Estimating vertical soil water fluxes with tracers and time domain reflectometry (TDR) in a sand column under controlled laboratory conditions

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Kreba, S.A and C.P. Maulé. 2010. **Estimating vertical soil water fluxes with tracers and time domain reflectometry (TDR) in a sand column under controlled laboratory conditions.** Canadian Biosystems Engineering/Le génie des biosystèmes au Canada. 52: 1.9–1.17. There are a number of methods that have been developed and used for determination of soil water flux. Soil water flux determination is a complex measurement, requiring measurement of various components over a period of time. The objectives of this study are to show that time domain reflectrometry (TDR) can be a useful tool for estimation of soil water fluxes using tracer methods and to evaluate the accuracy of a tracer method in estimating soil water flow under mostly saturated conditions. A TDR system within a sand column with a KCl tracer was used. The TDR was used to simultaneously determine tracer concentration and moisture content. Two methods of flow determination were used: peak migration and soil water balance. Both methods were used for upward and downward flow directions. The peak migration method is based upon change in peak location with time and utilized TDR measurements. The water balance method is based upon changes in soil moisture content and the rate water is added to or taken from the system. The peak migration method provided similar average upward and downward water fluxes as compared to the soil water balance method. The average estimated upward soil water fluxes by the peak migration and the soil water balance methods were 10.2 and 11.3 mm d^{-1}, respectively, and were 91.9 and 90.2 mm d^{-1}, respectively for the downward fluxes. An advantage of the peak migration method is that it shows greater spatial and temporal resolution. **Keywords:** Soil water flux, time domain reflectrometry, tracer method, soil column.

Plusieurs méthodes ont été développées et utilisées pour la détermination du flux d'eau du sol. La détermination du flux d'eau du sol est une mesure complexe qui requiert la mesure de diverses composantes durant une période de temps. Les objectives de cette étude visent à démontrer que réflectométrie temporelle (TDR) peut s'avérer un outil pratique pour l'estimation du flux d'eau du sol utilisant des méthodes avec marqueur et pour évaluer l'exactitude d'une méthode avec traceur dans l'estimation du débit de l'eau dans le sol sous des conditions quasi-saturées. Un système TDR à l'intérieur d'une colonne de sable avec KCl comme marqueur a été utilisé. Le réflectomètre temporel a servi à déterminer la concentration du marqueur et la teneur en eau simultanément. Deux méthodes de détermination du débit ont été utilisées; migration du pic et équilibre de l'eau dans le sol. Les deux méthodes ont été utilisées pour des directions de débit ascendant et descendant. La méthode de migration du pic est basée sur le changement dans l'emplacement du pic avec le temps et utilise des mesures du réflectomètre temporel. La méthode d'équilibre de l'eau est basée sur des changements dans la teneur en eau et le taux d'eau ajoutée ou retirée du système. La méthode de migration du pic a fourni une moyenne de flux d'eau ascendante et descendante similaire à la méthode d'équilibre de l'eau dans le sol. La moyenne estimée des flux ascendants d'eau du sol par les méthodes de migration du pic et d'équilibre eau-sol était de 10.2 et 11.3 mm d^{-1} respectivement; et 91.9 et 90.2 mm d^{-1} respectivement pour les flux descendants. Un avantage avec la méthode de migration du pic réside dans le fait qu'une meilleure résolution spatiale et temporelle a été obtenue. **Mots clés:** flux d'eau du sol, réflectométrie temporelle, méthodes avec marqueur, colonne de sol.

INTRODUCTION

Water movement in the vadose zone between the bottom of the root zone and a deep water table is an important process for the contribution and loss of water and solutes from the root zone, the process of soil salinization, and potential long-term pollutant movement from the top soil to the groundwater zone. Groundwater is increasingly being used as a water source. As the world's population grows, concerns are being raised about the overall health of groundwater and the possibility of contamination. Vertical water movement within soils varies in direction due to climate conditions and vegetative demands. Upward movement of vadose groundwater can occur due to winter freezing (during the Canadian prairie winter, the ground may be frozen up to 2 m in depth) (Maclean 1974) or due to summer drying caused by evaporation and plant transpiration. On the other hand, extended periods of rain or spring snowmelt will result in excess water moving downward through the soil and into the vadose groundwater zone.

