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MEASURING AND MODELING ROOT WATER UPTAKE BASED ON ^{36}CHLORIDE DISCRIMINATION IN A SILT LOAM SOIL AFFECTED BY GROUNDWATER

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Water uptake by plant roots was successfully simulated by use of a volumetric sink term S(z) added to the continuity equation for soil moisture flow, which generally requires detailed information about the root system as functions of root density, root distribution, and root length. Unfortunately, these factors are difficult to evaluate. This paper describes a simple method for the estimation of soil water extraction by roots based on root discrimination of selected solute species such as chloride. A silt loam soil planted with carrots and affected by groundwater at different depths was used for the investigation. The soil was characterized by soil matric potentials close to hydrostatic equilibrium conditions. Aeration in the root zone was impeded by high moisture content. Because chloride was strongly discriminated by the roots, root water uptake was found to be related to the increase in soil ^{36}chloride solution concentration. Consequently, the chloride in the soil water was found to be an ideal indicator of water uptake in plants. Based on this proposed approach, patterns of water extraction by carrot roots could be described using a quasi steady-state model. We also found that in the groundwater-affected silt loam soil with impeded aeration, about 80% of the water transpired was extracted from the top 5 cm of the root zone.

Upward flow from groundwater is often the main cause of a secondary salinization that threatens irrigated agriculture. This is the case in southwestern Switzerland, one of the most productive agricultural areas in the country. Here the groundwater represents an important source of water for plants. However, the plants are stressed on the fine-textured alluvial soils, not only because of solutes in the groundwater but also because of insufficient soil aeration (Schmidhalter and Oertli 1989). The salinization process is influenced variably by evaporation and transpiration. Cropping may considerably enhance upward flow and, thus, salinization. The rate of salinization increases with decreasing distance of the groundwater table from the surface in fallow soils (Schmidhalter and Oertli 1991). However, in cropped soils the sequence is reversed from the fallow soil. Biomass production and transpiration increased considerably by lowering the groundwater depth from 50 to 150 cm. Our explanation for this is that with the shallow groundwater depths, most soil pores were filled with water and root development was impeded as a result of oxygen deficiency.

The dynamics of soil salinization is strongly affected by plant water uptake in the investigated area. Therefore, an understanding of water uptake by plants is needed to identify the causes and mechanisms of soil salinization and to propose measures for improving the existing salinization problem. The objective of this work was to describe water uptake by plants in a soil with saline groundwater at 50, 100, and 150 cm depth. Soil columns planted with carrots (*Daucus carota* L.) were used for the investigation, which was carried out in a growth chamber for 121 days. Similar experiments, which are not reported here, were performed on a field site. Based on these experiments, we attempted to develop a solute transport model for nonreactive solutes to be used as an indirect method to predict the salinization process in cropped soils caused by capillary rise. This required the inclusion of a submodel describing water uptake by plant roots.

MODELING APPROACH

There are two approaches to modeling plant-root uptake of water in soils: a microscopic approach, where water flux to a single root is considered, and a macroscopic approach, where

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the entire root system as a whole is considered. The single root model approach cannot be used for the description of a complex root geometry with unknown hydraulic properties as found in carrots. Macroscopic approaches are most frequently used in water uptake models where water uptake by plant roots is represented by a volumetric sink term. The models of Molz and Remson (1970), Nimah and Hanks (1973), and Feddes et al. (1976) belong to the most common ones and are described briefly below. First, we evaluated whether one of these models could be used for the present investigation. Molz and Remson (1970) represented the sink term $S(z,\theta)$ for water uptake as

