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On the use of soil hydraulic conductivity functions in the field

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Abstract

The estimation of soil water fluxes using the Darcy–Buckingham flux-gradient approach is, after a century of use, still problematic under field conditions. Two features of the soil hydraulic conductivity (K) function are the main causes of failure, first the exponential character of the K versus soil water content (θ) relations, which lead to large variations in K for minimal variations of θ. Due to this, the level of precision of field measurements of θ and the spatial variability of θ itself, make the deterministic estimation of soil water fluxes unfeasible using this approach. Secondly, the spatial variability of the parameters of the $K(\theta)$ relations also contribute heavily to errors in soil water flux estimation from site to site. In a coffee crop water balance experiment, soil water fluxes below the root zone were estimated over one year, comparing the use of a soil hydraulic conductivity function obtained in the field, with an indirect climatologic approach in which the deep drainage is estimated from a water balance excess. Five replicates gave the possibility of calculating variances of both forms of calculation and their respective coefficients of variation (CV). Results show that CVs of the estimates made through the Darcy–Buckingham approach varied from 78 to 122%, in comparison to 8–23 for the indirect climatologic approach. It is therefore concluded that Darcy–Buckingham approach used deterministically under field conditions does not yield consistent results.

Keywords: Darcy–Buckingham equation; Soil hydraulic conductivity; Soil water flux; Soil variability

1. Introduction

According to Addiscott (1993), water was the earliest subject of experiments carried out on soils, citing the French meteorologist de la Hire as the first to install lysimeters in 1688. The first quantitative theory on the behavior of water in soils appeared much later, in 1907, published by Buckingham, based on the well-known law of Darcy (1856). This theory, now known as the Darcy–Buckingham law, was used over the whole XX Century, and continues valid today. It is based on the flux-gradient approach, which includes soil hydraulic conductivity functions, specific for each porous medium. These soil hydraulic conductivity functions $K(\theta)$ or $K(h)$, were and are extensively used in laboratory and field experiments to estimate soil water fluxes, reported in a voluminous literature, a good sample of which can mostly be found in Kutilek and Nielsen (1994). However, the exponential character of the relation between K and θ or h, which reflects the physical nature of the flow of water in porous materials such as soils, makes the application of this theory very difficult. For laboratory soil columns packed with

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homogeneously sieved soil material or surrogate soils, and for field situations in which these functions are obtained using the very same soil body in which soil water fluxes are estimated, results have been successful (Larue et al., 1968; Davidson et al., 1969; Vachaud et al., 1983; Minasny et al., 2004; Lazarovitch et al., 2005). This exponential character also influences the determination of the hydraulic conductivity functions, making results dependent of the chosen method. Jones and Wagenet (1984) used five simple methods to calculate the hydraulic conductivity functions at one field spot, on 100 sites with a wide range in vertical and horizontal particle size distributions. Large variability of K_0 and β in the adopted relation $K(\theta) = K_0 \exp[\beta(\theta - \theta_0)]$ lead them to conclude that the five studied methods can be most useful in developing relatively rapid estimates over large areas, however, not as useful when to be used at a particular location at which soil water fluxes are to be precisely known. In other situations, when K functions are obtained at one site and applied in other or when they are averaged, however, many reports indicate difficulties, e.g. Reichardt et al. (1998) and Villagra et al. (1995), mainly due to the spatial and temporal variability of the soil matrix along the landscape.

Many studies on the characterization of soil porous systems (Kutilek and Nielsen, 1994) point to the need of applying the concepts of double conductivity or of bi-modality of soils. The $K(\theta)$ relation being directly related to soil pore distributions, has to follow their nature. Many field soils present bi-modal pore size distributions, which lead to two equivalent radii and, consequently, two water transport processes (Bouma, 1984; Othmer et al., 1991). According to these authors, interpedal pores together with large macropores conduct an accelerated water flux, and the interpedal pores (between aggregates) have a much lower hydraulic conductivity as compared to the whole soil. They claim that when primary, secondary and tertiary peds are present in a field soil, the use of a single $K(\theta)$ relation is not correct.

