s < \phi$) implies a smaller denominator in eq.(1a) and hence a higher estimate of K_s (Di Prima et al., 2017). According to eq.(1b), the increase of K_s is partially compensated by a decrease of θ/θ_s, which explains why the estimate of θ_s had a smaller impact on the K calculations for more negative pressure head values.

In any case, when $\theta_s < \phi$, the BEST_P calculations appear conceptually less reliable than the BEST_R ones since, in the former case, infiltration parameters (i_s, b_s) obtained in a soil that contains air are treated as if they had been obtained in a completely saturated porous medium.

Upwards concavity of cumulative infiltration for the laboratory mini-disk runs

An attempt to explain the unexpected upwards concavity of the $I(t)$ curves obtained in the laboratory with the 1D MDI runs (**Fig. 3**) was made and at least three possible reasons were identified.

A possible reason was that the contact between the device and the infiltration surface improved during the run, notwithstanding that this surface appeared leveled and smoothed before firmly putting the MDI in place. In other terms, the actual infiltration surface was smaller than expected at the beginning of the run and then it became larger. Non-uniform wetting even under controlled laboratory conditions on repacked soil columns was reported by Close et al. (1998). These authors also concluded that contact problems can be particularly noticeable in the case of low established pressure heads. There was agreement between this suggestion and our results since a smaller h_0 value produced a greater upwards concavity.

An upwards concave cumulative infiltration curve can be obtained when the soil is initially water repellent but water repellency decreases during infiltration (Alagna et al., 2019). Water repellency should not have a strong effect on infiltration in initially wet soil conditions (de Jonge et al., 1999; Dekker et al., 2001) that were those of this experiment given that, at the beginning of a run, the base of a soil core was equilibrated at not less than -6 cm of water, However, this interpretation was not excluded because other authors also signaled water repellency phenomena in initially relatively wet soil conditions (Carrick et al., 2011).

According to Faybishenko (1995), air can initially be entrapped in the smallest soil pores but, under the influence of capillary forces, water is drawn into these pores so that the entrapped air is displaced into the largest pores. Therefore, another possible interpretation was that some air was entrapped in the smallest soil pores at the beginning of the infiltration run and then it escaped. Infiltration rates increased during the run with the smallest pressure head because the hydraulically active pores were small and they initially contained some entrapped air. This entrapped air did not affect appreciably infiltration rates at the highest pressure head since, in this case, the hydraulically active pores were large enough not to be blocked by air escaping from the smallest pores.

Of course, a less uncertain interpretation of the detected concavity would likely require additional experimental investigations that should particularly be carried out for small pressure heads. However, it does not seem that these investigations constitute a priority in a practical perspective since the detected concavities were not so appreciable as generating great doubts on the reliability of the *K* estimates.

An upward concavity was not observed in the field since, in this case, infiltration rates tended to stabilize after a transient decreasing stage. Therefore, it appeared plausible to believe that: i) the contact between the device and the soil was better in the field, probably because the leveling of the contact surface was easier in the absence of any confinement of the infiltration surface; ii) in the field, possible soil water repellency effects were masked by lateral capillarity forces that were not active in the laboratory; and/or iii) entrapped air effects were less noticeable in the field since air could escape more easily due to the lack of any confinement of the sampled soil volume. Moreover, infiltration occurred under larger pressure head gradients in the field than in the laboratory. Perhaps, even this difference contributed to determine a different shape of the experimental *I* vs. *t* relationship.

Comparing mini-disk experiments

With reference to saturated hydraulic conductivity of undisturbed soils, differences from a reference value by nearly 60% (Yilmaz et al., 2023) or by a factor of two or three (Elrick and Reynolds, 1992) could be considered negligible, at least for some practical purposes. Assuming that these suggestions also apply to the case of the near-saturated soil hydraulic conductivity, it could be suggested that the 1D and 3D MDI experiments overall yielded similar results, particularly close to saturation (**Table 1**). However, another interpretation was that the 1D experiment tended to generally yield higher *K* values than the 3D experiment, more noticeably for the smallest pressure heads.

Obtaining a larger *K* value with a 1D experiment than a 3D one could be due to a significant contribution to total flow of macropores extending from the surface to the bottom of the soil sample (Bagarello et al., 2007) or to an overestimation of the steady-state 1D flow rate

(Bagarello et al., 2010). However, explaining the differences between the 1D and 3D data as a consequence of preferential flow phenomena did not appear convincing since the largest discrepancies were detected with the lowest established pressure head, which did not activate the largest voids in the sample.