$$S(z,\theta) = \frac{TL(z)K(\theta)}{\int_0^R L(z)K(\theta)\, dz} \quad (1)$$

where T is the evapotranspiration rate per unit area of soil surface (cm hr^{-1}), $L(z)$ is the length of roots per unit volume of soil (cm cm^{-3}), $K(\theta)$ is the soil hydraulic conductivity (cm day^{-1}), z is the depth below soil surface (cm), θ is volumetric water content (cm^3 cm^{-3}), and R is depth of root zone (cm). Equation (1) was used successfully when the evapotranspiration rate T was met (Molz and Remson 1970). Such conditions are satisfied when high soil-water contents (low suctions) are maintained in the soil-root zone. Although this condition was met in the experiments reported in this paper, this approach could not be adopted because soil aeration played a crucial role in determining water uptake, and water uptake was reduced at high water contents. For a similar reason, the model by Feddes et al. (1976) could not be adopted

$$S(z) = S_{max}\ \alpha\ (h,z) \quad (2)$$

where S_{max} denotes the maximal water uptake rate and $\alpha(h,z)$ a dimensionless sink term, which varies as a function of soil water suction (h). In the very wet soil water range, soil aeration predominantly determines water uptake and cannot be described by a linearly decreasing term as a function of soil water suction. Other macroscopic approaches include that proposed by Nimah and Hanks (1973);

$$S(z,t) = \frac{[H_r - h(z,t)]RDF(z)K(\theta)}{\Delta x \Delta z} \quad (3)$$

where H_r is the root pressure head at the soil surface (cm), $h(z,t)$ is the pressure head (cm), $RDF(z)$ is the proportion of total active roots in the depth increment Δz (cm), and Δx is the distance between the plant root surface and the point in the soil where $h(z,t)$ is measured. Both $RDF(z)$, which varies with time and depth, and Δx are difficult to quantify.

Because required root parameters for the above three models are not easy to measure, an alternative method for characterizing root water uptake was sought. Certain solute species are not only nonreactive in the soil system but are also taken up by plant roots in small quantities only. Solute species also undergo changes in concentrations because of root water uptake. In this work, we investigated whether changes in solute concentration within the root zone are related to root water uptake and can thus be used to describe root activity as a function of depth and time. Therefore, one can avoid the need for detailed information on the root system, i.e., root density, root distribution, and root length, factors that are difficult to evaluate.

MATERIALS AND METHODS

Experimental

Carrots (*Daucus carota* L., var Nandor) were grown for 121 days in a growth chamber in soil columns with controlled groundwater table depths of 50, 100, and 150 cm. The treatments were replicated 3 times. The experimental design is depicted in Fig. 1; columns were made of PVC with a diameter of 0.25 m. The soil used in this investigation was illitic-chloritic silt loam (fine mixed mesic Aquic Ustifluvent) from Charrat in the canton of Wallis, Switzerland. The soil was 9.1% clay, 59.5% silt, 31.4% sand, and 0.85% organic matter. Cation exchange capacity was 4.8 cmol kg^{-1}, and the pH(H_2O) 8.2 (Schmidhalter 1986). Each soil column was homogeneously packed to a bulk density of 1.32 g cm^{-3} in the subsoil (depth >30 cm) and 1.25 g cm^{-3} in the topsoil (0–30 cm depth). The groundwater level was maintained constant by using a float (Fig. 1). Water was supplied from graduated bottles, which allowed daily determination of water consumption. Results from a previous experiment with fallow soil columns and the same experimental design were used to estimate the contribution of evaporation to total water loss (Schmidhalter and Oertli 1991). The soil columns were further equipped with tensiometers and salinity sensors (Soil Moisture Equipment

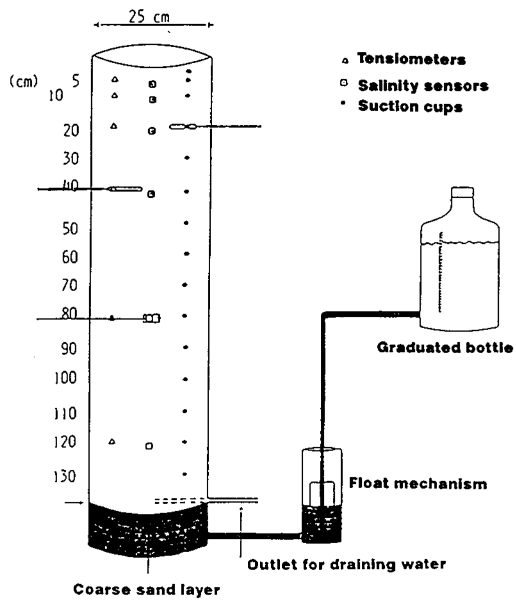


FIG. 1. Experimental set-up with location of tensiometers, salinity sensors, and suction cups.

