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# Capillary rise quantification by field injection of artificial deuterium and laboratory soil characterization

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## Abstract

In arid contexts, water rises from the saturated level of a shallow aquifer to the drying soil surface where evaporation takes place. This process plays important roles in terms of plant survival, salt balance and aquifer budget. A new field quantification method of this capillary rise flow is proposed using micro-injections (6 µL) of deuterium-enriched solution ($\delta$ value of 63 000‰ vs. V-SMOW) into unsatured soil at 1 m depth. Evaluation of peak displacement from a profile sampling 35 days later, delivered estimates that were compared with outputs of numerical simulation based on laboratory hydrodynamic measurements. A rate of 3.7 cm $y^{-1}$ was observed in a Moroccan site where the aquifer level was 2.44 m deep. This value was higher, than other estimates based on natural diffusion with the same depth of aquifer, but lower than the estimates established using integration of van Genutchen closed-form functions for soil hydraulic conductivity and retention curve.

## 1 Introduction

In arid contexts with long periods of high evaporative demand, water may rise from the saturated level of a shallow aquifer to the near surface of the drying soil where evaporation takes place. Although resulting evaporation rate is known to decrease quickly with the aquifer depth, this process may play important roles in terms of plant survival, salt balance and aquifer budget. Field quantification of capillary rise in soil is uneasy to perform. Firstly, the stability of water contents along the soil profile does not allow the closure of traditional soil budget to be performed by using time differences in soil water stocks. Secondly, high suction values in soil (lower than −800 cm), frequently encountered near the soil surface, outrange tensiometer measurement capability. Fluxes smaller than 100 mm $y^{-1}$ are not easily measured except by external surface measurements like the eddy-covariance method (Garcia et al., 2009). Significant advances were achieved in the early eighties by field methods, conceived to compute capillary rise rates, using natural equilibrium between convection and diffusion of natural stable isotopic species ($^2$H,$^{18}$O) in the liquid and vapor phases of the vadose zone profiles assumed to be in a so called "pseudo steady-stage of evaporation" (Allison, 1982; Barnes and Allison, 1983, 1984; Barnes and Hughes, 1983; Fontes et al., 1986). In more recent studies, this methodology was extended to others contexts (Liu et al., 1995; Yamanaka and Yonetani, 1999; Grünberger et al., 2008) and modeling was refined, taking in account dynamics of drying processes and interactions with plant evapo-transpiration (Walker et al., 1988; Barnes and Walker, 1989, Shimojima et al., 1990; Walker and Brunel, 1990; Brunel et al., 1995, 1997; Melayah and Bruckler, 1996a,b; Braud et al., 2005a,b). The isotopic content profile allowed the discrimination to be done between the top layer where only vapor fluxes take place, and the deeper layers where liquid-phase fluxes are dominant. A better understanding of the partition processes between water and vapor fluxes was the main output provided by these studies, but concerning field flux measurements, some uncertainties remained. Indeed, isotope soil profile sampling cannot be repeated in the same place, and evaporation rates calculated are "instant rates" submitted to night and day alternation and temperature changes, and may differ from time average value. Gardner (1958) was the first author to propose a relationship based on a specific integration of the Richards equation to determine capillarity rise rate. The relationship links depth of the saturated level, the soil hydrodynamic characteristics and an ascending rate assumed to be constant. This relationship was confirmed by measurements on experimental columns of disturbed or undisturbed-soils submitted to capillary rise under different conditions (Gardner and Fireman, 1958). Although it did not take in account the presence of a superficial layer where the vapor fluxes are dominant and did not represent layered soils, this method has been widely used to compute the relation between the depth of the water table and the stabilized evaporation flux. A comparison between fluxes estimates based on Gardner approach and isotopic profiles has been performed by Coudrain-Ribstein et al. (1998), who concluded that the scattering of the rates estimated by Gardner's method could be caused by the lack of measurements of soil



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permeability reflecting vapor transfer capabilities. The introduction of van Genutchen (1980) "closed-form equation" to represent soil permeability, was used recently by Hu et al. (2008) to perform a theoretical analysis of the limiting rate of phreatic evaporation.

Putting apart the profiles of natural isotope contents in soil water already evocated, a large number of studies considered the use of artificial isotopic species of water as tracers in soil hydrodynamic studies (Koeniger et al., 2010). However, field and laboratory soil hydrodynamic studies involving isotopic tracers mainly focused on infiltration and/or mixing fronts and implied large volumes of tracing solution resulting in almost (or totally) saturated conditions. Then, to our knowledge, artificial tracing was never used to quantify field capillary rise, with the exception of assessment of radioactive contamination by tritium in dry environments (Garcia et al., 2009).

