Estimating the saturated soil hydraulic conductivity in a farm constructed wetland by the borehole permeameter infiltration method

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***Abstract*— The borehole permeameter infiltration method was used to determine the soil hydrodynamic properties at different depths in a farm constructed wetland (CW) in which the saturated soil hydraulic conductivity, K_s, was already investigated for the uppermost surface layer. With the aim to estimate K_s and the sorptive number α*, a non-linear curve fitting approach was used to fit both the Philip (1993) and the Reynolds (2011) models to the experimental infiltration data. The estimated parameters were compared with those obtained with the graphical approach by Philip (1993) and the empirical approach by Regalado et al., (2008). Both K_s and α* varied along the soil profile but also between the inlet and outlet zone of the CW. The graphical approach was not applicable whereas the empirical approach overestimated both K_s and α* compared with the non-linear fitting approach. Despite the differences among the considered approaches, the borehole permeameter technique confirmed that the subsurface K_s was at least two orders of magnitude lower than that obtained by the ring infiltrometer at the CW surface thus confirming the importance of conducting these experiments for the study of the groundwater recharge processes.**

Keywords*— *Philip-Dunne permeameter, groundwater recharge, saturated soil hydraulic conductivity, infiltration experiment, wetland

I. INTRODUCTION

Constructed wetlands (CWs) are extensively applied for preventing nonpoint source pollution from agricultural activities [1], [2]. Apart from treatment of agricultural runoff and drainage water, CWs can also serve to control flood peak and retain stormwater [3] and to recharge groundwater when the CWs is not waterproofed.

The groundwater recharge volume is mainly controlled by the infiltration and drainage processes that act into the quasi-saturated soil profile from the surface down to the groundwater table. Measurement of saturated soil hydraulic conductivity, K_s, is therefore crucial for modelling downward water movement from the CWs, also with the aim to assess the overall wetland water balance, as well as to control excessive leakage that may be harmful in case of dissolved NO_3-N contaminants [4].

Surface saturated soil hydraulic conductivity, K_s can be measured with a relatively limited experimental effort by ring infiltrometer techniques (e.g., [5]). In their investigation [6] applied the BEST-steady approach [7] to estimate the hydrodynamic parameters (soil sorptivity and saturated hydraulic conductivity) at three locations of a meandered, 470 m long, pervious CW designed to treat the agricultural drainage water. Mean estimated K_s ranged from 30.5 mm/h at the inlet to 293 mm/h at the outlet as a consequence of sealing due to selective settling of suspended soil particles. However, surface point measurements of K_s were at least two order of magnitude higher than the groundwater recharge rate estimated from a global water balance, thus indicating that this process is mainly governed by the hydraulic conductivity of the deeper soil layers [6].

Measuring the subsurface hydrodynamic properties is not easy given the difficulty of carrying out an experiment below the soil surface thus making the results uncertain. Several methods have been proposed to measure K_s along the soil profile that basically make use of infiltration data collected into cased or uncased well permeameters under constant or falling head of water.

The cased or borehole permeameter, also known as the Philip-Dunne permeameter [8] consists of a circular tube, with an inner radius equal to r, that extends to the base of a vertical borehole. The bottom of the borehole represents the infiltration surface in which a falling-head infiltration process is imposed. Water is suddenly introduced into the tube to a depth of D_0 at time $t = 0$ and then the process continues up to $t = T$, that is the time when the borehole empties. During the infiltration process, the water depth on the infiltration surface, D, is measured repeatedly to obtain the experimental $D(t)$ drawdown curve.

Reference [8] proposed a graphical method to calculate K_s and the sorptive number, α*, from measurement of infiltration times at two pre-established water levels, i.e., $D = D_0/2$ and $D = 0$. However, making use of only two (D, t) data pairs alone implies sensitivity of K_s and α* determinations to

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random measurement errors, short-term flow variations caused by soil heterogeneities, and systematic "lack of fit" between the data and the selected model [9]. To avoid this problem, numerical fitting of the theoretical curve to the whole $D(t)$ curve was proposed [9], [10]. Reference [11], starting from the Philip's approach, proposed two empirical relationships that simplify K_s and α^* determination. Reference [9] developed an extended analysis for falling-head lined borehole permeameter introducing a variety of discharge geometries. For the case of vertical discharge only, the extended analysis differs from the model proposed by Philip only for the choice of parameters accounting for flow efficiency and gravity effects.