Corporation, Santa Barbara, CA), inserted horizontally at depths of 5, 10, 20, 40, 80, and 120 cm, and soil water samplers (suction cups) inserted every 10 cm and at depths of 2 and 5 cm. Readings from the tensiometers, salinity sensors and extraction of soil solution were made twice a week. Soil water content was determined gravimetrically at the beginning and at the end of the experiment. During the experiment, soil matric potentials were converted to soil water contents based on a previously established soil moisture retention curve. The electrical conductivity of the soil solution was determined with a conductometer (ES 18, Metrohm Herisau, Switzerland). Prior to planting, the salt concentration in the soil (EC_{SE}-soil 13. 7 mS cm^{-1}) was reduced by leaching with tap water to obtain mildly saline conditions (<3.5 mS cm^{-1}). The saturated hydraulic conductivity of the soil was 8 cm day^{-1}. The dependence of the unsaturated hydraulic conductivity k(h) on the soil water suction (h) was determined for the subsoil (<30 cm depth) with the instantaneous profile method (Watson 1966) and described by $k(h) = 1208/(229 + h^{1.355})$ (Schmidhalter and Oertli 1991). The electrical conductivity of the water used was 2.6 mS cm^{-1} and its composition in meq L^{-1}: Ca, 12; Mg, 16; Na, 5; K, 0.35; SO_4, 21; Cl, 5; HCO_3, 7.35. A chloride-36 tracer as $Ca^{36}Cl_2$ (54 kBq L^{-1}) was also added to the groundwater. Chloride concentration and ^{36}Cl activity of the extracted soil solution were determined with a chloride analyzer (Corning Limited, Essex, England) and a Berthold Betaszint BF 8000 liquid scintillation counter (Berthold AG, Regensdorf, Switzerland), respectively.

At depths where the groundwater levels were established, the soil water suction did not change appreciably and remained constant (h = 0) throughout the experiment. The plants received no water from the soil surface; water supply occurred exclusively by upward flow from the groundwater. Average daily mean temperatures in the growth chamber were 11.0, 15.2, 18.2, and 18.2°C for days after planting 0–31, 32–64, 65–94, and 95–121, respectively. The daily relative humidities and photon flux densities were constant at 64% and 580 μmol m^{-2} s^{-1}, respectively. Fifty seeds were planted in each soil column. Twenty days after planting, the plants were thinned to 10 per soil column. At the end of the experiment, the storage root length was measured.

Root water uptake model

In this study, ^{36}Cl was used as a tracer for water transport and tested as an indicator of root water uptake activity. No interaction with the nonadsorbing solute chloride was expected in this soil, which is characterized by a low clay content of illitic-chloritic origin. It was further assumed, and later on tested, that compared with changes in chloride concentration in the soil solution, the amount taken up by carrot plants is only small. A model was formulated that attempted to describe ^{36}Cl distribution in the soil resulting from its discrimination by plant roots. The model assumes that the fate of Cl in solution is given by convective and dispersive processes. In order to account for the effect of plant-water uptake on chloride concentration in the soil solution, the following adjustment in C was made

$$C = \frac{C'\theta}{\theta - S(z)\Delta t} \tag{4}$$

where C' represents the chloride concentration before adjustments to changes in soil-water content (θ) attributable to water uptake, t = time (day), and $S(z)$ represents the moisture extraction rate per unit volume of soil. One-dimensional vertical transport for a nonadsorbing solute in soil can be described by