Both physical and tracer methods have been used to determine and study the groundwater and soil water fluxes; however, tracer methods are recommended especially in arid and semiarid environments (Gee and Hillel 1988; Allison et al. 1994; Scanlon et al. 1997) because the soil water fluxes are low and variable and chemical methods are easier to use (Christie et al. 1985; Zebarth and de Jong 1989; Gaye and Edmunds 1996; Joshi and Maule 2000; Le Gal La Salle et al. 2001). Other advantages of using tracer methods are that natural tracers can represent a spatially uniform input to the soil water and groundwater systems and some tracers are part of the water molecule or travel with water. However, there are several disadvantages of using tracer methods. Also, some tracers (e.g., Cl^{-}) may face the problem of anion exchange or exclusion in soils with high clay content (Hillel 1980). Using tracers that naturally occur within precipitation water requires long-term records of precipitation chemistry and of land use. Determining the groundwater recharge using tracer methods may be influenced by seasonality because seasonality may affect the downward flow rates and shape and concentration of the tracer profile (Kreba and Maule 2008).

Three techniques have been suggested (Allison et al. 1994) for estimating recharge rate from tracer profiles in the unsaturated zone: (1) from the total amount of tracer stored in the profile; (2) from the shape of the tracer profile in the soil; and (3) from the position of the tracer peak (the peak migration method). For this column study it was not possible to determine the soil water fluxes using the first technique. For the second technique there is insufficient information within the literature and, thus, just the peak migration method is considered in this review.

The peak migration method is based on the assumption that a volume of water equal to that present above the peak at the time of sampling has been displaced. Also, this method relies on steady-state flow and spatially uniform solute input assumptions (Joshi and Maule 2000). The peak method is used mostly to estimate recharge fluxes from tritium data (Smith et al. 1970; Allison and Hughes 1974; Gaye and Edmunds 1996) because the number of years elapsed since peak injection, 1963, is known. Piston flow through the unsaturated zone is also assumed (Daniels et al. 1991). Annual precipitation is assumed to be infiltrating as a slug and that it vertically displaces residual precipitation from the preceding year. Since tritium originates in the atmosphere and is deposited with precipitation, a low-high-low tritium profile will be recognized in the field. This reflects the movement of peak tritium concentration in the soil through time. The moisture content is taken as the average moisture content from the ground surface to the peak depth (Wood et al. 1997). However, most studies did not specify how the moisture content was calculated (Ward 2003). Potential problems with this method are the violation of the piston flow assumption and the absence of a distinct tritium peak (Allison et al. 1994).

To study the effects of climate on the soil water movement in semiarid regions, it is necessary to consider using the water balance method. The water balance method may be considered as similar to an accounting procedure where water inputs and outputs to the soil are algebraically added. It assumes that air temperature and day-length represent the energy required for evaporation and transpiration. Precipitation represents the water input and the soil moisture storage is regulated by assuming a maximum capacity based on the soil texture. Other operations regulate the water through the cycle for example, snowmelt, runoff, infiltration, and vegetative interception. The reliability of the calculated recharge of flux using this approach depends on the accuracy and precision with which these parameters can be measured (Gee and Hillel 1988).

Investigating the use of time domain reflectrometry (TDR) and the accuracy of using the tracer profile methods for estimating soil water and solute transports will improve understanding about the contribution and loss of water and solutes from the root zone, the process of soil salinization, and the potential long-term pollutant movement from the root zone to the groundwater zone. Groundwater inflow and outflow can also strongly affect the water quality of wetlands. As tracer profiles have become a more common way of studying recharge in deep unsaturated regimes (Allison et al. 1994; Scanlon et al. 1997; Dyck et al. 2003; Si and Kachanoski 2003), it is hoped that a detailed laboratory study will enable improved interpretation of field cores involving evaluation of the potential impact of seasonality, rain versus snowmelt contributions, and long-term effects of climate change upon groundwater recharge and discharge.