Another recent view is the stochastic approach (Vauclin and Vachaud, 1990) which involves the use of $K(\theta)$ probability functions, having as a result not a single deterministic value of water flow, but a range of water flows, with their respective probabilities.

In view of these difficulties of estimating soil water fluxes in field soils, this study makes a comparison between two ways of estimating the deep drainage component of water balances: (i) indirectly from the water balance equation; and (ii) directly, using the Darcy–Buckingham flux-gradient equation.

2. Materials and methods

2.1. Experimental field

A water balance experiment was carried out at Fazenda Areão, ESALQ/USP in Piracicaba, SP, Brazil, (22°42′S, 47°38′W, 580 m above sea level) on a soil classified as Rhodic Kandiudalf, locally called ‘‘Nitossolo Vermelho Eutroférrico’’, with a moderate A horizon of clayey texture. The climate is Cwa, according to Köppen's classification, with annual average temperatures, rainfall, and relative humidity of 21.1 °C, 1257 mm, and 74%, respectively.

Coffee plants (*Coffea arabica* L.), cultivar ‘‘Catuaí Vermelho’’ (IAC-144), were planted in rows along topographic contour-lines in May 2001, with plant spacing between rows of 1.75 and 0.75 m between plants of one row. Water balances were established for 14 days intervals, starting on 1 September 2003 (days after beginning, DAB 0) and continued until 30 August 2004 (DAB 364), using five replicates:

$$P + I + \text{ER} + \Delta S + \text{RO} + Q_L = 0 \quad (1)$$

where P is the rainfall, I the irrigation, ER the actual evapotranspiration, ΔS the soil water storage changes in the soil 0–1 m layer, RO the runoff, and Q_L the deep drainage at the lower boundary of the soil volume at the depth $z = 1$ m; all expressed in mm for 14-day periods.

P and I were measured using classical rain gauges installed in the center of each replicate. ΔS was estimated from neutron-probe measurements of θ made at depths of 0.2, 0.4, 0.6, 0.8, and 1.0 m, using the trapezoidal rule. To measure RO, plots of 12 m^2 were framed and the surface water flow was collected downhill, by gravity, in 200 L tanks. The actual crop evapotranspiration was estimated as a remainder in Eq. (1), using two distinct approaches:

(i) As an alternative to the Darcy–Buckingham approach, we suggest that in wet periods with deep drainage (Q_Li) is likely to occur, it is considered equal to zero in Eq. (1). In doing so, ERi is an overestimation of ER because it includes Q_Li. Thus, in periods in which this overestimated ERi was higher than the potential evapotranspiration (ETc), we considered ERi = ETc and the difference ERi − ETc = Q_Li. This way of calculating the actual evapotranspiration and the deep drainage is actually very similar to what is made in climatological water balances (Pereira et al., 2002) in which they call (ERi − ETc) as excess, not making a distinction between RO and Q_L. The potential evapotranspiration was estimated from the reference evapotranspiration (ET_0), corrected by the crop coefficient (K_C). ET_0 was calculated using the Penman–Monteith equation (Pereira et al., 1997) with meteorological data collected by a station near to the experimental area. In periods during which plants were not under stress, when the soil water storage was relatively high and without drainage, K_C was calculated dividing ERi by ETc. For all other periods, K_C was taken as an the average of the calculated values.

(ii) ERii calculated as an unknown in Eq. (1), estimating the drainage ($Q_L$ii) using Darcy's equation at the lower boundary of the soil volume element, $z = 1.0$ m. To estimate $Q_L$ii, tensiometers were installed in each plot at the depths of 0.9 and 1.1 m, with three replicates, to estimate the total soil water potential gradient (∇H, mm^{-1}) at $z = 1.0$ m, by means of finite difference (Eq. (3), below). The total soil water potential (H, m) was considered as the sum of the gravitational potential (z, m), negative downwards from the soil surface, and the matric soil water potential (h, m). The tensiometers, with mechanic vacuummeters (Bourdon capsules), read directly H values in mmHg. These values were transformed in meters of water column by the factor 1.36. The soil hydraulic conductivity is usually expressed as $K(\theta)$ or $K(h)$ functions but, although H depends on the depth z, which is fixed in our case for one depth only ($z = 1$ m), we chose the form $K(H)$ to facilitate calculations. The instantaneous water fluxes (q_L, mm day^{-1}) at $z = 1.0$ m were calculated as follows:

$$q_L = -K(H)\nabla H \tag{2}$$

where q_L is the Darcy–Buckingham flux density (mm day^{-1}) and ∇H is calculated by finite difference:

$$\nabla H = \frac{\Delta H}{\Delta z} = \frac{H_{1.1} - H_{0.9}}{0.2} \tag{3}$$

The numerical integration of the Eq. (2) in relation to the time, resulted in $Q_L$ii values:

$$Q_L\text{ii} = \int_0^{14} q\,dt = \sum_{i=1}^{14} q_i\,\Delta t \tag{4}$$

using $\Delta t = 1$ day, 14 q_i values were obtained for each water balance period. Tensiometer readings were made only during wet periods, when q_L was likely to occur, and not always daily. The response of tensiometers installed deeply, e.g. 1.0 m, is slow since the advance of a deep drainage front is also slow after having traveled for 1 m. Interpolations were needed in order to obtain 14 readings per period in which $Q_L$ii was calculated.

The $K(H)$ relation was established in a separate plot, near to the experimental area, due to the impossibility of making this on the five replicates of 12 m^2 on which P, I, RO and ΔS were continuously measured. In this extra plot, coffee plants were cut keeping their root systems in order to maintain the soil under nearly the same conditions as in the five plots where the water balances were carried out. Reichardt et al.'s (2004) method, which is based on previous soil saturation by means of infiltration followed by internal drainage, was used. A set of concentric rings (1.5 m internal diameter and 2.5 m external diameter) received a 5.0 cm water depth for more than 6 h, time enough to reach the dynamic equilibrium, judged through constant tensiometer readings, for the measurement of the saturated hydraulic conductivity K_0 (mm day^{-1}). Three neutron-probe access tubes were installed in the central ring for measuring volumetric soil water contents (θ, m^3 m^{-3}) at the same depths used in the experimental plots (0.2, 0.4, 0.6, 0.8 and 1.0 m); and three tensiometers at 0.9 and 1.1 m to measure the gradient ∇H and the total water potential (H), taken as an average of all six ($\bar{H}$), to represent the value at $z = 1.0$ m. This procedure was also applied at the five replicates within the coffee crop.

After the infiltration ($t = 0$, for the drainage), the rings were covered with a plastic sheet in order to avoid water losses by evaporation and measurements of $\theta_{(z,t)}$ and $H_{(z,t)}$ started in increasing time intervals, for 13 days. According to the method proposed by Reichardt et al. (2004), these data were used in semi logarithmic regressions to obtain the parameters for a $K(\theta)$ relation at a chosen depth $L = 1.0$ m:

$$K = K_0 \exp[\beta(\theta - \theta_0)] \tag{5}$$

in which K_0 and θ_0 are the field saturated values of K and θ.

The regressions suggested by the method are:

$$\theta_L(t) = a + b \ln t \tag{6}$$

$$S_L(t) = c + d \ln t \tag{7}$$

where S_L (mm) is the soil water storage of the 0–1.0 m layer, and

$$H_L(t) = e + f \ln t, \quad \text{for } z = 0.9 \text{ and } 1.1 \text{ m} \tag{8}$$

Once these regressions are made, and if they are statistically significant, their parameters are substituted in the following equation:

$$K(\theta)_L = \left\{ \frac{[(-d\exp(a/b))][\exp(-\theta/b)]}{[(e' + (f'/b))(\theta - a)]} \right\} \quad (9)$$

where $e' = (e_1 - e_2)/0.2$ and $f' = (f_1 - f_2)/0.2$; e_1, e_2, f_1, and f_2 are the coefficients of Eq. (8) for $z = 0.9$ and $z = 1.1$, respectively. Using relations (6) and (8), the $K(\theta)$ relation was transformed into a $K(H)$ relation, to be used in the water balance calculations.