Our objectives were to assess a new method of field measurement of capillary rise flow and to bring out a sound base of comparison with the results of simulation based on laboratory hydrodynamic measurements. We based our approach on the practices of ecophysiologists, who perform small volume injection of concentrated solution to quantify tree sap-flow (Marc and Robinson, 2004). A small quantity of artificial $D_2O$ was injected into a soil profile, assumed to be at "steady state" of evaporation, and the displacement of the tracer was measured 35 days later.

## 2 Materials and methods

### 2.1 Site description

The experiment site is located at 30° 22′ 05″ N 09°34′ 38″ W near Agadir City, Moroco. The experiment was performed in 2001 on a sandy soil with a shallow free aquifer (Bouchaou et al., 2008) at some distance of the coast on a fine aeolian sand cover with a modal size around 100 µm and moderate carbonate contents (Weisrock et al., 2002). The annual rainfall was exceptionally low in 2001 with a value of 86 mm, whereas the average annual rainfall recorded for 1961–2004 period at Agadir meteorological station



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was 255 mm $y^{-1}$ (Stour and Agoumi, 2008). The experiment was started during the dry season, on the 16 May 2001. The piezometric level of the free aquifer was at 2.44 m depth and remained at the same depth 35 days later. At the nearest meteorological station, no rain was recorded between the tracer injection day and the soil sampling day. A small rain event (0.7 mm) occurred 25 days before the injection and two rains occurred 70 and 75 days before the injection with a total amount of 3.5 mm. Relative humidity was monitored with a moisture capacity probe at the soil surface and at four different depths during the sampling day (24 h). The values observed at the soil surface ranged from 0.31 at midday to 0.89 at night, while at 31 cm depth the observed values ranged from 0.83 to 0.89.

### 2.2 $^{2}H$ Injection, sampling and measurements

The 16 May 2001, a squared pit was excavated down to more than one meter depth. On the wall of the pit a horizontal injection line was settled at 1 m depth (Fig. 1). This depth was chosen assuming that, at depth between 50 cm and 100 cm, in the soil layer, transfer in vapor phase was not dominant and negative potential heads were not higher than –800 cm. Fifty injections points were defined by driving horizontally 47 mm-long medical needles spaced 1 cm apart in the soil along a horizontal line on a side of the pit. The injections were performed through the medical needles that played the role of catheters, as the thinner needle of a 10 µL micro syringe was inserted inside to inject 6 µL of deuterium-enriched solution. This artificial solution was made by mixing a small amount of heavy water $D_2O$ (99.9%) with water, resulting in a $R_{D/H}$ of 0.009968 (corresponding roughly to a $\delta$ value of 63 000 $\delta$‰ vs. V-SMOW). The total input corresponded to $1.67 \times 10^{-4}$ moles of $D_2O$. Spray-paint was used to seal pit wall surface in order to: i) prevent drying during the injection time, ii) mitigate the effects of the disturbance due to the pit, iii) ease correct placement of the sampling as a land mark. After the injection, the medical needles were not removed from the soil, and the pit was filled up respecting the sequence of the extracted soil. Thirty five days later, the soil sampling was performed above the injection line to retrieve the tracer. A steel frame of 0.15 m by 0.05 m was used to locate the sampling column adjacent to the spray-paint coating. Twenty samples were collected and treated by distillation to dryness under vacuum. The D concentrations were determined after reduction of the distilled water by reaction with zinc (Coleman et al., 1982) using a mass spectrometer VG602C. The results are reported in $\delta$-notation, permil deviation of the measured isotopic ratio relative to V-SMOW international standard, with an uncertainty lower than $\pm 2\delta$‰. The injection of tracer produced a sharp peak of content in the soil profile. The peak displacement with time may be interpreted as the result of the cumulative convective upward flow. The amount of water between the injection depth and the final peak depth represents the volume of displaced water during the period between injection and sampling.

$$E_{35\,\mathrm{days}} = \int_{\mathrm{inj}}^{\mathrm{peak}} \Theta(z).dz \qquad (1)$$

With $E_{35\,\mathrm{days}}$ being the flux during the 35 days (cm), inj: the injection depth (cm), peak: the peak depth (cm), $\Theta(z)$ the measured volumetric water content.