With the aim to estimate the hydrodynamic properties of the soil profile, falling-head infiltration was measured in lined boreholes dug at different depths at the inlet and outlet of the farm CW already studied by [6]. A non-linear curve fitting approach was applied to fit both the Philip and the Reynolds models[8], [9] to the experimental $D(t)$ data to simultaneously estimate K_s and α^*. The nomograph approach by [8] and the empirical approach by [11] were also applied for comparative purposes.

II. Theoretical Background

The extended analysis developed by [9] allows to obtain an estimate of both the saturated soil hydraulic conductivity, K_s (L T^{-1}), and the sorptive number α^* (L^{-1}), on the basis of the approximate analytical solution for three-dimensional Green-Ampt (GA) [12] infiltration model. The soil is assumed to be homogeneous and isotropic with an initially uniformly unsaturated water content. The GA pressure head at the wetting front, ψ_f (L), is assigned as $\psi_f = -\alpha^{*-1}$. For the case of vertical flow, i.e., water infiltrating only from the bottom of the borehole, the simplifying hypothesis that the infiltration surface has a spherical shape with a radius r_0 (L) equal to $r/2$ can be made. By applying the derivation procedure of [8], the following relationship was proposed between the scaled time, τ, and the scaled bulb radius, ρ:

$$\tau = \left(1+\frac{1}{2A}\right)\ln\left(\frac{A^3-1}{A^3-\rho^3}\right) - \frac{3}{2A}\ln\left(\frac{A-1}{A-\rho}\right) + \frac{\sqrt{3}}{A}\left[\tan^{-1}\left(\frac{A+2\rho}{\sqrt{3}A}\right) - \tan^{-1}\left(\frac{A+2}{\sqrt{3}A}\right)\right] \quad (1)$$

where:

$$A^3 = \frac{3\left(\alpha^{*-1}+D_0+CG\right)}{r_0\Delta\theta}+1 \quad (2)$$

$$\tau = \frac{K_s}{Cr_0}t \quad (3)$$

$$\rho^3 = \frac{3\left(D_0-D(t)\right)}{r_0\Delta\theta} \quad (4)$$

in which $\Delta\theta$ is the difference between the saturated, θ_s (L^{3}L^{-3}), and the initial, θ_0 (L^{3}L^{-3}), soil water content, C is a flow efficiency correction coefficient and G is a coefficient related to gravity that varies between 0 and r_0 [8]. Reference [9] used $C = 1$ instead of $C = \pi^2/8$ originally assumed by [8][8] because radial and combined radial-vertical discharge through the borehole screen should be about as efficient as flow through the equivalent sphere surface. As for G constant, [8][8] assumed $G = r_0$ that corresponds to maximum gravity flow, whereas [9] numerically proved that the choice of a G had virtually no effect on K_s estimates and little effect on α^* estimates for strongly capillary soils. Hence, $G = 0$ was suggested as the best choice.

Equation (4) allows to retrieve the maximum scaled radius of the wetted bulb at the end of the experiment, ρ_{max} (i.e. for $D = 0$ and $t = T$), and the ρ corresponding to $D = D_0/2$, ρ_{05}:

$$\rho_{max} = \left(\frac{3D_0}{\Delta\theta r_0}+1\right)^{1/3} \quad (5a)$$

$$\rho_{05} = \left(\frac{\rho_{max}^3}{2}+1\right)^{1/3} \quad (5b)$$

The graphical procedure proposed by [8] first involves the determination of the relationship between α^{*-1} and τ_{max}/τ_{05}, where τ_{max} and τ_{05} are the τ values obtained by (1) and (5) with ρ_{max} and ρ_{05}, respectively. At this aim, a sequence of α^{*-1} values with a certain step is established. For a given α^{*-1}, A is computed by (2) with the appropriate values of ρ (ρ_{max}, ρ_{05}) and then τ_{max} and τ_{05} obtained by (1). Considering τ_{max}/τ_{05} must be equal to T/t_{05}, the developed α^{*-1} vs. τ_{max}/τ_{05} relationship allows to identify the value of α^{*-1} corresponding to the experimental information T/t_{05}. To estimate K_s, the τ_{max} values are plotted to τ_{max}/τ_{05} ratio with the aim to individuate the τ_{max} value corresponding to $\tau_{max}/\tau_{05} = T/t_{05}$. Finally, this τ_{max} value is used in the following relationship:

$$K_s = \frac{\pi^2 r_0 \tau_{max}}{8T} \quad (6)$$

In their analysis of Philip's model, [11] showed that reliable estimates of the suction parameter can be obtained only for $T/t_{05}<5$. They proposed the following statistical relationships for the estimation of τ_{max} and α^{*-1} (within the range from 1 to 100 m^{-1}):

$$\tau_{max} = 0.731\frac{T}{t_{05}} - 1.112 \quad (7)$$

$$ln\left(\alpha^{*-1}\right) = -13.503 + 19.678\left(\frac{T}{t_{05}}\right)^{-1/2} \quad (8)$$

The most interesting practical-applicative aspect of the proposed procedure is represented by the lack of need to determine $\Delta\theta$ for the estimation of K_s and α^{*-1}.

III. Material and Method

The study area is located at the experimental agricultural farm (12.5 ha) of Canale Emiliano Romagnolo land reclamation consortium (CER) in the Metropolitan city of Bologna. The volume drained by the whole farm area is collected to a CW that was constructed in 2000 with a surface of around 0.4 ha and an overall volume close to 1,470 m^3 (Fig. 1). Due to the pervious nature of the CW surface both infiltration and evapotranspiration occur during functioning.

In 2020 and 2021, two sampling campaigns were conducted at the inlet and outlet zone of the CWs in an area of 3x3 m^2. At each sampling site, two replicate boreholes excavated by a hand auger at the depths z = 0.5, 1.0 and 1.5 m (Fig. 1). PVC tubes with an external diameter of 6.3 cm were inserted in the boreholes. To avoid an incomplete contact between the external surface of the tube and the soil,

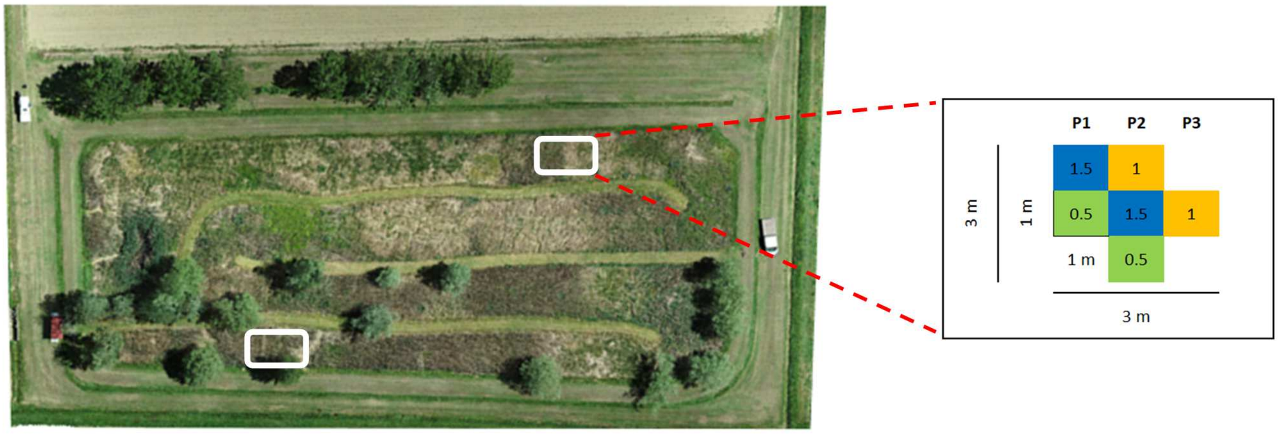

Fig. 1. Areal view of the wetland and the experimental design.

which could determine upflow of water in the hollow space during the run, the auger diameter was a few millimetres smaller that the external diameter of the tube. To facilitate the insertion of the pipes some grease was applied on the outer side of the tube. The base of the permeameter hole was carefully cleaned using a flat base screw ground anchor. Once the borehole permeameter was prepared, a Diver water level data logger sensor (Eijkelkamp, The Netherlands) was inserted inside each borehole to register the depth of water during the falling-head infiltration process. A fixed volume of water (670 mL corresponding to a water depth of 30 cm) was then poured into the pipe at time zero, and the time required for infiltration was recorded. The pipes were sealed in the upper part, above the ground, with plastic film to prevent water loss due to evaporation but also to prevent any rain from bringing water inside them.