$$\frac{\partial(\theta C)}{\partial t} = \frac{\partial}{\partial z}\left[\theta\, D_{sh}\,\frac{\partial C}{\partial z}\right] - \frac{\partial(qC)}{\partial z} \tag{5}$$

where z represents depth (cm), C is solute concentration (g cm^{-3}), q is average velocity (cm day^{-1}), and D_{sh} is the hydrodynamic dispersion coefficient (cm^2 day^{-1}). Spatial-temporal water uptake was obtained by weighing the measured evapotranspiration rates with a distribution function of depth and time as described in the results section. ^{36}Chloride transport and water extraction patterns of carrot roots were described by a quasi steady-state model. Therefore, equation (5) could be simplified to

$$\frac{\partial c}{\partial t} = D_{sh}\frac{\partial^2 c}{\partial z^2} - v\frac{\partial c}{\partial z} \tag{6}$$

which is the commonly known convective-dispersive (C-D) equation for steady water flow conditions. Here v is the pore water velocity (q/θ) in cm day^{-1}. The initial conditions and boundary conditions chosen were:

$$C = 0 \quad t = 0 \quad 0 \le z \le L \tag{7}$$

$$vC_o = D_{sh}\frac{\partial C}{\partial z} + vC \quad t \le t_1 \quad z = 0 \tag{8}$$

$$\frac{\partial C}{\partial z} = 0 \quad t > 0 \quad z = L \tag{9}$$

which describe chloride transport in a soil column of length L, initially ($t = 0$) devoid of ^{36}Cl. A solution having a soluble chloride concentration C_o is applied at the bottom of the soil columns ($z = 0$) for a length of time t_1. The numerical model is valid for a semi-infinite system, and, therefore, ^{36}Cl for $z > L$ is considered as precipitated. The governing partial differential Eq. (6) was numerically solved by replacing with a second order finite difference approximation. A finite difference scheme was used for the numerical solution with an implicit-explicit scheme (Crank-Nicholson) for the diffusive term and an implicit scheme for the convective term (Selim et al. 1976). Further details of the model are given in the results section.

RESULTS AND DISCUSSION

The investigated groundwater-affected soil was characterized by high soil matric potentials that approached hydrostatic equilibrium conditions (Schmidhalter and Oertli 1991). Its capacity to transmit water at low pressure head was extraordinarily high, which is reflected in the k(h)-relation as well as in the vertical upward water fluxes for groundwater depths at 50, 100, and 150 cm (Table 1). The observed temporal and spatial changes in soil water content (Table 2) were small. This supports the assumption that water transport could be described by a quasi steady-state condition where q is constant.

For the description of ^{36}Cl transport and water uptake, the experiment was divided into 12 phases with different but constant water fluxes (Table 1). Volumetric moisture content was specified for 5-cm intervals and changed every 10 days, with a corresponding adjustment of the solute concentration. Under the experimental conditions, water uptake was not related to the soil moisture pressure head. Rather, soil aeration was crucial in determining water uptake. This is demonstrated by higher water fluxes at lower groundwater depths (Table 1). A strongly impeded aeration in the subsoil (>30 cm soil depth) (Schmidhalter and Oertli 1991) restricted rooting depths to 30 cm for the treatments with groundwater depths of 100 cm and 150 cm and to 20 cm for the treatment with a 50-cm groundwater depth. Average storage root lengths recorded at the end of the experiment (day 121) were 10.6, 13.7, and 16.7 cm in the treatments with groundwater depths of 50, 100, and 150 cm, respectively.