3. Results and discussion

Results of all water balance components shown in Eq. (1) are presented elsewhere (Silva et al., 2006). Here, we focus on the estimation of the deep drainage Q_L made through approaches (i) and (ii). Calculations of Q_Li are straightforward and presented in Table 1, showing individual values for each replicate, their average $\bar{Q}_L$i, standard deviation (S.D.), and coefficients of variation (CV). Through approach i, the variability of Q_Li is mainly due to the variabilities of P and ΔS, since I and RO are either zero or very small. P had CVs of the order of 4% and in the case of ΔS, several times they lose their meaning. As ΔS approaches 0, which can occur by coincidence in any balance where $S_L f$ is close to S_Li, the standard deviation becomes larger than the average, and the CV increases very much. The magnitude and variability of the S.D. for S_L was, however, in the order of the other components. The CVs of Q_Li presented in Table 1 reach high values for balances containing replicates with values 0.0, which in the same way have no meaning. Their standard deviations are larger than their respective small averages, indicating that they do not differ significantly from zero. In this way, balances 11, 12, 13, 20 and 24 are the main contributes to Q_Li, with CVs in the range of 6.1–22.6%. For the whole year, Q_Li summed up to 245.5 mm.

The saturated soil hydraulic conductivity (K_0) and the saturated soil water content (θ_0) for the $K(H)$ relation used to estimate the drainage $Q_L$ii using the Darcy–Buckingham approach were: 468.64 mm day^{-1} and 0.459 m^3 m^{-3}, respectively, measured when the infiltration reached the dynamic equilibrium monitored through tensiometers readings. The series of measurements of $\theta(z, t)$ and $H(z, t)$ obtained during the internal drainage process at increasing time intervals allowed the establishment of the regressions indicated by the Eqs. (6)–(8), shown in Figs. 1–3.

Changes of θ with time, during interval drainage process are somewhat scattered, however, presenting a

Table 1
Deep drainage below root zone ($z = 1.0$) calculated by approach (i) (Q_Li) for five replicates

Balance number	DAB	Q_Li (mm)					$\bar{Q}_L$i	S.D.	CV
		1	2	3	4	5			
3	28–42	−7.0	−0.9	0.0	0.0	0.0	−1.6	3.1	193.0
6	70–84	−3.5	−0.4	0.0	0.0	−4.0	−1.6	2.0	125.0
7	84–98	−8.6	0.0	0.0	0.0	−3.4	−2.4	3.8	157.1
8	98–112	−1.9	0.0	−4.3	0.0	0.0	−1.2	1.9	152.9
9	112–126	−10.7	−5.2	0.0	0.0	−2.7	−3.7	4.5	120.1
10	126–140	−0.5	−0.3	0.0	−5.9	0.0	−1.3	2.6	191.0
11	140–154	−41.2	−59.1	−69.4	−68.5	−46.2	−56.9	12.8	22.6
12	154–168	−37.8	−32.8	−26.6	−27.8	−32.8	−31.6	4.5	14.2
13	168–182	−82.1	−80.0	−89.5	−83.9	−76.0	−82.3	5.0	6.1
14	182–196	−5.9	−3.7	0.0	0.0	0.0	−1.9	2.7	142.5
15	196–210	0.0	−1.6	0.0	0.0	−2.6	−0.8	1.2	142.2
16	210–224	−0.6	−11.5	0.0	0.0	0.0	−2.4	5.1	209.1
17	224–238	−2.5	−5.8	−0.9	0.0	0.0	−1.8	2.4	133.7
18	238–252	0.0	−1.4	−1.2	−2.1	0.0	−0.9	0.9	97.2
19	252–266	−2.6	−3.3	0.0	−0.9	0.0	−1.4	1.5	111.4
20	266–280	−21.4	−25.2	−16.3	−22.5	−28.0	−22.7	4.4	19.3
21	280–294	0.0	0.0	−6.9	−1.2	0.0	−1.6	3.0	184.2
22	294–308	−1.1	−1.0	0.0	0.0	0.0	−0.4	0.6	137.1
23	308–322	−3.7	−8.9	0.0	0.0	0.0	−2.5	3.9	155.5
24	322–336	−19.2	−19.4	−22.3	−22.5	−19.9	−20.7	1.6	7.8
26	350–364	−8.8	−9.9	0.0	0.0	−10.4	−5.8	5.3	91.8

$\bar{Q}_L$i: average; S.D.: standard deviation; CV: coefficient of variation; DAB: days after beginning, starting 1 September 2003.