Time domain reflectrometry has been used widely in laboratory and field studies to measure the soil water content, electrical conductivity, and other soil hydraulic properties (Topp et al. 1980; Ward et al. 1994; Buttle and Leigh 1995; Wang et al. 1998; Ferre et al. 1998; Amente et al. 2000; Vogeler et al. 2000; Vogeler et al. 2001; Ritter et al. 2005; Ebrahimi-Birang et al. 2006; Hansson and Lundin 2006). Time domain reflectrometry has the advantage of allowing for continuous and simultaneous measurements of the soil water content and the electrical conductivity (Robinson et al. 2003) and usually does not require site-specific calibration (Wraith and Baker 1991). Time domain reflectrometry has been used to determine and investigate soil hydraulic properties (Ward et al. 1994; Buttle and Leigh 1995; Wang et al. 1998; Si et al. 1999; Lee et al. 2001; Noborio et al. 2006); however, no literature has been found that shows the application of TDR for estimation of soil water flux using tracer methods. The objectives of this study are to show that TDR can be used for estimation of soil water fluxes with tracer methods and to evaluate the accuracy of a tracer method, normally used for field studies, in determination of flow rate under mostly saturated conditions. The peak method was used as the tracer method and evaluated against the soil water balance method that is assumed to represent actual conditions. Although a laboratory study cannot simulate all field conditions, it does offer the advantage of isolating a few important parameters and changing a complex system into something that can be easily studied. Solute transport from the soil to the groundwater vadose zone may take decades to occur or reach equilibrium given a land use change that affects water or solute input. As a preliminary study we focused upon a sandy soil (no preferential flow paths) of homogeneous density with depth and under controlled conditions of upward and downward fluxes.

MATERIALS and METHODS

The objectives were studied in the context of a 1.05-m deep sand column with KCl used as a tracer. The column was maintained at moisture contents between saturation and field capacity. Water and the KCl solution were added using a drop sprinkler to ensure an even distribution across the surface. A column, 1.2 m in length and 0.25 m in diameter, constructed of PVC, was used in this study (Fig. 1). The sand at the upper end was exposed to allow water addition by a sprinkler system or water loss by evaporation. Fifty pairs of TDR probes were installed in the column spaced at 20 mm intervals with depth. The distance between each rod in a pair was 12 mm. The TDR probes were installed slightly offset from each other in a helical configuration to avoid the influence of each other with regard to vertical water and solute flow.

Beaver Creek sand was chosen for this study. Beaver Creek sand is found southeast of Saskatoon, SK and has been extensively researched in other column studies (Wilson 1990; Bruch 1993). A particle size analyses showed that 95.5% of the material is sand, 3.5% silt, and 1% is clay. There was evidence of salt precipitates (assumed to be primarily calcium carbonate) in the sand. This was indicated by light to moderate fizzing when a 7% HCl solution was applied. After the column was packed with disturbed sand, physical and hydraulic properties [bulk density, porosity, and saturated hydraulic conductivity (K_{sat})] were determined. Bulk density, determined from the mass of the empty column and the column filled with dry packed sand, was 1588 kg m^{-3}. Porosity was determined to be 0.37 m^3 m^{-3} using soil moisture data measured by TDR for saturated conditions. To determine the saturated hydraulic conductivity, the sand column was saturated from the bottom then a water head and downward flow were established. Estimated saturated hydraulic conductivity, using Darcy's law, was 0.973 mm min^{-1}.

To ''rain'' water, a peristaltic pump was used to supply the water to a rain cap. The rain cap was designed to add the water equally across the sand surface of the column. The rain cap was 0.25 m in diameter with water received in a top inlet and water outflow through 20 equally spaced 0.5-mm (i.d.) needles.

Evaporation was applied to cause upward movement of soil water. To control the evaporation of water, a 90-mm diameter variable-speed fan placed within a 0.25-m diameter plastic dish was located on top of the column. A tygon tube of 9.5 mm inside diameter was used for draining the water from the column. The tube outlet was maintained in a container of water located beside the sand column. This tube was used to initially saturate the sand from the bottom, and to control the water level in the sand column. Four soil temperature probes (TMC6-HD HOBO by Onset Computer, Pocasset, MA, USA), 5 mm diameter and 32 mm long, were located at depths of 15, 50, 200, and 500 mm from the soil surface. The soil temperature was recorded every hour, and average daily soil temperature at the depth 15 mm varied between 20.7 and 21.9°C during the evaporation period. Later these