### 2.3 Soil parameters and rise rate computation

Soil hydrodynamic parameters were characterized on five undisturbed samples taken on a side of the pit using the evaporation method (Wind, 1968). The Software Hydrus 1-D.4.13 was used to calculate Van Genutchen (1980) hydrodynamic parameter values by inversion (Simunek et al., 1998). The inversion was based on observation nodes data (8 couples of time and piezometric heads and 8 couples of time and water contents). Initial input data was based on values computed by the manual method in a spread sheet file and consistency was checked. Real saturated water contents were kept unchanged. Results are presented in Table 1. We respected van Genutchen



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(1980) formalism, $S$ being the saturation index, $\Theta_r$ and $\Theta_s$ the residual and saturation contents, $\psi$ tension head [L], $K(\psi)$ the hydraulic conductivity [L.T$^{-1}$] and $K_s$ the saturated hydraulic conductivity [L.T$^{-1}$], $\alpha$ ([L$^{-1}$], cm$^{-1}$), $n$, $m = 1-1/n$, the other van Genutchen's parameters, using the relations:

$$S = \frac{\Theta - \Theta_r}{\Theta_s - \Theta_r} = (1 + (\alpha\psi)^n)^{-m} \tag{2}$$

$$K(\psi) = K_s S^{1/2}(1-(1-S^{1/m})^m)^2 \tag{3}$$

Vertical flux is described by Richards' differential equation linking the flux $E$ ([L.T$^{-1}$], cm min$^{-1}$) and the hydraulic conductivity ([L.T$^{-1}$], cm min$^{-1}$) to the variation of the potential head $\psi$ ([L], cm) along the depth $Z$ ([L], cm$^{-1}$):

$$E = K(\psi).\frac{d\psi}{dz} - K(\psi) \Leftrightarrow dz = \frac{K(\psi)}{E + K(\psi)} d\psi \tag{4}$$

At steady state, between two depths $z_1$ and $z_2$ ([L], cm), stabilized at the constant potential heads $\psi_1$ and $\psi_2$ ([L], cm), the flux $E$ is constant and ruled by the integration of Eq. (4):

$$\int_{z_1}^{z_2} dz = \int_{\psi_1}^{\psi_2} \frac{K(\psi)}{E + K(\psi)} d\psi = z_2 - z_1 \tag{5}$$

In a "steady state" evaporation profile, when the soil characteristics relationships $K(\psi)$ and $\psi(\theta)$ are known, then the vertical convective flux ($E$) between these two points can be computed. A particular starting point of integration may be the underground water level where soil saturation implies that tension head is equal to zero (then $z_1 = 0$ and $\psi_1 = 0$). Potential maximum evaporation flux may be computed setting $z_2$ to the depth of underground water level and $\psi_2$ to $\infty$. When the soil saturation depth is known, the knowledge of another potential head at a different depth may lead to another estimate of evaporation. The Eq. (5) was solved numerically by the Eq. (6) developed in a spreadsheet using 36 600 steps of integration:

$$z_2 - z_1 \approx \sum_{i=0}^{i=36600} \frac{K(\psi(i))}{E + K(\psi(i))} \Delta\psi(i) \tag{6}$$

using Eq. (2) and with:

$$\psi(i+1) = \psi(i) + \Delta\psi(i) \tag{7}$$

computing proportional integration steps:

$$\Delta\psi(i) = \lambda\psi(i) \tag{8}$$

or using a constant for the integration steps:

$$\Delta\psi(i) = \gamma \tag{9}$$

Uprising fluxes were estimated, for each set of unsaturated soil characteristics established by Wind's method, using 3 distinct sets of integration limits:

The first integration (i) was computed between the aquifer level (saturation depth) at 244 cm depth where $\psi(0) = 0.0001$ and the soil surface with a maximum tension head of $\psi(i = 36600) = -1.64 \times 10^6$ cm, corresponding to the minimum air relative humidity of 31% observed at the soil surface, and the parameter $\lambda$ was set to 1.000624294 (Eq. 8).

Second integration (ii) was calculated between the saturation depth at 244 cm where $\psi(0) = 0.0001$ cm and the 50 cm depth where water contents was set to the measured value (0.0495). The parameter $\lambda$ was set in order to reach this value for the same number of integration steps (Eq. 8).

The third integration (iii) considered, the lower limit at 100 cm depth, where water content was set to 0.076227 and a upper limit, set to 37 cm depth, where water contents was 0.03546 in order to mimic the trend of average water content increase. Due to the





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smaller gap of tension heads constant $\gamma$ was set in order to reach the same number of integration steps (Eq. 9).