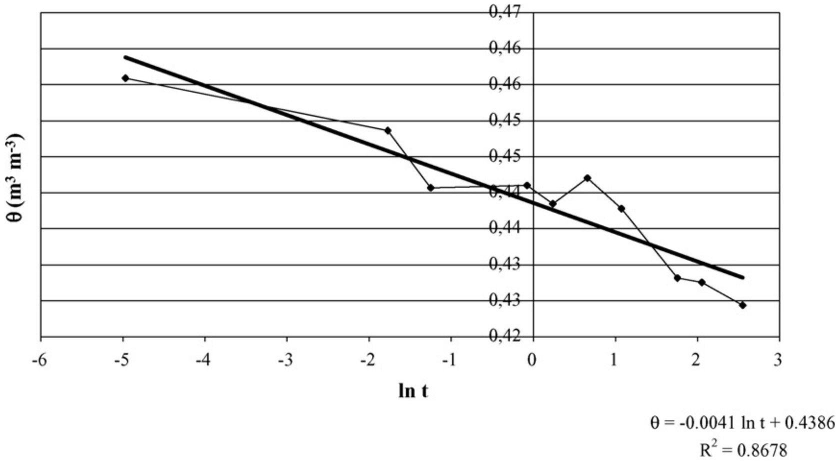

Fig. 1. Regression of soil water content θ ($m^3\ m^{-3}$) changes as a function of ln t (days), as indicated by Eq. (6), during internal drainage.

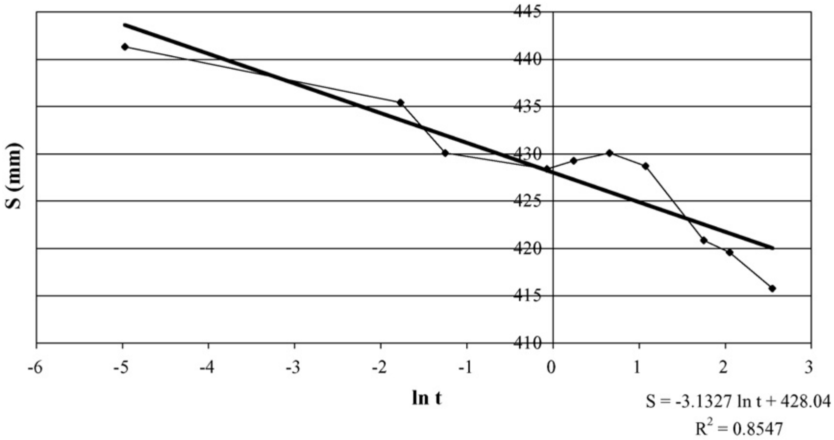

Fig. 2. Regression of soil water storage S (mm) of the 1 m soil layer as a function of ln t (days), as indicated by Eq. (7), during internal drainage.