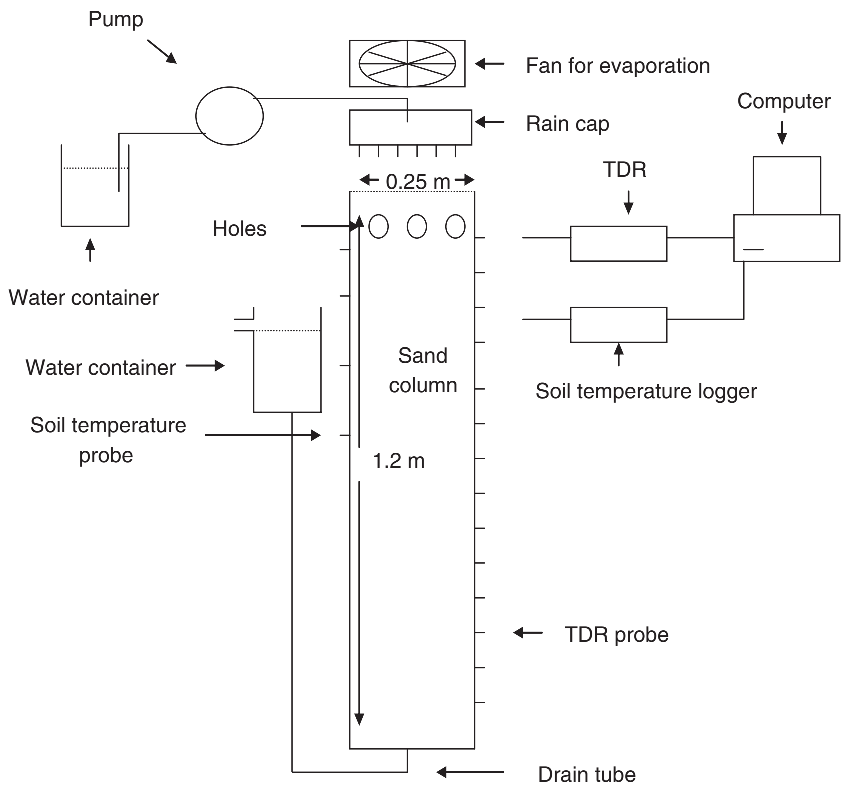

Fig. 1. The column with the rain, evaporation, and the drainage systems with the MP917 Moisture Point TDR and the temperature logger.

values were used to correct TDR measurements and to aid evaporation interpretation.

We used an MP917 Moisture Point (Environmental Sensors, Victoria, BC) TDR instrument. The length of the TDR probes was 210 mm with approximately 190 mm of the probe inserted into the sand in the column. Thirteen millimeters of each probe was left outside the column for the TDR's cable connection. A silicone sealant was used to seal the gap between each probe and the column's wall. Methods described by Ebrahimi-Birang et al. (2006) were used to calculate soil moisture and solute concentration. The value of a constant voltage approached by the TDR signal and relative to the TDR input signal can be used to obtain the bulk electrical conductivity (EC_b) of the porous media (Ebrahimi-Birang et al. 2006). Time domain reflectrometry readings were calibrated using the electrical conductivity (EC) meter and a small column (0.102 m in diameter and 0.305 m length) with the disturbed sand (Beaver Creek sand) packed dry into the column. A pair of TDR probes (210 mm in length) was inserted vertically into the sand in the column. Four KCl solutions (0.5, 1.0, 2.0, and 3.0 g L^{-1}) were prepared and added to the saturated sand column, starting with the lower concentration, to the sand column. Each time the solution was added to the column, a TDR reading (dV) was taken. Also, EC readings were taken of the drainage waters using the EC meter. A polynomial relationship was observed and yielded a coefficient of determination (R^2) of 0.95 between the TDR readings and the EC measured in drainage by the EC meter (Eq. 1). The soil moisture content was not considered in calibrating the TDR readings because the sand column, with exception of the top 75 mm of soil, was at saturated soil moisture content for the study duration.

$$EC = 0.0002dV^2 - 0.0467dV + 3.0326 \tag{1}$$

Where EC is the electrical conductivity (dS m^{-1}) and dV is the change of voltage measured by TDR (V).