TABLE 1

Water fluxes in a silt loam soil planted with carrots and affected by three different groundwater depths. Fluxes (q) are indicated for different times (days) after the beginning of the experiment

Days	Groundwater depth		
	50 cm	100 cm	150 cm
		q (cm day^{-1})	
0–9	0.25	0.33	0.24
9–22	0.31	0.34	0.29
22–31	0.28	0.38	0.36
31–41	0.50	0.65	0.61
41–52	0.53	0.66	0.71
52–64	0.50	0.71	0.68
64–74	0.59	0.78	0.92
74–83	0.71	0.89	1.13
83–93	0.66	0.87	1.06
93–102	0.60	0.87	1.01
102–113	0.60	0.79	0.90
113–121	0.60	0.78	0.87

TABLE 2

Volumetric soil moisture contents (θ) in a silt loam soil with groundwater tables at 50, 100, and 150 cm measured at the beginning of the experiment (day 0) and at the termination of the experiment (day 121)

Soil depth (cm)	Groundwater depth					
	50 cm		100 cm		150 cm	
	θ_0	θ_{121}	θ_0	θ_{121}	θ_0	θ_{121}
	Volumetric water content ($cm^3\ cm^{-3}$)					
0–10	38.5	34.1	33.3	31.0	29.1	26.5
10–20	41.7	36.7	36.2	33.9	32.0	29.5
20–30	44.3	39.6	37.2	34.9	33.0	30.4
30–40	48.4	45.8	42.6	41.5	38.3	37.0
40–50	49.4	47.9	44	42.9	39.2	37.9
50–60			45.6	44.5	40.1	38.8
60–70			46.4	45.3	41.0	39.7
70–80			47.3	46.3	41.9	40.6
80–90			48.1	46.9	42.8	41.5
90–100			48.9	47.8	44.2	42.9
100–110					45.8	44.5
110–120					46.6	45.3
120–130					47.5	46.3
130–140					48.2	47.0
140–150					49.1	47.9

^{36}Chloride transport in the root zone occurred from about day 50 and day 85 for the treatments with groundwater depths at 50 and 150 cm, respectively. At these times, the roots had already extended to maximum length. The insignificant changes in the soil water pressure head (data not shown) suggest that the availability of water was unlimited. Evapotranspiration rates were controlled by the evaporative demand and varied as a function of leaf area. With minimal changes in soil moisture content occurring during the experiment, water flux by upward flow nearly equaled evapotranspiration. Decreases via evapotranspiration in soil water content accounted for less than 1.8–2.8% of the total water loss.

Except in the top 1.5 cm of the soil, no precipitation of chloride salts occurred. Because ^{36}Cl was strongly discriminated by the roots, root water uptake was related to the increase in soil ^{36}Cl solution concentration in the root zone. Uptake of ^{36}Cl by the plants was negligible and amounted to 0.6%, 0.4%, and 0.4% of the total amount present in the soil for the treatments with groundwater depths of 50, 100, and 150 cm, respectively.

The dispersion coefficient was the only unknown parameter below the root zone at the beginning of the experiment. A best fit value of 1.3 $cm^2\ day^{-1}$ was determined, which resulted in a good agreement between measured and calculated values for all treatments (e.g. Fig. 2: day 31). The sensitivity of the dispersion coefficient to the investigated water fluxes was small. Therefore, a constant value was found to be adequate for all water fluxes. This agrees with data of Moreale and van Bladel (1984). They determined a value of 1.48 $cm^2\ day^{-1}$ for a silty soil at $v = 0.66\ cm\ day^{-1}$. Nielsen and Biggar (1962) and Paetzold and Scott (1978) report values of 2.2 $cm^2\ day^{-1}$ and 0.86 $cm^2\ day^{-1}$ at $v = 1.39\ cm\ day^{-1}$ and $v = 2.27\ cm\ day^{-1}$ in a clayey loam soil and a silty loam soil, respectively.

The computing scheme was numerically stable and not affected by changes in water flux or soil water content. On average, the difference between calculations and input amounts (mass balance) was smaller than 1 to 2% and the maximum difference found was 4%. Table 3 further indicates ^{36}Cl measured between the top 2 cm of the soil and the groundwater table in percent of the input after 64, 93, and 121 days. The rest was distributed in the top 2 cm.