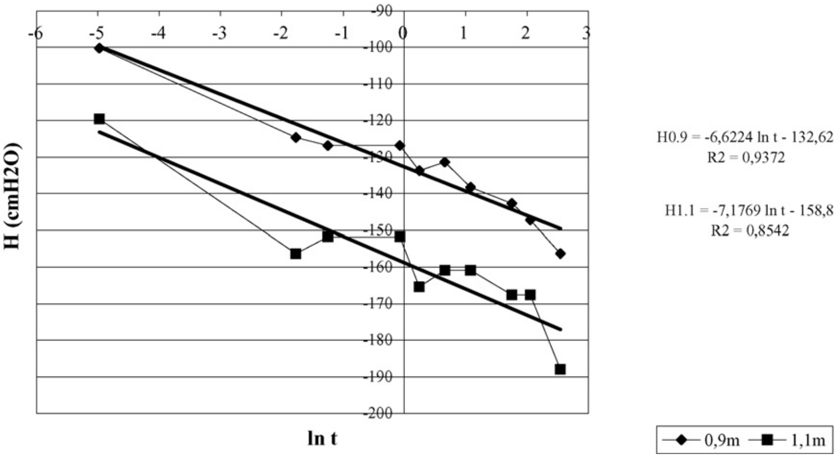

Fig. 3. Regressions of total soil water potential H (cmH_2O) at depths 0.9 and 1.1 m, as a function of ln t (days), as indicated by Eq. (8), during internal drainage.

significative R^2. This soil presents very high conductivities close to saturation, with a remarkable flow reduction after one day of drainage. As a result, decreases in θ are so small that values almost fall into the measurement error, causing the scatter of the data shown in Fig. 1. Similar comments can be made for the soil water storage changes, shown in Fig. 2. This behavior can be explained by the considerations of Othmer et al. (1991) who suggest that several soils present bi-modal pore size distributions, with interpedal pores that together with macropores conduct an accelerated water flux, and intrapedal pores with a much smaller conductivity than that of the whole soil. We believe that this is not our case since, at the depth of 1 m, primary and secondary peds are not present and the $K(\theta)$ relation represents well the whole soil at the measurement site. Between different measurement sites, however, spatial variability plays an important role.

Tensiometer readings shown in Fig. 3 also present scattering. As we know, these readings are temperature dependent and since they have been taken at different hours of the day, this effect might be the cause. Another problem is their reading precision, which is in the order of ± 10 cmH_2O, in readings from -100 to -250 cmH_2O, in the wet range. This accuracy is common for Bourdon type manometers. However, the R^2 of the regressions shown in Fig. 3 were also significant. In this way, these regressions supplied the parameters for the $K(\theta)$ relation, which are: $a = 0.4386$, $b = -0.0041$, $d = -3.1327$, $e' = 1.309$, and $f' = 0.0277$:

$$K(\theta)_L = \left\{ [(-(-3.1327)\exp(0.4386/-0.0041))] \frac{[\exp(-\theta/-0.0041)]}{[(1.309+(0.0277/-0.0041))(\theta-0.4386)]} \right\} \quad (10)$$

Which transformed into $K(H)$, resulted in:

$$K(\bar{H}) = \frac{4.651757 \times 10^9 \exp(\bar{H} - 21.57/6.8996)}{0.723486 - 0.004018 \times (\bar{H} - 21.57)} \quad (11)$$

in which $\bar{H}$ is the average of the six tensiometers installed in each plot, three at 0.9 m and three at 1.1 m, representing, therefore, the average potential at the lower boundary ($z = 1.0$ m). Fig. 4 shows the variations of $\bar{H}$ as a function of DAB, where two main drainage periods can be noted: the first period from DAB 130–200, and the second, from DAB 260–340. During these periods, the readings were intensified and by means of curve adjustment interpolation made using Microsoft Excel® 14 values of $\bar{H}$, grad(H), and $K(\bar{H})$were obtained for calculating $q_L$ii and $Q_L$ii (Eqs (2)–(4)). An example of these calculations is shown in Table 2.

Eq. (11) obtained from a distinct plot in relation to the five experimental plots used to establish balances, when used in these plots for the calculations of water balance, yielded extremely low K values for all water balances including balances 11, 12, 13, 20 and 24 which presented large values of Q_Li. This is a typical problem of the use of the $K(\theta)$ or $K(H)$ relations in the field, due to site spatial variability. Two main aspects have to be considered.