The EC values measured by the EC meter were also calibrated. A solution of 2.2 g L^{-1} was prepared and diluted numerous times (22 times and the lowest concentration was 0.084 g L^{-1}). Electrical conductivity measurements were taken using the EC meter each time the solution was diluted. A linear relationship yielded a coefficient of determination (R^2) of 0.99 between the EC readings, and the concentration in solution was obtained and used to calculate the soil solute concentration (Eq. 2).

$$C = 0.6844EC - 0.0749 \tag{2}$$

Where C is the salt (KCl) concentration (g L^{-1}).

The method described by Rhoades et al. (1999) was used to transform electrical conductivity data to a reference temperature (25°C).

Two methods were used to determine upward and downward soil water fluxes: the peak migration and the water balance. The peak migration method depends on the concentration peak in the sand column. The soil water fluxes can be estimated using this method by considering the movement of the concentration peak through time and the average water content for the distance where the peak moved. The soil water flux can be calculated by the peak migration method using the following equation:

$$Q = \frac{D\theta}{T} \tag{3}$$

Where Q is the upward or downward soil water flux (m s^{-1}), D is the distance that the peak moved up or down (m), T is the time for the peak to move D (s), and θ is the average volumetric water content for D (m^3 m^{-3}).

The soil water balance technique can give an estimate about the actual upward and downward soil water fluxes under evaporative and rain conditions. This method considers the total amount of accumulated moisture in the sand column rather than the peak. This method was used to evaluate the performance of the peak migration method in determining soil water fluxes in a column study. The following equation was used to estimate the soil water fluxes in the sand column using the soil water balance method:

$$Q = P - E \pm \Delta S \tag{4}$$

Where Q is the upward or downward soil water flux (m s^{-1}), P is the rate of rain (m s^{-1}), E is the rate of evaporation (m s^{-1}), and ΔS is the rate of the change of total soil moisture (storage) in the sand column, where a negative value means a loss, and a positive value a gain (m s^{-1}).

The water table depth was maintained at 0.325 m from the sand surface to enhance the evaporation from the sand column and the upward flow. The original intent was to have a greater degree of unsaturation by having a lower water table, but when this was done the evaporation rate was far too low (less than 1 mm d^{-1}) and thus proved impracticable to test similar upward and downward flow amounts within reasonable time periods. The location and concentration of the starting peak were established by adding 20 mm (depth in soil column) of 7 dS m^{-1} KCl solution followed by 188.7 mm of distilled water to the top of the sand column using the rain system. The upward flow was caused by evaporation and downward flow was caused by adding distilled water to the sand column (rain). The changes of soil moisture, concentration, and mass of salts were considered under both upward and downward flow conditions. The total mass of dissolved salts was calculated by multiplying concentration by soil moisture volume. Thirteen days was the evaporation period and TDR readings were taken every 2 d. The rain period was 3 days and TDR readings were taken every day. Between the evaporation and the rain periods the sand column was left (covered) for 2 days to allow for equilibrium to be established. Actual average evaporation rate from the sand column calculated from change of moisture storage, and the average lost water from the water container was 11.3 mm d^{-1}, and the rain rate was assumed constant at 89.6 mm d^{-1}. The intent of these evaporation and rain rate values were not to represent typical soil water rates, but to obtain measurable rates for purposes of the study.

RESULTS and DISCUSSION

Changes of soil moisture under upward and downward flow conditions

The change of soil moisture measured by the TDR was considered under upward and downward flow conditions. The soil moisture from the water table to the column bottom varied between 0.362 and 0.378 $m^3\ m^{-3}$ before and during the evaporation period. This indicates that the porosity for this sand column was approximately 0.37 $m^3\ m^{-3}$. The saturated soil condition above the water table to a depth of about 0.1 m (Fig. 2) is due to the capillary fringe. The variation of soil moisture beneath the water table is likely due to resolution of the TDR. The soil moisture between the sand surface and the depth of 0.07 m varied with time between 0.141 and 0.283 $m^3\ m^{-3}$ under evaporative conditions (Fig. 2). Total soil moisture varied with time under evaporative conditions, decreasing from 382.4 to 365.8 mm after 10 days (Fig. 2; Table 2). Most of this decrease was due to drying of the soil between the sand surface and a depth of 0.07 m (Fig. 2).

During the 3 days of rain, the soil moisture content remained constant with time (0.36 $m^3\ m^{-3}$) beneath 0.11 m depth. The only change in moisture content was within the top 0.11 m and the change could be related to measurement resolution of the TDR. Total soil moisture varied between 369 and 371 mm of water in the sand column during the three days of rain (Table 1).