The main difficulty in simulating the ^{36}Cl distribution was the description of the water uptake in the root zone. Spatial-temporal water uptake was obtained by weighing the measured evapotranspiration rates with a distribution function R(z). Water uptake in the top 2 cm included loss of water by evaporation. The observed distribution could not be explained by linear (40, 30, 20, 10% rule; Rhoades and Merrill 1976), trapezoidal (Gardner 1983), or the distribution as described by van Genuchten (1987) nor by exponential distribution functions (Raats 1974). The latter, for example, underestimated the water uptake in the lower root zone.

Water uptake from the soil was successfully described by the following extraction expression of an exponential and a linear term:

$$S(z,\, t) = \frac{aT}{d}\, e^{-z/d} + (1 - a)Tb(z) \qquad (10)$$

$S(z, t)$ is the rate of moisture extraction per unit volume of soil, T the rate of evapotranspiration per unit soil surface, a indicates a fraction of the evapotranspiration rate, d is an empirical parameter chosen so that the integral of S over the

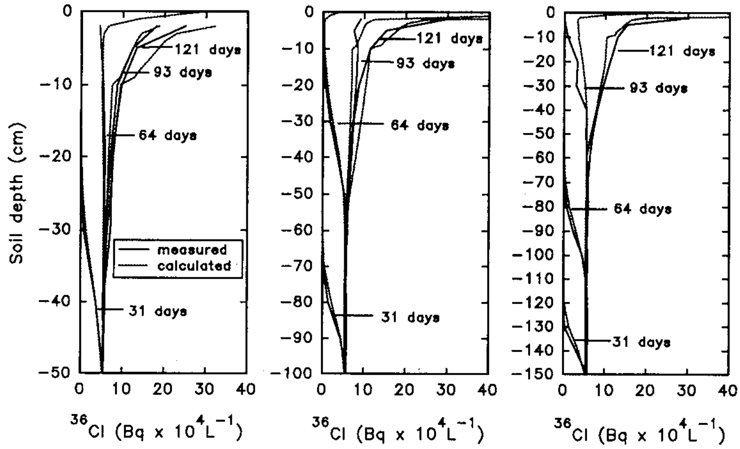


FIG. 2. Measured and predicted ^{36}Cl distribution at different times in a silt loam soil affected by groundwater and planted with carrots. Groundwater depths were 50, 100, and 150 cm. ^{36}Chloride was added as a tracer to the groundwater.

TABLE 3

Measured amounts of ^{36}Cl in groundwater affected soil columns planted with carrots. ^{36}Cl was added as a tracer to the groundwater and was measured between the top 2 centimeters of the soil and the groundwater table and is indicated in percent of input after 64, 93, and 121 days. The rest was distributed in the top 2 cm. The difference between numerical calculations and input amounts (mass balance) was smaller than 1-2%.

Groundwater depth (cm)	Days after the beginning of the experiment		
	64	93	121
	^{36}Cl in percent of the input quantity		
50	85.7	64.5	50.9
100	99.1	88.4	69.9
150	100.3	94.1	90.8

root zone is equal to T, and $b(z)$ are empirical parameters representing the nonexponentially distributed part of the evapotranspiration rate, specified for 5-cm increments. The exponential term in Eq. (10) is very similar to the distribution function as proposed by Raats (1974). The parameters d and $b(z)$ were obtained by fitting the root water uptake function to the measured chloride distribution data. Both parameters revealed to be time invariant in this experiment and were therefore assumed to be constants. The parameter d was calculated to be 5 cm (equal to 0.1666 R for the 100- and 150-cm depths and 0.25 R for the 50-cm groundwater depth where R indicates the rooting depth). Restricting the exponential uptake to the top 5 cm and using constant parameters $b(z)$ for 5-cm increments between the 0 to 30-cm (treatments with groundwater depths at 100 and 150 cm) and the 0 to 20-cm (treatment with groundwater depth at 50 cm) soil depths significantly improved the simulation results. Sixty-one percent of ^{36}Cl was thus distributed exponentially, and the remainder linearly. The distribution of b(z) in different soil depths for the three investigated treatments is indicated in Table 4.