Firstly, the soil space variability is the factor that mostly affects the parameters used in $K(\theta)$, $K(h)$ or $K(H)$ relations. This variability was studied by several researchers (Greminger et al., 1985; Villagra et al., 1988), who indicated its importance and discussed its implications on the calculations of field soil water fluxes. Villagra et al. (1995) mentioned that most field K methods require a considerable amount of expensive and time consuming equipment on both, upper and

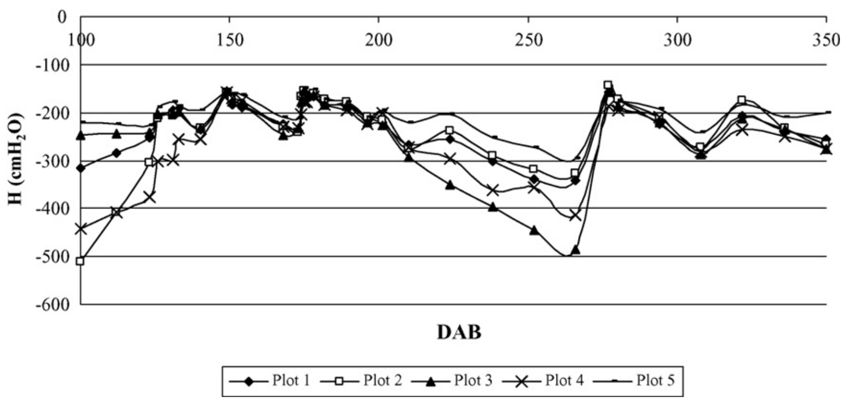

Fig. 4. Changes of the total soil water potential $\bar{H}$ (average of six tensiometers per plot) as a function of time. DAB: days after beginning.

Table 2
An example of deep drainage calculation showing values of $\bar{H}$, grad(H), $K(\bar{H})$, $q_L$ii and $Q_L$ii for balance 11

DAB	$\bar{H}$ (cmH$_2$O)	Grad(H) (cm cm^{-1})	$K(\bar{H})$ (mm day^{-1})	$q_L$ii
141	−222.5	−1.5	0.0007	0.0
142	−214.1	−1.6	0.0023	0.0
143	−205.1	−1.7	0.0086	0.0
144	−195.6	−2.0	0.0350	−0.1
145	−188.0	−1.9	0.1076	−0.2
146	−179.1	−2.0	0.3975	−0.8
147	−171.3	−2.1	1.2759	−2.7
148	−165.1	−2.1	3.1545	−6.6
149	−157.4	−2.3	9.8671	−22.6
150	−160.1	−2.3	6.6416	−15.3
151	−172.8	−2.2	1.0090	−2.3
152	−174.0	−2.1	0.8533	−1.8
153	−175.7	−2.1	0.6571	−1.4
154	−178.5	−2.2	0.4366	−1.0
$Q_L$ii (mm)				−54.6

lower limits of the volume element, implying in a significant limitation for space variability studies which require a large number of replicates. As a consequence, few studies discuss the difficulties imposed by the space variability of the K relations when used in the field and in water balance calculations (Timm, 2002).

Secondly, the exponential characteristic of the $K(\theta)$ and $K(\theta)$ relations is the other factor: for very small θ, h or H variations, large K variations result. For our Eq. (11), which is valid within the interval $H = -131.4$ to -350 cm, for $H = -131.4$ (field saturation point) the value of $K_0 = 468.64$ mm day^{-1} is obtained. This is a very high value (almost 0.5 m day^{-1}), typical of sandy, permeable soils. Eq. (11) is strongly exponential, drastically reducing K for small variations in H. Table 3 shows K variations in relation to H variations in steps of -10 cmH$_2$O. This step ΔH is nearly the precision of the tensiometers used in this study and a common precision

Table 3
An example of variations in $K(H)$ for 10 cm steps for Eq. (11)

H (cmH$_2$O)	$K(H)$ (mm day^{-1})
−131.4	468.64
−140	130.8310
−150	29.7141
−160	6.7557
−170	1.5375
−180	0.3502
−190	0.0798
−200	0.0182
−210	0.0042
−220	0.0010
−230	0.0002

value for this kind of measurement. It can be observed that K varies 340% for each of these variations in H. This fact is discussed in the literature and contributes to the failure of the use of Darcy's equation in field soils, like Warrick and Nielsen (1980) who showed data with coefficients of variation of 100% for the saturated soil hydraulic conductivity, and of 400% for non-saturated soil hydraulic conductivity exposed to lower moistures. Despite of all this, Libardi and Saad (1994) comment that it is common to find water balance studies based on only one hydraulic conductivity measurement for the experimental area, leading to errors in deep drainage estimates or in capillary rise.