Changes of concentration and mass of dissolved salts under upward and downward flow conditions

Bulk electrical conductivity (EC_b) was determined using the TDR at a datum of 25°C. Before starting the experiment, the peak was located at a depth of 0.51 m from the sand surface and its concentration was 1.76 g L^{-1} (Fig. 3). Under evaporative conditions, there was an upward movement for the concentration peak such that after 13 days of evaporation the peak was 0.15 m from the sand surface and its concentration decreased to 1.59 g L^{-1}

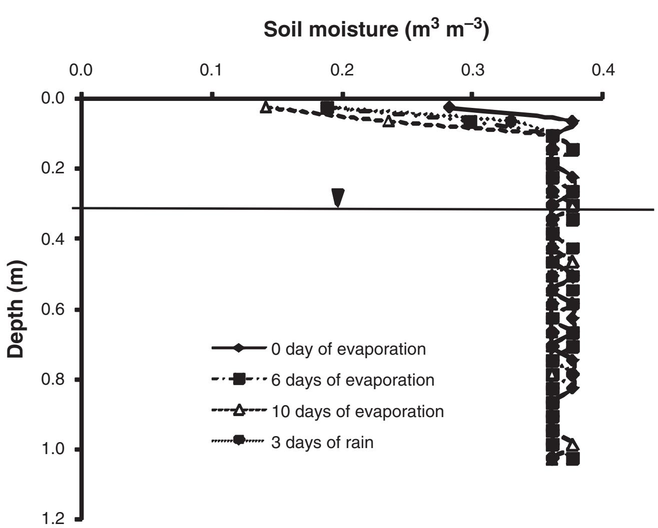

Fig. 2. The distribution of soil moisture as a function of depth under evaporative and rain conditions.

(Fig. 3). At the bottom of the sand column, the concentration decreased by 0.40 g L^{-1} after 13 days of evaporation. The decrease of the concentration in the bottom of the sand column was because of the inflow of water of low concentration from the inlet at the column bottom. In the top of the sand column (0.03 m depth), the concentration increased by 0.10 g L^{-1} after 13 days of evaporation due to concentration by evaporation.

The total mass of dissolved salts in the sand column varied between 14.2 and 15.2 g (Table 1) during the evaporation period with some of this increase coming from calcium carbonate dissolution and some due to instrumental and operator error as well as the variation of soil moisture and concentration with time.

Under rain conditions the concentration peak moved downward (Table 1; Fig. 4). The concentration near the sand surface (0.03 m depth) did not change after day 1; however, the concentration in the bottom of the sand column was increasing with time under the downward flow conditions (Fig. 4). The exfiltrate concentration was measured every day during the rain period, and it increased from 0.17 to 0.88 g L^{-1} over the 3 day period. The total mass of dissolved salts, calculated using soil moisture and concentration data, decreased with time under the downward flow conditions (Table 1). There was 4.60 g of salts lost from the system (sand column) during the rain period by drainage. The decrease in total mass of salts with time and the increase of concentration at the bottom of the sand column were because of the downward peak movement and loss of salts with drainage.

Estimation of upward and downward fluxes

Peak migration method The average estimated upward flux, over the entire 13 day evaporation period was 10.2 mm d^{-1} (moisture flux). The estimated upward flux varied with time and depth between 7.4 and 14.8 mm d^{-1} (Table 2). The increase of the estimated upward flux in the period between the sixth and tenth days of evaporation can be related to the difficulty in determination of the peak location due to the double peak on day 6 and loss of apparent peak for day 10 (Fig. 3).

The average estimated downward flux for the three day rain period was 91.9 mm d^{-1}. The downward soil water flux for the first day of the rain period was slightly high relative to the 2nd and 3rd days (Table 2), and it could be due to errors in changes of storage in the period between the end of evaporation and the beginning of downward flow.

Soil water balance method The average upward flux for the 13-day evaporation period was estimated to be 11.8 mm d^{-1} and flux varied between 9.8 and 14.3 mm d^{-1} (Table 2). These variations of upward soil water flux with time can be because of measurement or instrumental error.