The results of integration methods (ii) and (iii) were compared to numerical simulation computed with Hydrus 1D4.13 with the same parameters set and boundary conditions. The initial condition for water content was settled to an increasing linear trend with depth from a water content of 0.1 in surface to saturation at 244 cm depth. Total run time was set to 100 days and stability of fluxes and water contents was always observed after 20 days. The value reported is the value of "surface flux" for 100 days.

## 3 Results

### 3.1 Soil water contents and van Genutchen parameters

The profile of soil volumic water contents is presented in Fig. 2. Soil water content between soil surface and 120 cm depth was always less than 0.097. Between surface and 40 cm depth, water content was inferior to 0.026. A linear increase trend of soil water content with depth was observed with a low correlation ($R^2$ =0.66) and a slope of $6.64 \times 10^{-4}$ ($\theta\,cm^{-1}$). This correlation was stronger ($R^2$ =0.84) between 60 and 97 cm depths with a slope of $1.543 \times 10^{-3}$ ($\theta\,cm^{-1}$). The scattering of the values with depth indicates that the retention parameters were quite heterogeneous at small scale. Saturation water content was quite constant along the profile, ranging from 0.29 to 0.31 (Table 1). Parameter $n$, ranging from 1.6 to 2.2 is representative of Sandy Loam, Loamy Sands and Loams and parameter $\alpha$ ranging from 0.029 to 0.056 $cm^{-1}$, typically corresponds to Loams and Sandy Loams. Observed saturated hydraulic conductivity ranged from 0.25 to 0.66 cm $min^{-1}$, values associated to fine sands and Loamy Sands. The small size of the sand particles (100 µm) (Weisrock et al., 2002) may explain the high hydraulic conductivity values combined to low $\alpha$ values. The application of Eq. (2) showed that tension heads corresponding to water content value of 0.0495 (the value at 50 cm depth) ranged from 455 to 266 cm with two groups of similar values. First group corresponded to sampling characterization depths of 37 and 77 cm and second one to the samplings at depths 53, 92 and 112 cm.

### 3.2 Fluxes estimates using soil laboratory characterization

Computed vertical fluxes for 35 days assuming steady state evaporation with stability of water contents are reported in the Table 1. Cumulative fluxes range from 0.59 to 3.46 cm for 35 days (corresponding to 61–390 mm $y^{-1}$). The geometric means of estimates by the 3 methods ranges from 0.8 to 1.2 cm for 35 days (89.7–146 mm $y^{-1}$), with an overall average of 1.21 cm for 35 days corresponding to 126.7 mm $y^{-1}$. The comparison of the results obtained by methods (ii) and (iii), with Hydrus runs estimates, shows that, although the order of magnitude of fluxes is similar in all situations but one (37 cm depth, method (iii)), Hydrus 1-D, estimates are higher than those computed directly by the integration function, the difference between geometric means being less than 10% with the two methods (ii) and (iii). The influence of an error on the field measured volumetric water content of 1% could be characterized on the estimates given by method (ii). A negative error of 1% on water content would lead to an over-estimation of 0.19 to 0.37 cm for 35 days (corresponding to 19.8–38.5 mm $y^{-1}$), but a positive error of 1% would lead to an under-estimation that could reach 92% (53 cm depth parameters).

### 3.3 Injection results

A sharp peak of high $^2$H concentrations was observed in soil water (Fig. 3). Maximum $\delta^2$H content was +6.7‰ vs. V-SMOW at a distance of 4.7 cm from the injection point but at a distance of 56.8 cm (44.2 cm depth) the $\delta^2$H value was −11.7‰ vs. V-SMOW. Positive $\delta$ values were recorded between +9.5 cm and −5 cm distances from the injection point. The average soil water content between 100 cm and 95.3 cm depths being 0.0747, a 47 mm displacement corresponds to a flux of 0.351 cm in 35 days or 3.66 cm $y^{-1}$ (Eq. 1). In comparison, the values computed from the laboratory measurements were 1.6 to 9.84 times higher. The tracer recovering was computed by