Instead, it can be supposed that 1D K > 3D K was obtained since ir_s (slope of the stabilized part of the I vs. t relationship) was used instead of sl (slope of the linear regression line fitted to all the I vs. t data points) for calculating 1D K and $ir_s > sl$ was obtained for the smallest pressure heads (**Fig. 4**). Although this interpretation could find a numerical support, there were physical reasons for using ir_s instead of sl. The assumption that K coincides with the steady-state infiltration rate can only be made if the flow process is stable. The last part of each infiltration run appeared clearly linear in all cases (**Fig. 3**) denoting that, starting from a certain point, the process stabilized. Instead, considering a not perfectly stable process as if it was stable, i.e. using sl for the K calculations, would have made the reliability of the estimated slopes more questionable in the perspective to determine K.

It was also deemed unlikely that differences were attributable to soil compaction during sampling since, in this case, the opposite result (3D K > 1D K) would have been expected.

An effect of the methods used to determine K from the MDI infiltration data was not completely excluded since a dependence of the K values on the applied calculation method is documented in the literature (Dohnal et al., 2010; Jacques et al., 2002; Logsdon and Jaynes, 1993). Nor can it be ruled out that differences occurred as a consequence of spatial variability of soil hydrodynamic properties at the field site, given that different points were sampled with the 3D and 1D MDI runs. Moreover, it could also be considered that the 1D data were expressive of a vertical infiltration process whereas the 3D data were representative of a combined vertical and lateral flow process. Therefore, forcing water to move vertically perhaps reduced the overall tortuosity of the flow paths, especially with reference to the smallest active pores given that the differences between the 1D and 3D results increased as h_0 decreased.

Field vs. laboratory determination of soil hydraulic conductivity

According to the BEST_R, CHP and 1D MDI data (**Table 1** and **Fig. 6**), the laboratory prediction of K_s was appreciably larger than the field one, that is by 5.6 times. For unsaturated soil conditions, the K estimates were nearly independent of h with BEST_R but they decreased appreciably with smaller pressure heads according to the MDI data. Consequently, the two approaches yielded relatively similar results for h = -1 cm (difference by 1.7 times) but BEST_R predicted appreciably higher K values than the MDI in more unsaturated conditions (by 3.4 and 8.1 times for h = -3 and -6 cm, respectively).

A possible reason of the differences between the laboratory and the field determination of K_s was that the laboratory experiment actually yielded excessively high K_s values. It would not be the first time that soil cores yield higher (even by orders of magnitude) K_s values than those obtained by establishing in the field an infiltration process on an initially unsaturated soil (e.g., Bagarello and Provenzano, 1996; Jačka et al., 2014; Paige and Hillel, 1993; Reynolds et al., 2000). High K_s values could depend on rapid pipe flow through worm holes, old root channels and cracks that extended through the core (Reynolds et al., 2000) or on cracks and fissures created during the collection procedure (Jačka et al., 2014). According to Paige and Hillel (1993), discontinuous macropores in the field can become continuous in a particular soil sample. It cannot be said without any doubt that pipe flow phenomena occurred in this investigation since the length of the soil cores (10 cm) was appreciably greater than that considered in other investigations (e.g. 3 cm in Paige and Hillel, 1993) and continuity of large pores between the upper and the bottom ends of the soil sample appears less probable with a longer sample. On the other hand, it could also be suspected that these cavities may have formed during the experiment given that K_s was measured after three previous K determinations with the MDI, that is, after an intense use of the sample albeit with all possible precautions.

On the other hand, it cannot ever be ruled out that the K_s values obtained with the field experiment were too low. In this investigation, there were some signs that this circumstance occurred even if these signs were not unequivocal. In particular, the 1D MDI experiments (but also the 3D MDI ones) yielded $K_{-1} > K_s$ (**Table 1**), which is physically impossible. This result did not necessarily represent a proof that the field K_s values were too low since it could also be a consequence of spatial variability of soil hydrodynamic properties (e.g., Logsdon and Jaynes, 1996; Prieksat et al., 1994). In other words, these differences occurred because the points sampled with the beerkan runs were inherently less permeable than those sampled for the MDI runs. However, a rather large number of runs were carried out with each method on an overall small area, according to existing guidelines (Reynolds et al., 2002), and this circumstance induced to be cautious in proposing an interpretation exclusively based on spatial variability considerations. Instead, the conclusion that $K_{-1} > K_s$ signaled that BEST_R yielded too low K_s values appeared more convincing than, or at least as plausible as, the suggestion that particularly low permeability points were sampled with the beerkan runs but not with the MDI experiments.