As mentioned above, since our $K(\bar{H})$ relation yielded negligible soil water fluxes when applied to the five balance plots, we tried to adjust Eq. (11) in order to overcome the problem of spatial variability between the extra plot in which $K(\bar{H})$was determined and obtain consistent results. For this, we considered the procedure (i) for calculating Q_Li as correct, which is reasonable, and Eq. (11) was adjusted so that for balance 11, $Q_L\text{i} = Q_L\text{ii}$ for average H data of the five plots. The adjustment consisted of a correction in $\bar{H}$, which was replaced by $\bar{H} - 21.57$. This correction means that, due to the space variability of the soil under study, the average $\bar{H}$ value measured in the plot where the function $K(\bar{H})$ was obtained, differs in 21.57 cm from the average obtained from the $\bar{H}$ values from all the five water balance plots. This is a small correction that can be very well justified being aware of the variability of soil water retention curves of field soils, however, large enough to make Eq. (11) usable. Once adjusted for water balance 11, Eq. (11) was maintained with the same factor for all other calculations.

Table 4 presents individual values of $Q_L$ii, their averages $\bar{Q}_L$ii, standard deviations, and coefficients of variation, for all balances presenting deep drainage by approach (ii). The variability between replicates is much higher, confirming the above discussion. The CVs varied from 77.9 to 176.0%, not taking into account balance 22 which presents a too low average and a consequent CV of 182.3%. These coefficients of variation are much higher than those of Q_Li. The total amount of drainage reached 391.2 mm in the whole year, almost the double of the amount obtained through approach (i). These high values are actually unrealistic since they yield underestimated values of actual evapotranspiration ER when introduced into the balance Eq. (1). For the balances of high $\bar{Q}_L$ii (balances 11, 13 and 20), the application of Eq. (1) yielded positive values of ER, which are contradictory. Based on these considerations, we made the option of calculating Q_L

Table 4
Deep drainage below root zone (z = 1.0) calculated by approach (ii) ($Q_L$ii) for five replicates

Balance number	DAB	$Q_L$ii (mm)					$\bar{Q}_L$ii	S.D.	CV
		1	2	3	4	5			
11	140–154	−15.8	−69.7	−64.7	−64.3	−183.4	−79.6	62.0	77.9
12	154–168	−0.3	−0.9	−0.5	−2.3	−19.4	−4.7	8.2	176.0
13	168–182	−39.4	−106.6	−64.7	−15.4	−264.7	−98.2	99.0	100.9
14	182–196	−5.0	−19.8	−3.7	−1.3	−19.0	−9.8	8.9	91.3
20	266–280	−71.5	−550.6	−120.3	−1.3	−140.2	−176.8	215.7	122.0
21	280–294	−14.7	−22.2	−2.5	−0.5	−15.5	−11.1	9.3	83.6
22	294–308	0.0	−0.1	0.0	−0.1	−1.0	−0.2	0.4	182.3
23	308–322	−0.13	−18.4	−0.1	0.0	−8.4	−5.4	8.1	150.2
24	322–336	−0.1	−18.4	−0.1	0.0	−8.4	−5.4	8.1	149.9

$\bar{Q}_L$ii: average; S.D.: standard deviation; CV: coefficient of variation; DAB: days after beginning, starting 1 September 2003.

through methodology (i), disregarding the results of methodology (ii).

4. Conclusions

Water balances were established to estimate internal drainage below the root zone of a coffee crop and evaluate two methodologies using deterministic approaches: (i) based mostly on atmospheric parameters in the water balance equation; and (ii) based on the Darcy–Buckingham flux-gradient equation. Soil spatial variability, the exponential character of the hydraulic conductivity functions, and the precision of soil water content and soil water potential measurements in the field did not yield consistent estimates of drainage through methodology (ii), leading to the adoption of methodology (i).

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