The results show good agreement between experimental and simulated results (Fig. 2). Since the plants could discriminate chloride easily, it proved to be an ideal indicator of water uptake. A similar discrimination of chloride has been reported by Prenzel (1979), who found that only 8.6% of the chloride transported by mass flow to the roots of beech (*Fagus sylvatica* L.) trees was taken up. Under quasi steady-state conditions, the moisture extraction pattern was time invariant (Molz and Remson 1970), and a model such as Eq. (10) can yield good results. Water loss by evaporation and water uptake from the 0 to 5-cm soil depth and lower soil depths (>5 cm) are indicated in percent of the total evapotranspirational water loss in Table 5. Transpiration rates were obtained as the difference between evapotranspiration and evaporation (Schmidhalter and Oertli 1991). About 80% of the transpired water was extracted by the

TABLE 4

Distribution of the empirical constants b(z) representing the linearly distributed part of the evapotranspiration rate specified for 5-cm increments for the investigated treatments with groundwater depths in 50, 100, and 150 cm

Soil depth (cm)	Treatments (groundwater depth)		
	50 cm	100 cm	150 cm
		$b(z)$	
0–5	0.21	0.25	0.28[1]
5–10	0.03	0.04	0.04
10–15	0.04	0.02	0.04
15–20	0.04	0.03	0.03
20–25	0.04	0.03	
25–30	0.03	0.02	

[1] 0.21 is distributed linearly in the 0 to 5-cm soil depth and the rest, 0.07, is restricted to the 0–2-cm soil depth and distributed in fractions of 4/10, 3/10, 2/10, and 1/10 at 0.5-cm increments from the top downward.

TABLE 5

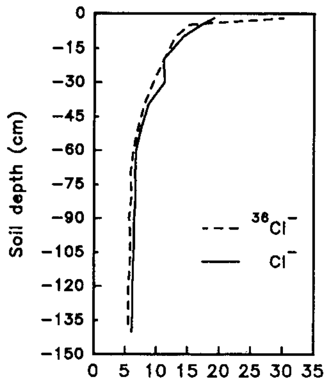
Water loss by evaporation, water uptake from 0–5-cm soil depth, and lower soil depths (>5 cm) in percent of the total evapotranspirational water loss for the investigated treatments with groundwater depths in 50, 100, and 150 cm

	Treatments (groundwater depth)		
	50 cm	100 cm	150 cm
		%	
Evaporation	35	24	17
Water uptake 0–5 cm	53	62	65
Water uptake >5 cm	11	14	18

roots from the top 5 cm. The results of this study show that in soils affected by groundwater and with impeded aeration, water uptake occurs mainly from the top soil layer.

The proposed method for describing root water uptake based on chloride discrimination appears to be ideally suited for our data sets. Conventional approaches as described previously could not be adopted. Water uptake by plants could not be inferred from changes in soil matric potential, because hydrostatic equilibrium conditions were maintained throughout the experiment. In the presence of a water table, an aeration factor is needed for modeling the sink term in addition to others that may influence physiological activity of the plant (Reicosky et al. 1972). The relationship between soil matric potential and water uptake is extremely complex in the narrow range between the so-called anaerobiosis point and soil matric potentials considered optimal for water uptake. No evidence for a simple linear relationship between high soil matric potentials, where soil aeration is critical, and root water uptake was found in these experiments. It is often tacitly assumed that root water uptake correlates with the root density distribution. There are numerous exceptions invalidating or relativizing this assumption. A comparison of the sink profile and the root density profile indicated that in the presence of a water table, a small portion of the root system can be responsible for the major portion of the water uptake (Reicosky et al. 1972). Root density distributions, whether expressed as root length, root mass, or root surface area per volume, do not necessarily correspond with root activity. At decreasing soil water availability a major portion of the root system may be found in the top soil layer without contributing much to water uptake. Root water uptake at very different soil water distributions could successfully be described by a simple model taking into account root dry mass, a time-factor indicating its activity, and the water potential gradient in the soil-plant continuum (Schmidhalter et al. 1992). Our current knowledge about temporal and spatial activities of roots with respect to water and nutrient uptake is very meager. This paper suggests a method that can furnish valuable information about the activity of roots in water uptake. Chloride or other nonreactive solutes discriminated by roots might be used as indicators and/or integrators of the activity of roots in water uptake. This method may be used not only for the description of water uptake in groundwater-affected salinized soils or soils where plants rely on water stored in the soil, but should be applicable to other situations as well. It is currently being tested whether this approach can be extended to more transient conditions characterized by intermittent rainfall and more drastic changes in the soil moisture content.