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peak integration. The barycentre of $^{2}$H concentration weighted by water contents was determined using a initial $\delta^{2}$H content of −11.7‰ vs. V-SMOW and indicated an average distance of 2.4 cm. This result has to be mitigated due to the fact that only a few measurement points were performed under the injection point because a preliminary assumption was that the convective flux would be higher. As the sampling was performed on a determined surface imposed by the sampling steel framework, a recovering rate of tracer can be estimated. From $4.698 \times 10^{-5}$ moles of $D_2O$ injected into the soil column (on 0.15 m of the line of injection) $2.29 \times 10^{-5}$ moles were recovered in the 0.15 × 0.05 m sampling area corresponding to a recovering rate of 48.7%. The remaining tracer, non-recovered by sampling, probably moved outside the 5 cm strip above the needles in relation with diffusion and dispersion. A numerical simulation was performed with Hydrus 2-D in order to check the plume aspect and the coherency of the results taking in account that the paint coating sealed one side of the soil column (Fig. 4). Soil water flow parameters used in the simulation were those measured for 53 cm depth (Table 1), but according to the lower convective speed measured with the tracer, $K_s$ was set to 375 cm d$^{-1}$ to coincide with the maximum of contents (0.22 cm min$^{-1}$). The best fit with inverse modelling, was found for longitudinal and lateral dispersion of 8.15 cm and 2.24 cm, respectively (Fig. 4). The tracer recovering rate based on this run was computed to be 48.5% in very good agreement with the estimate based on measurements.

## 4 Discussions

The water flow values computed using soil characteristics determined at the laboratory were systematically higher than the capillary rise flux found from tracer displacement. The later is dependant on the evaluation of peak displacement and average water content of soil between the injection point and the peak. If uncertainties on tracer displacement and water content were ±0.5 cm and ±1%, respectively, then the evaporation flux would be between 2.7 and 4.4 cm y$^{-1}$ and therefore lower values could not easily be attributed to measurement error. Considering a soil volume of 1 $cm^3$, with a water content of 7%, the increase of water content due to the injection of the tracer solution (6 µL) represented less than 10% relatively to the initial content. Consequently the perturbation of water regime by the injection should be negligible, justifying the use of high $^2H$ concentration in injected solution. The stability of the evaporation state was a main assumption to compute fluxes using integration of the unsaturated hydraulic conductivity and tension head functions. To check that this equilibrium was reached, we used Hydrus 1-D and simulated two years of water movements in the soil profile, for all set of hydrodynamic parameters. The boundary conditions are representative, in the upper part, of the rain data and evaporation demand, and for the lower part, of a stabilized aquifer depth at 2.44 m. Regardless of initial water contents, the simulation indicated a steady-state evaporation regime between injection and sampling dates, consequently it can be assumed that discrepancy from steady-state water regime was not the reason of over estimation of fluxes. Cumulative flow of 0.351 cm between injection and sampling dates was coherent with the observation of a constant depth of aquifer level during this period, i.e. the cumulative flux would be related to an increase of saturation depth of 1 cm, which is roughly the accuracy of the piezometric measurement.

Gowing et al. (2005) emphasized the influence of vapour–liquid phase transition occurring in the upper parts of the profile on the evaporation flux. Measurements of relative humidity performed from the soil surface to 31 cm deep revealed that during the study period, the evaporation front was presumably located above a depth of 40 cm, where pore air moisture was saturated with water vapour, even during the mid-day maximum drought. Inside the water vapor transfer layer (VTL), processes are not well represented with the equations developed inside van Genutchen formalism and moreover tension head values higher than −800 cm are not easily measured, impeding correct extrapolations in very dry soil. Hence, the resolution of Richard's equation based on the extrapolation of the relationships defining hydraulic conductivity, tension heads and water contents is questionable. This would explain the over-estimations encountered when integration function is applied to high suctions, likely expressions





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involving, soil surface and/or VTL would easily lead to over estimation of fluxes, if the surface layers were involved. Nevertheless, for soil layers deeper than 50 cm, the water content values corresponded to low tension head falling within the range of the tension heads that were measured in the laboratory. Inside VTL, vapour flow processes may mitigate the evaporation flux, but quantifications based on water contents observed under the VTL, should correspond to actual fluxes. No hydrodynamic characteristics were determined for soil layers under 1 m depth because lower excavation would have endangered the stability of the floor of the pit due to increasing water contents near the aquifer water level. However, deeper soil layers could present lower permeability, particularly when getting close to the permanent water saturation level where pedogenesis processes may favor clay formation. In the case of the existence of deep layers with lower hydraulic conductivity values, discrepancy with fluxes should be observed between estimates i and ii based on extrapolations of soil characteristics to 2.44 m depth and estimate iii, based on 37 to 100 cm depths. As no strong differences of fluxes were observed in this study, the assumption could not be sustained. Based on the compilation of 20 different isotopic profiles Coudrain-Ribstein et al. (1998) proposed a relationship linking depth and evaporation flux that would imply a rate of 1.9 $\mathrm{cm\,y^{-1}}$ for the observed aquifer depth (2.44 m). Our measured rate is 1.8 time higher than this value.