The average estimated downward soil water flux for the 3 day rain period was 90.2 mm d^{-1}, varying between 88.8 and 92.3 mm d^{-1} (Table 2). The downward flux was slightly higher at the first day than that in the last 2 days, and it could be due to measurement error that cannot fully

Table 1. **The variation of peak location, rate of evaporation and rain, total soil moisture, and total mass of salts under the upward and downward flow conditions.**

Flow direction	Period of time (day)	Peak depth (m)	Evaporation and rain rates* (mm d^{-1})	Total soil moisture (mm)	Total mass of dissolved salts (g)
Upward flow	0	0.51	NA	382.4	14.2
	2	0.47	9.1	380.9	15.1
	4	0.43	10.3	NA	15.1
	6	0.39	NA	375.7	15.2
	8	0.31	11.5	370.1	15.0
	10	0.23	8.7	365.8	14.9
	13	0.15	12.2	368.8	14.5
	average	–	10.4	374.0	14.9
Downward flow	1	0.43	89.6	371.0	14.4
	2	0.67	89.6	371.0	12.7
	3	0.91	89.6	371.0	9.8
	average	–	89.6	371.0	12.3

*Evaporation rate as measured from the water container and thus does not include changes in soil moisture. NA = not available.

account for changes in storage. The change of the total soil moisture was relatively high in the first day due to the soil depth between the sand surface and the water table not being saturated.

Comparing the two methods

Both the peak migration and the soil water balance methods gave similar average upward and downward soil water fluxes (Table 2). The water balance method gave higher values of upward flux than the peak migration method during the first 6 days of evaporation. Both methods estimated a relatively high downward flux in the first day of rain then decreased in the last 2 days. Both methods showed that 1 day is needed for equilibrium conditions to be established under rain conditions. There was 14% difference between averages of upward flows estimated by the peak migration and the soil water balance methods relative to the highest average. The difference between averages of downward fluxes determined by both methods was 2% relative to the highest average. Although these measurements are within the same column, a 14% difference between the two methods may be acceptable. Saturated hydraulic conductivity, as measured from soil cores within the same field, can have coefficients of variation (CV) between 86 and 190% (Warrick and Nielsen 1980). The peak migration method gave more variable estimate (higher standard variation) than the soil water balance method did. Coefficient of variation shows that the downward flow rates are less variable than the upward flow rates, and this could be due to the magnitude of flow (Table 2).

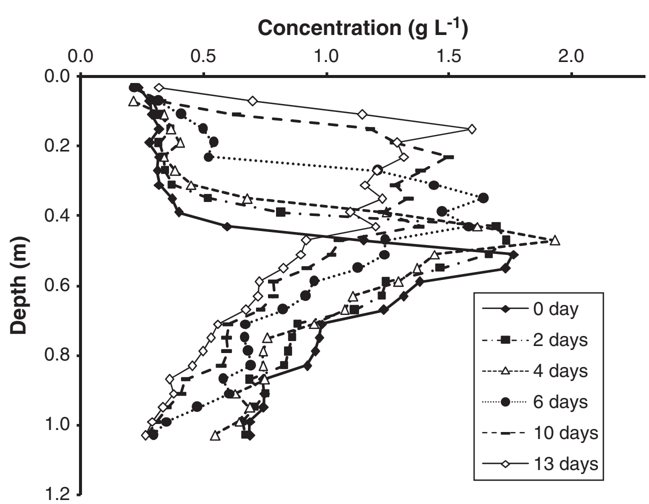

Fig. 3. The change of distribution of concentration with depth under evaporation conditions.

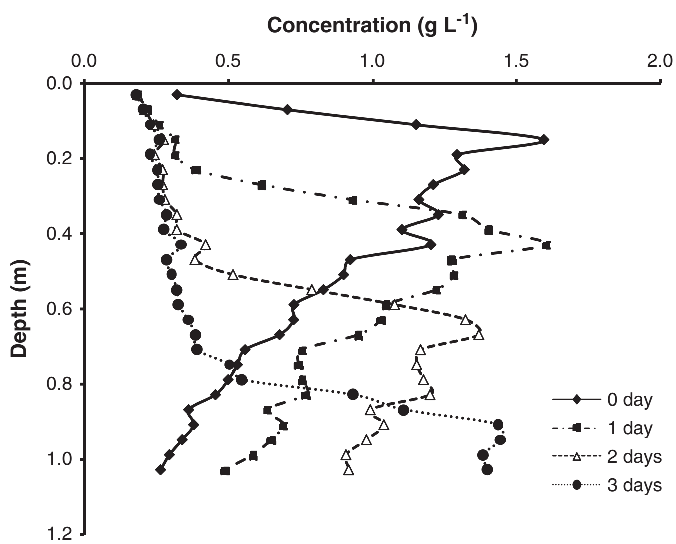

Fig. 4. The change of the concentration distribution as a function of depth and time under the downward flow conditions (rain conditions).