At each sampling site (CW inlet and outlet), four undisturbed soil cores (5 cm in height by 5 cm in diameter) were collected at the three depths (0.5, 1 and 1.5 m), in separated boreholes, by means of a multiple sample auger kit (Eijkelkamp, The Netherlands) for determination of soil bulk density, ρ_b (M L^{-3}) and volumetric water content, θ_0 (L^3 L^{-3}), at the time of sampling. The saturated soil water content, θ_s, was assumed to be equal with soil porosity, ϕ, which was estimated by ρ_b, and soil particle density.

IV. Results

Table I shows the soil physical characteristics detected from undisturbed cores. Small differences were observed among the mean ρ_b and θ_0 values independently of the sampling site and explored layer. It can be concluded that bulk density and water content were relatively uniform, both vertically and horizontally. It is worth noting that $\Delta\theta$ varied between 0.02 and 0.11 cm^3 cm^{-3}, thus indicating that the soil was relatively close to saturated conditions. Despite the contribution of capillarity can be considered of minor importance under such wet initial conditions, according to [9], (1-4) remain valid and accurate for almost the entire $\Delta\theta$ range including near saturated conditions.

The infiltration $D(t)$ curves at inlet and outlet of the wetland are shown in Fig. 2. At the inlet, complete empting of the permeameter took from 90 to approximately 1200 h for the sampling depths up to 1 m. Specifically, the two experiments conducted at z = 0.5 m showed similar trend and duration (respectively 656 and 826 h) whereas the experiments conducted at z = 1.0 m were highly variable (Fig. 2). In the two experiments conducted at z = 1.5 m, after around 2 months of monitoring, less than 10 cm of water had infiltrated.

At the outlet zone of the wetland, most of the $D(t)$ curve showed a downward concavity. Permeameters emptied only in three out of six experiments, whereas in the other cases the water depth D had not reached half of the initial height even after 48 days of infiltration. The very long time the water took to infiltrate is probably due to the combined effects of very low hydraulic conductivity and small capillarity as a consequence of initial high soil water content but also to the massive soil structure as highlighted by the high ρ_b values (Table I).

Considering that Philip's graphical approach requires the complete emptying of the permeameter in order to obtain the times at two pre-established water levels, corresponding to $D = D_0/2$ and $D = 0$, the procedure was potentially applicable only in five out of the total 12 runs. However, even for these cases, the graphical approach did not allow obtaining K_s and α^{*-1} values with the only exception of a single infiltration test conducted for z = 0.5 m at the CW inlet (data not showed). Failure of the Philip's graphical approach was attributed to t_{05} values that were more than half the T value corresponding to the end of the experiment, thus indicating that the infiltration rate did not slow down as requested by theory (Fig. 2).

TABLE I. Physical Characterisation of the Wetland Soil for the Inlet and Outlet Points (N = 4)

Soil depth (cm)	Statistic	ρ_b [a]	θ_0 [b]	θ_s [b]	ρ_b [a]	θ_0 [b]	θ_s [b]
		Inlet			***Outlet***		
50-60	Min	1.54	0.34	0.40	1.49	0.39	0.43
	Max	1.59	0.39	0.42	1.51	0.42	0.45
	Mean	1.56	0.36	0.41	1.47	0.40	0.44
	CV%	1.40	6.00	2.00	1.89	3.08	2.37
100-110	Min	1.42	0.30	0.43	1.41	0.40	0.44
	Max	1.50	0.38	0.46	1.48	0.45	0.47
	Mean	1.47	0.34	0.45	1.44	0.43	0.46
	CV%	2.43	10.28	3.03	2.27	6.12	2.72
150-160	Min	1.43	0.42	0.45	1.36	0.44	0.48
	Max	1.45	0.45	0.46	1.37	0.48	0.49
	Mean	1.44	0.43	0.46	1.37	0.46	0.48
	CV%	0.62	2.81	0.74	0.80	5.92	0.85

a. ρ_b unit is g cm^{-3}.

b. θ_0 and θ_s unit is cm^3 cm^{-3}.