## 5 Conclusions

Field tracer introduction in the wall of a soil pit, and sampling after 35 days delivered a peak displacement that could be interpreted as the result of a slow water rise from the aquifer. This study provided then an alternative method for estimation of water flux in soils to previous methods based on natural isotope contents and/or soil hydrodynamic features measuring. Compared to other ones, presented method is relatively sparing in assumptions: (i) The measured flux must be representative of all profile fluxes in the profile, but no hypothesis around steady state water regime is compulsory, (ii) apart from isotope measurements only density and water contents of the zone of the peak





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of tracer content are necessary (iii) The measurement is cumulative over a predetermined period which avoids considerations on night and day alternation and particular sampling time (iv) The method does not require knowledge of soil characteristics under and above the sampling zone. In the case of the studied experimental site a value of $3.7\,\mathrm{cm\,y^{-1}}$ was proposed. This value is higher than other estimates based on natural diffusion with the same depth of aquifer (Coudrain et al., 1998) but lower than the estimates established using integration of van Genutchen closed-form functions for soil hydraulic conductivity and retention curve.

## List of parameters

| | |
|---|---|
| $E$ | Convective flux ($\mathrm{cm\,d^{-1}}$) or ($\mathrm{mm\ y^{-1}}$) |
| $E_{35\,\mathrm{days}}$ | Cumulative convective flux during 35 days (cm) |
| z, inj, peak | Depth (cm), Injection depth (cm), Peak of isotopic content depth (cm) |
| $\psi$, $\psi(z)$, $\psi(i)$ | Tension heads (cm), at depth $z$, at step $i$ |
| $\psi_1$, $\psi_2$ | Integration limits for tension heads: tension 1 (cm), tension 2 (cm) |
| $\Theta$, $\Theta(z)$, $\Theta_r$, $\Theta_s$ | Volumetric water content, at depth $z$, residual, saturated ($\mathrm{cm^3\,cm^{-3}}$) |
| $S$, $S(\theta)$, $S(\psi)$ | Saturation index, at water content $\Theta$, at tension head $\psi$ |
| $K$, $K_s$, $K(\psi)$ | Hydraulic conductivity, at saturation, at tension head $\psi$ ($\mathrm{cm\,d^{-1}}$) |
| $\alpha$ | van Genutchen parameter ($\mathrm{cm^{-1}}$) |
| $n$, $m$ | van Genutchen Parameters ($m = 1-1/n$) |
| $z_1$, $z_2$ | Integration limits depths : depth 1 (cm), depth 2 (cm) |
| $\Delta\psi(i)$ | Integration step (cm) for potential head at step $i$ |
| $\lambda$, $\gamma$ | Value of integration steps: proportional step, constant step (cm) |





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**Table 1.** Van Genutchen parameters for 5 samples and resulting "steady state" fluxes for 35 days computed (using Hydrus or Eq. 5) from the aquifer depth (2.44 m) to soil surface or 50 cm depths, and from 1 m to 37 cm soil depth. (in bold: geometrical averages).

| | Van Genutchen parameters | | | | | Stabilized uprising computed fluxes for 35 days (cm) | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | | From saturation at 244 cm depth | | | | | From 100 cm depth to 37 cm depth | |
| Sampling depths | $\Theta_r$ | $\Theta_s$ | $\alpha$ | $n$ | $K_s$ | To surface $\psi = 1.64 \times 10^6$ cm | To determined water content ($\Theta$) at 50 cm depth $\Theta = 0.0495$ | $\Theta = 0.0495$ | $\Theta = 0.0595$ | $\Theta = 0.0395$ | $\Theta_{37\,cm} = 0.03546$ ($\Theta_{100\,cm} = 0.07623$) | |
| cm | – | – | $cm^{-1}$ | – | $cm\,min^{-1}$ | Eq. (5) | Hydrus | Eq. (5) | | | Hydrus | Eq. (5) |
| 37 | 0.012 | 0.291 | 0.0289 | 1.7770 | 0.2478 | 1.56 | 3.74 | 3.46 | 3.04 | 3.65 | 1.13 | 1.17 |
| 53 | 0.030 | 0.287 | 0.0304 | 2.2067 | 0.6615 | 0.58 | 1.62 | 1.41 | 0.10 | 1.78 | 2.15 | 1.82 |
| 77 | 0.008 | 0.314 | 0.0565 | 1.6320 | 0.6045 | 0.59 | 1.26 | 1.09 | 0.91 | 1.28 | 0.79 | 0.75 |
| 92 | 0.014 | 0.295 | 0.0413 | 1.8360 | 0.5996 | 0.72 | 1.34 | 1.33 | 0.51 | 1.68 | 1.87 | 1.60 |
| 112 | 0.014 | 0.310 | 0.0378 | 1.9157 | 0.3639 | 1.20 | 0.83 | 0.77 | 0.16 | 1.04 | 1.14 | 1.03 |
| Average | 0.016 | 0.299 | 0.0390 | 1.8735 | **0.4645** | **0.86** | **1.53** | **1.40** | **0.47** | **1.71** | **1.33** | **1.21** |