One of the possible reasons why field K_s values were too low was that some air was entrapped in the soil during the ponding infiltration runs (Lee et al., 1985; Mohanty et al., 1994). Indeed, there were many opportunities to induce some air entrapment in the sampled soil volume since the beerkan run was performed in accordance with Lassabatère et al. (2006) and hence by adding a new water volume when the previous water volume had completely infiltrated and the infiltration surface was entirely exposed to air (Bagarello et al., 2021). Field-saturated soil hydraulic conductivity values can be expected to be two or even more times smaller than completely saturated soil values (Gupta et al., 1993; Jačka et al., 2014; Reynolds and Elrick, 1987). Moreover, the results of this investigation were consistent with the conclusion by Sakaguchi et al. (2005) that the saturated hydraulic conductivity measured on a soil containing entrapped air can be smaller than the unsaturated hydraulic conductivity close to saturation.

Another possible reason why the field K_s values were too small was that repeatedly pouring water on an initially unsaturated soil altered the upper soil layer that became progressively less conductive. Mechanical impact of the applied water volumes was minimized by applying water close to the infiltration surface and dissipating the energy of the water on the fingers of the hand. However, some slaking perhaps occurred in the early stages of the run since water was suddenly applied on an initially unsaturated soil (Le Bissonnais, 1996). Consequently, pore sizes decreased and flow paths became perhaps more tortuous. Slaking did not occur in the laboratory experiments since, in this case, the soil was initially wetted slowly from the bottom.

Finally, another possible explanation for low K_s values was that soil structure collapsed to some extent during ring insertion. However, this explanation appeared rather unrealistic since the ring was inserted to a very short depth in the soil (1 cm) and with great caution.

With reference to the unsaturated soil hydraulic conductivity, a similar result to that of this investigation was obtained by Alagna et al. (2016) in a comparison, for a loam soil, between BEST and the classical tension infiltrometer (TI). In particular, differences between corresponding K values were relatively small (by 1.2-3.0 times, depending on the BEST algorithm) for h = -1 cm but BEST yielded 9 to 35 times higher K values than the TI for $-12 \leq h \leq -3$ cm. According to these authors, this difference occurred because the assumed hydraulic conductivity function in BEST did not reproduce satisfactorily the changes in the soil pore system for $h < -1$ cm. In other words, representation of the soil as a single permeability system was responsible of the poor matching between the two tested methods.

Even in this investigation, differences between the predicted and the measured K values for a given pressure head appeared to depend on the inability of BEST to describe the soil hydraulic properties close to saturation. In particular, the experimental θ values decreased appreciably as h decreased from -1 to -6 cm (from 0.371 to 0.284 m^3/m^3) but the BEST_R predictions of θ varied only minimally in this range of h values (from 0.382 to 0.377 m^3/m^3; **Fig. 2**). The experimental and modelled θ values were rather similar for $h = -1$ cm and an approximate similarity was also detected with reference to the corresponding estimates of K (**Table 1** and **Fig. 6**). Moreover, plotting K against θ (**Fig. 7**) showed that i) regardless of the imposed h_0 value, the 1D MDI experiment yielded increasing K values with θ, as expected; and ii) the K values predicted with BEST_R fell in a zone of the figure which indicated a certain correspondence between the laboratory and the field K data for similar θ values. In particular, with BEST_R, the means of θ and K were equal to 0.380 m^3/m^3 and 67.5 mm/h, respectively. According to the fitted regression line to the K(1D MDI) vs. θ data, $K_{\theta=0.38}$ was equal to 94.0 mm/h, that is larger than the BEST_R prediction by 39.3%. This difference was rather small (Elrick and Reynolds, 1992; Yilmaz et al., 2023) and it suggested that the BEST_R predictions of K and the 1D MDI values were of the same order of magnitude when the soil water content was the same. In other words, BEST_R overestimated K at small pressure heads since it was unable to predict the same decrease in θ that was detected experimentally. This result reinforces the need to define the setup of BEST-2K, that extends the existing BEST methods for use in dual-permeability soils (Lassabatère et al., 2014, 2019).