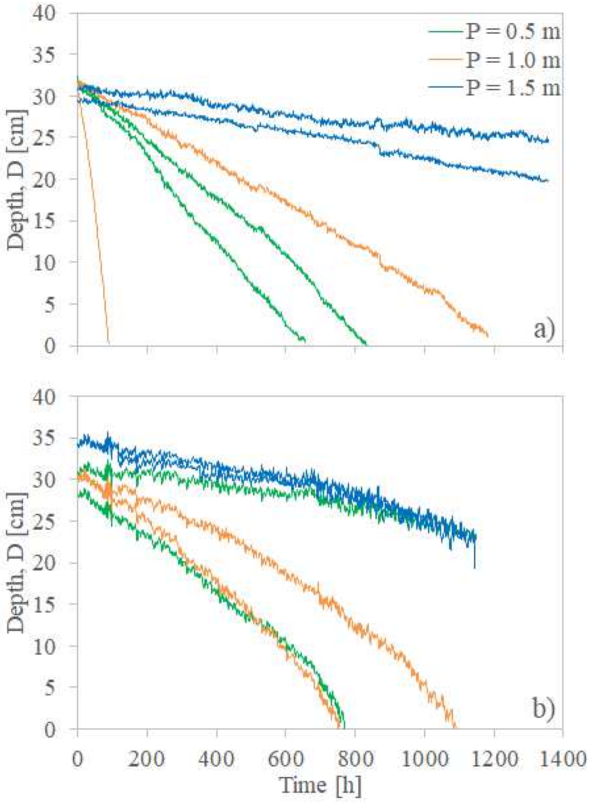

Fig. 2. Experimental D(t) drawdown curves at a) the inlet and b) outlet zone of the wetland.

Table II shows the mean estimated K_s and α^{*-1} values obtained by non-linear fitting of both the Philip and the Reynolds models to the experimental $D(t)$ data. The results of the empirical approach by [11] are also showed.

At the CW inlet, both approaches, i.e., [8] and [9], indicated that K_s at z = 0.5 m was up to one order of magnitude lower than that obtained at z = 1.0 and 1.5 m. Specifically, the differences were more relevant for the Philip model that estimated a mean K_s value into the upper layer (z = 0.5 m), respectively, 16 and 2.5 times lower than at the two deeper layers. This result can be in part explained by the higher mean ρ_b value observed in the upper layer (Table II). The α^{*-1} estimated by the two approaches decreased along the soil profile but, at a given depth, the [8] model yielded α^{*-1} that were from 5 to 30 times higher than the corresponding estimates obtained by [9]. Where applicable, the empirical procedure proposed by [11] gave K_s values that increased from 0.0068 at z = 0.5 m to 0.0152 mm h^{-1} at z = 1.0 m (i.e., a factor of 2.3) and α^{*-1} that were less variable along the profile. A trend to overestimate K_s in comparison with Philip and Reynolds models was observed, whereas estimated α^{*-1} values were either close to those estimated by [9] or [8] approach.

At the CW outlet, the mean K_s estimated by non-linear fitting of the [8] and [9] models to the experimental $D(t)$ data were generally lower than the corresponding values estimated at the CW inlet and less variable along the soil profile (Table II). In this case too, the differences found can probably be attributed to the ρ_b values that decreased along the soil profile (Table II). It was confirmed that maximum K_s value were observed into the intermediate layer (z = 1.0 m) and that the upper layer was the less conductive in the explored profile. Also, the mean α^{*-1} values decreased with depth and the tendency of the Philip model to yield higher α^{*-1} values than the Reynolds one was confirmed. The empirical procedure proposed by [11] was inapplicable for the experiments conducted at z = 1.5 m and for one experiment at z = 0.5 m. However, the mean K_s value at z = 1.0 m was in agreement with those obtained by the fitting approaches (i.e., [8] and [9] models). The α^{*-1} estimated by the empirical approach [11] increased from z = 0.5 to 1.0 m differently to α^{*-1} values estimated by the other two approaches that decreased with depth.