The proposed method is especially useful for areas with saline soils where upward flow from groundwater is the principal mechanism leading to secondary salinization. Salt and water transport are linked closely. The movement of water is a rough indicator for the salinization potential. This is particularly true for the present soil,



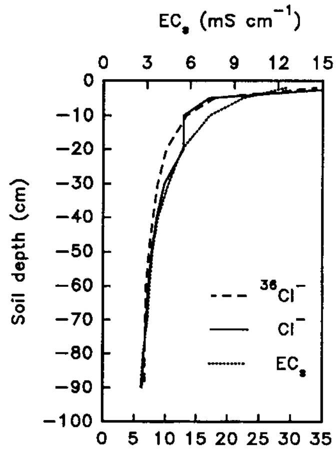


FIG. 3. (a) (left figure) ^{36}Chloride activity (Bq L^{-1}) and chloride concentration (meq L^{-1}) measured at the end of the experiment in the treatment with the groundwater depth at 150 cm. (b) (right figure) The same distributions as well as the salt concentration (EC_s, mS cm^{-1}) as measured in the treatment with the groundwater depth at 100 cm.

in which, because of a low cation exchange capacity, the transport of the bulk of salts is closely related to the movement of water and noninteracting solute species such as chloride. ^{36}Chloride distribution closely agreed with chloride (Fig. 3). A good agreement was also found between the salt concentration distribution (EC_s) and either chloride distribution, as depicted in Fig. 3, for the treatment with 100-cm groundwater depth. An even closer agreement was realized for the 50-cm groundwater depth treatment (data not shown). Initially high salt concentration for the 150-cm groundwater depth treatment accounted for a somewhat larger difference between the total salt and chloride distribution (data not shown). Gardner (1967) considered the salt profile beneath a crop irrigated with saline water. Neglecting solute diffusion or dispersion, and any extraction of salts by plant roots, he showed that the steady-state solute profile could be related to water uptake by the roots. Diffusion might be significant in moving chloride against an upward hydraulic gradient where evaporation has produced a high chloride concentration at the soil profile. Figure 3 shows that chloride is predominately accumulated in the top 2 cm of the soil with a sharp decrease in concentration in lower depths. Surface soil is especially prone to drying out by evaporation, and, thus, the backward diffusive term will be reduced as well as by precipitation of salts. An analysis of this situation has been made by Gardner (1965), assuming that the water transmitting properties of the soil, rather than evaporation, control the flux of water to the surface, allowing the steady-state condition to be applied. These studies involving nitrate found that diffusion had an influence no more than several centimeters below the evaporating surface. In previous studies using the same experimental approach and conditions, we found that convection is almost always the dominant mechanism in the movement of salts (Schmidhalter 1986), and the diffusive term is negligible. This is confirmed by similar studies of Peck et al. (1981). Hence, the chloride distribution in the root zone is essentially caused by root water uptake. The inclusion of the water uptake submodel in a solute transport model allowed us to describe the salt distribution in the investigated cropped soil in and below the root zone.

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