In this investigation, standard approaches were applied to determine K with the different devices and experimental methodologies, meaning that different representations of the $K(h)$ relationship were considered. For example, the Brooks and Corey (1964) model was assumed for BEST whereas no assumptions were made with the 1D MDI since individual data points (i.e., a K value for a pre-established h value) were obtained by direct application of the Darcy law. Moreover, a specific BEST algorithm was used to analyze the infiltration data collected by the beerkan runs but other algorithms are available in the literature (Lassabatère et al., 2006; Yilmaz et al., 2010). Future developments of this investigation should also explore these methodological issues taking into account that soil hydraulic conductivity may be expected to vary with the adopted model (Lenhard et al., 1989; Mubarak et al., 2009a; Valiantzas, 2011) and different BEST algorithms could yield different estimates of the soil hydrodynamic properties, depending on the specific circumstances (Angulo-Jaramillo et al., 2016, 2019).

CONCLUSIONS

This investigation contributed to expand, for a sandy-loam soil, our knowledge on different simple field and laboratory methods usable to determine the saturated and near-saturated soil hydraulic conductivity.

Using, for the BEST (Beerkan Estimation of Soil Transfer parameters) calculations, the soil porosity, ϕ, instead of the true saturated soil water content, θ_s, yields higher estimates of soil hydraulic conductivity but differences decrease in more unsaturated soil conditions. It is recommended to make use of θ_s since the established beerkan infiltration process will likely give rise to air entrapment in the soil. In practice, however, using ϕ instead of θ_s could be expected not to introduce large uncertainties in soil hydraulic conductivity estimation.

Pooling the data from the unconfined MDI (mini-disk infiltrometer) infiltration measurements performed in the field and the confined MDI measurements performed in the laboratory appears a possible way to proceed in the perspective to obtain a mean K value for close to saturation soil with reference to an area of interest. The validity of this conclusion becomes weaker in more unsaturated conditions but even in this case it can be expected that only a moderate noise will be introduced in the estimate of K.

In the close to saturation region, the soil hydraulic conductivity function predicted with BEST does not generally reproduce direct measurements of K obtained in the laboratory at different pressure heads by a direct application of the Darcy's law with the MDI and the CHP (constant-head permeameter) method. In particular, the expectation could be that BEST will yield a too small K_s value and too high K values for the more unsaturated conditions. However, a satisfactory correspondence between BEST and laboratory determination of K can be expected for the same soil water content.

Supporting these conclusions with other comparisons is of course necessary. Based on the experience gained in this study, improving the organization of the experiment could be advisable even if delineating these improvements is not easy. For example, spatial variability of soil hydrodynamic properties was suggested to perhaps influencing to some degree some of the established comparisons. In principle, spatial variability effects could be prevented or appreciably reduced by applying all measurement methods at a single sampling point according for example to the sequence i) 3D MDI, ii) beerkan run, iii) collection of the undisturbed soil core, iv) 1D MDI and, finally, v) CHP application. In this case, all measurements refer to nearly the same soil volume. However, it cannot be said that this is the best choice, since the experiment will become unavoidably longer, for example because the soil needs to dry out after the 3D MDI run and before performing the beerkan run. Moreover, repeated solicitations on exactly the same soil volume could promote soil structure alterations having an impact that would be difficult to detect. Therefore, a possible alternative could be to reduce uncertainties attributable to soil spatial variability by performing more runs for each applied measurement method. However, interpreting differences between methods does not solve any problem since the most appropriate measurement method for an intended use of the data has to be chosen. A way to reach this objective could be comparing a monitored soil hydrological process with the corresponding one predicted by a mechanistic model and the measured soil hydrodynamic properties.

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AUTHOR CONTRIBUTION

Vincenzo Alagna: Investigation, Writing - Reviewing and Editing. **Dario Autovino:** Investigation, Data Curation, Writing- Original draft preparation. **Vincenzo Bagarello:** Conceptualization, Writing- Original draft preparation. Methodology, Formal analysis, Data Curation. **Gaetano Caltabellotta:** Investigation **Mariachiara Fusco:** Investigation. **Massimo Iovino:** Methodology, Writing - Reviewing and Editing. **Girolamo Vaccaro:** Investigation**.**

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