Apart from the Philip's graphical approach, which was not applicable, the two fitting approaches [8], [9] and the empirical approach [11] consistently signalled that the K_s at the CW outlet was lower than at the CW inlet and that, for both sampling sites, K_s was minimum in the upper layer and increased with depth. There were inconsistencies between the K_s estimates obtained with the different approaches that raised up to one order of magnitude, with the empirical approach [11] that yielded the highest estimates of K_s and the Reynolds one [9] the lowest ones. The estimates of α^{*-1} were high variable among the three approaches. A tendency of α^{*-1} to decrease with depth was observed that could be attributed to a reduced effect of capillarity as the initial soil moisture condition approach saturation.

The differences observed in terms of K_s, between the inlet and outlet of the wetland cannot be explained by the different soil bulk density since ρ_b values at the inlet zone are higher than at the outlet. Thus, it was hypothesized that the different hydraulic conductivity is due to the wetter initial moisture conditions at the outlet than slowed water infiltration.

It is worth noting that the K_s values for the subsurface layer decreased from the inlet to the outlet and this behaviour is opposite to that found by [6] for the surface layer. In addition to clogging phenomena, also [6] observed that higher θ_0 values yielded lower K_s values. Furthermore, the very low K_s values obtained at the point scale by the Philip-

TABLE II. MEAN SATURATED SOIL HYDRAULIC CONDUCTIVITY, K_s, AND PRESSURE HEAD AT THE WETTING FRONT, $-\alpha^{*-1}$, AT THE INLET AND OUTLET ZONE (N = 6) OF THE FARM WETLAND OBTAINED WITH DIFFERENT ANALYTICAL PROCEDURES

Zone	Models	K_s (mm h^{-1})			α^{*-1} (m)		
		P_0.5	*P_1*	*P_1.5*	*P_0.5*	*P_1*	*P_1.5*
Inlet	Philip [8]	0.0004	0.0066	0.0010	33.265	3.085	0.468
	Reynolds [9]	0.0007	0.0040	0.0004	1.074	0.589	0.066
	Regalado et al., [11]	0.0068	0.0152	-	1.488	3.009	-
Outlet	Philip [8]	0.0001	0.0010	0.0010	36.725	3.260	0.640
	Reynolds [9]	0.0001	0.0010	0.0006	22.014	0.454	0.003
	Regalado et al., [11]	0.0020	0.0010	-	5.366	7.970	-

Dunne permeameter were in agreement with the global infiltration rates estimated by the application of a water balance model to the wetland [6]

One critical aspect that deserves further investigation is related to the shape of the $D(t)$ curve that in most cases was linear, or even concave downward, thus apparently violating the theory that prescribes higher infiltration rates in the initial stage of infiltration as a consequence of the higher hydrostatic head into the borehole. Probably such finding is a consequence of the lower importance of D_0 term in (2) compared to the capillary term.

V. CONCLUSIONS

The borehole permeameter infiltration method was applied to estimate the hydrodynamic properties at the inlet and outlet zones of a farm constructed wetland for different soil depths down to 1.5 m. The saturated soil hydraulic conductivity, K_s, and the reciprocal of the sorptive number, α^*, were estimated by a non-linear curve fitting approach and the analytical solutions proposed by [8] and extended by [9]. The graphical [8] and empirical [11] approaches were also applied.

The model [8] tended to overestimate α^{*-1} as compared with model [9], whereas K_s estimates differed by a factor that varied from 0.57 to 2.50. The graphical approach was not applicable and the empirical approach, when applicable, overestimated K_s compared to the fitting procedures.

Despite the differences in K_s and α^{*-1} estimated by the considered approaches, a consistent trend was observed signalling that mean K_s values were higher at the CW inlet than at the CW outlet and minimum in the upper layer (z =0.5 m). The low conductivity observed close to the CW surface was probably a consequence of the settling of suspended soil particles that was already observed in a previous study [6]. More in general, the data obtained by the borehole permeameter technique confirmed that the subsurface hydraulic conductivity was several orders of magnitude lower than that obtained at surface soil layer indicating that sampling of the surface K_s alone could not be appropriate for interpreting the groundwater recharge processes that take place from a CW.

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