Table 2. Upward and downward soil water fluxes estimated using the peak migration and the soil water balance methods.

Flow direction	Period of time (day)	Estimated flux (mm d^{-1}) Peak migration	Soil water balance
Upward flow	0–2	−7.4	−9.8
	2–4	−7.4	−11.9
	4–6	−7.4	−12.5
	6–8	−14.8	−14.3
	8–10	−14.8	−10.8
	10–13	−9.6	−11.2
	average	−10.2	−11.8
	SD	3.6	1.6
	CV	35.3	13.6
Downward flow	0–1	101.4	92.3
	1–2	86.9	88.8
	2–3	87.4	89.6
	average	91.9	90.2
	SD	8.2	1.8
	CV	8.9	2.0

SD: standard deviation; CV: coefficient of variation.

The variations of estimated soil water fluxes with time and any differences between both methods can be because of measurement or instrumental error. Measurements of evaporation from the water container located beside the column might vary by ±0.5 mm. Measured soil moisture content by TDR under the water table (Fig. 2) showed that there might be a measurement error of ±0.016 m^3 m^{-3}. The peristaltic pump was tested for 24 h, and it gave a constant rate of water (rain) that it did not vary by more than 1 mm over 91 mm d^{-1}. Figure 3 shows that there might be instrumental error for measuring the EC by TDR because of the presence of double peaks in some readings (e.g., 6, 10, and 13 days of evaporation in Fig. 3). This error might be caused by TDR probes not remaining parallel or sand layers in the column caused by the backing method. Also, there might be an error of estimation of the concentration peak depth due to the 20 mm vertical distance between each two pairs of the TDR probes, therefore the actual peak might be somewhere in this depth. There might be variation of sand porosity with depth resulting in some bypass flow, air pockets, or temporal discontinuities in flow in different parts of the column. The instrumental and measurement errors and the effect of sand packing on flow may explain the variability of estimated fluxes by the peak migration method due to the effect on the peak depth estimation. The columns could not be deconstructed or tested for probe or packing error and thus a more quantitative measurement could not be done.

SUMMARY and CONCLUSIONS

The objective of this study was to show that TDR can be used for determination of soil water fluxes using tracer methods under mostly saturated conditions. A lab-column study, with a TDR system and a KCl tracer was setup to investigate this. A 1.2 m high column with sand was prepared and used in this study with a water table maintained at 0.32 m depth. The soil moisture measured by TDR under the upward and downward flow conditions was relatively constant beneath 0.07 m the sand surface. However, in the depth between the sand surface and 0.07 m, it varied with time. Under the downward flow conditions, the total mass of salts decreased with time and there was 4.75 g of salts lost from the system with drainage during the rain period.

The peak migration and the soil water balance methods gave similar average upward and downward soil water fluxes; however, the upward soil water fluxes varied with time when measured over short time periods (1 to 2 days). This result of estimated upward and downward fluxes indicates that TDR can be a useful tool for determination of soil water fluxes, and the tracer method can be recommended for determination of soil water fluxes in fields or in laboratory studies for sufficient time and depth. The TDR with the tracer method is suitable if a sufficiently long enough time and a sufficiently high depth interval are used, in comparison to the soil water balance method. An advantage with the peak migration method is that it shows greater spatial and temporal resolution than that of the soil water balance method. Moreover, it indicates that the tracer method can be successfully used in evaluating alternating direction flow under controlled laboratory conditions.

Recommendations towards improving methods and thus confirming our results would be having sand with no precipitated salts, and being able to verify packing homogeneity with depth. Recommended future work can be using other soils under unsaturated conditions. Also, the same methods can be used to evaluate the accuracy of another tracer method, such as the mass balance method, for estimating the soil water fluxes.

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