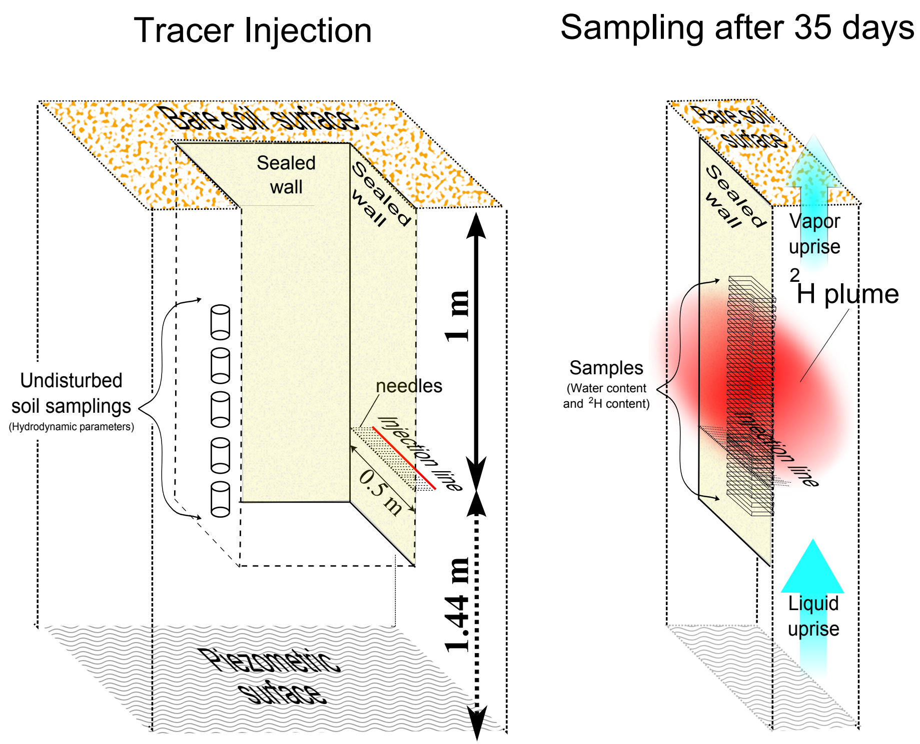


**Fig. 1.** Block diagram presenting injection pit and recovering sampling procedures of D tracer in the soil profile.





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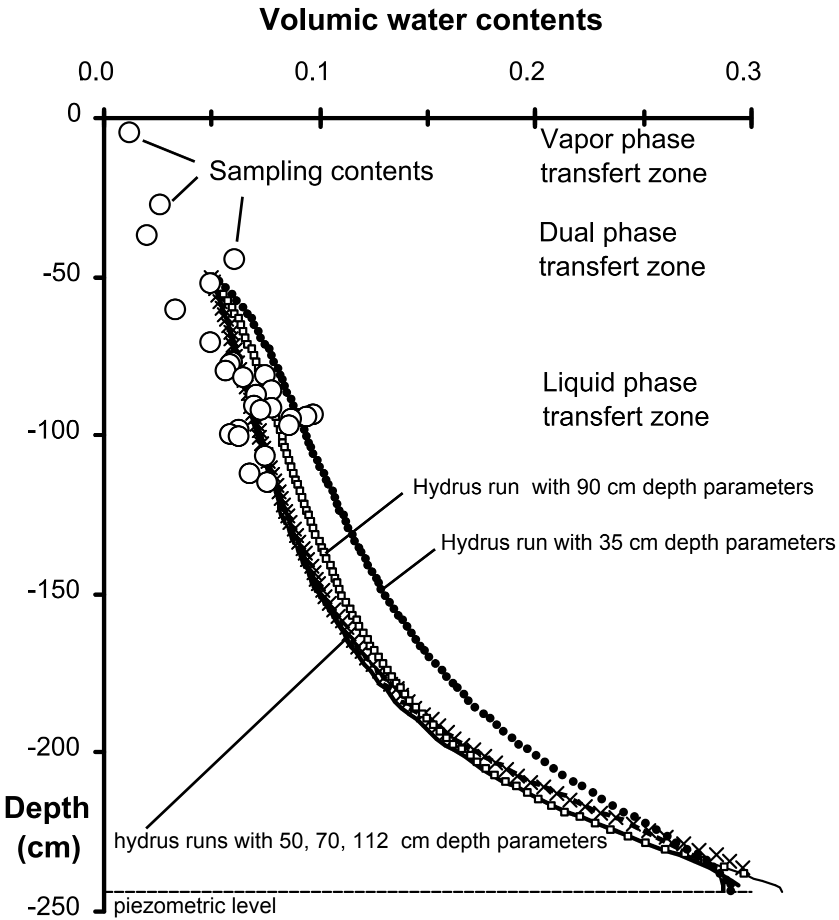


**Fig. 2.** Volumetric water contents profile the day of sampling and Hydrus 1-D stabilized water content profiles with the different set of hydrodynamic parameters measured at different depths by Wind's method.





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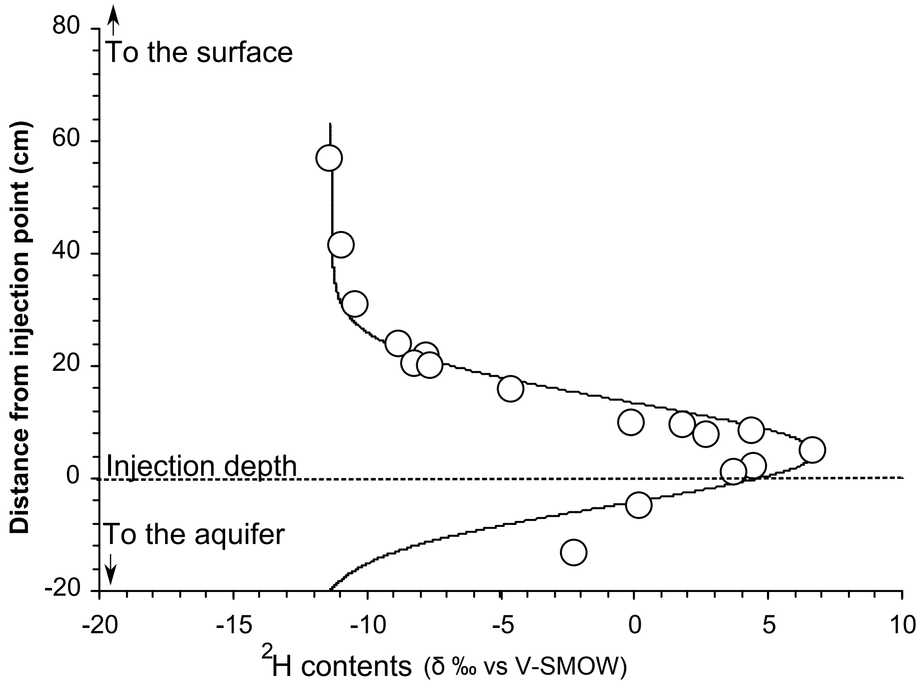


**Fig. 3.** $^2$H contents ($\delta$‰ vs. V-SMOW) of soil water in function of the distance to the injection point 35 days after tracer injection. Curve corresponds to the best Hydrus 2-D fit.





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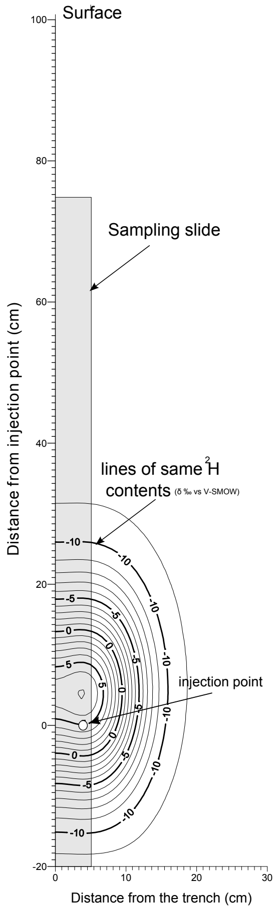


**Fig. 4.** Isocurves of $^{2}$H contents ($\delta$‰ vs. V-SMOW) after 35 days of tracer displacement simulated with Hydrus 2-D. $\Theta_r = 0.0304$, $\Theta_s = 0.287$, $\alpha = 0.0304\,\text{cm}^{-1}$, $K_s = 0.22\,\text{cm}\,\text{min}^{-1}$, $n = 2.207$. Longitudinal dispersivity = 8.15 cm, lateral dispersivity = 2.24 cm.





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