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Leaching and Water-Type Effects on Ground-Water Quality

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Introduction

Leaching to prevent the accumulation of soluble salts in the crop rootzone is essential for sustained irrigated agriculture. Under present irrigation practices, however, many irrigated lands are leached more than is necessary to prevent yield reductions from excess salinity buildup. Reduced leaching should not result in reduced yields in many irrigation projects in the United States (10).

When leaching fractions are reduced, the total salt load of the drainage water is reduced (3,5), although the salt concentration in the lower part of the rootzone is increased. Thus, irrigation management offers a means of controlling the salt content of drainage water under certain circumstances. In a previous study, the first author and Rhoades (7) analyzed the effects of improved irrigation efficiency on ground-water salinity for closed river basins. Depending upon water chemistry, changes in irrigation efficiency at steady-state water and salt fluxes may or may not affect downstream water quality. They classified irrigation waters into three groups: Type 1—waters initially undersaturated with $CaCO_3$; Type 2—waters initially saturated with $CaCO_3$; and Type 3—waters nearing saturation with gypsum and saturated with $CaCO_3$. Improved irrigation management resulted only in a slight reduction in downstream salinity if Type 1 water was used, in no steady-state reductions in downstream salinity with Type 2 water, and in substantial steady-state reductions in downstream salinity with Type 3 water.

The effects of irrigation management on steady-state ground-water salinities have also been analyzed (6). As with surface waters, reduced leaching, at steady state, may or may not reduce degradation of ground waters receiving irrigation

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drainage waters, depending on water type and hydrologic conditions. These evaluations were made for situations in which it was assumed that no salts, other than those present in the irrigation water or those derived from the dissolution of calcium carbonates or silicates in the soil profile, contributed to ground-water salinity. If additional readily soluble salts are present in the soil, or saline waters are present in the drainage water flow path (as is often the case), then reduced leaching will always decrease the rate of ground-water salination.

These previous studies have only considered steady-state conditions without considering travel times in either the unsaturated or saturated zones. Reduced leaching substantially increases solute residence times in and below the rootzone (4) and thus influences the rate of groundwater salination. The purpose of this study is to demonstrate the important transitory effects of changes in irrigation management on ground-water salinity, including consideration of solute transport in the unsaturated zone as well as chemical precipitation.

The effect of irrigation management on water quality can best be determined by computer simulation. Although the chemical and physical components of the model have been independently tested and verified, at the present time it is not possible to verify the simulation with data from a real ground-water basin. Accurate historical data on leaching fractions, irrigation, and groundwater quality and soil CO_2 levels are not available. It is our belief that a major reason water quality implications of irrigation management have been neglected is that its effects are observable only over long time periods (often decades or longer). Because of the considerable lag between implementation and observable results, any changes in water quality could also be caused by changes in other variables within that time frame.

Procedure

The effects of changes in irrigation management (leaching fraction), travel time, and irrigation water type on ground-water quality will be demonstrated with a hydrologically very simple transport model. Although the approach taken here does not apply to any particular ground-water basin, the calculations serve to illustrate the underlying physico-chemical principles. Changes in soil type, soil layering, and depth to water table may therefore affect the numerical results of the analysis, yet they will not alter the qualitative conclusions to be drawn from these results.

Figure 1 gives a schematic diagram of the hypothetical ground-water basin used in the analysis. A water table is assumed present at a depth of 20 m below the rootzone, while the unconfined aquifer is assumed to be bounded below by an impermeable layer at a depth of 50 m. The ground-water basin is managed to maintain the water table at a depth of 20 m, with no mass flow into or out of the basin. Because of evapotranspiration, this requires importing surface water into the basin at a rate equal to the net evapotranspiration rate, E (evapotranspiration minus rainfall; E is taken as 1.00 m/yr). Water for leaching is pumped from the underlying groundwater through a series of wells which drain water uniformly from the 30 m thick saturated zone. Thus, the rate of pumping per unit surface area, Q, is equal to the average drainage flux, D. The spatially distributed wells in the basin are assumed to be sufficiently close together so that, on the average, horizontal flow is not significant in the saturated zone. The basin is also assumed to be irrigated uniformly. Then no important areal effects are present, and the ground-water basin can be treated as a one-dimensional vertical system. Such a hydrologically simple system lends itself to analysis with relatively modest computer expenses. Although the exact simulation of an actual ground-water basin may require the use of a more complex two- or three-dimensional transport model, this will depend on the specific hydrology of that basin. The one-dimensional model will properly account for the vertical flow component, which is examined in the study.

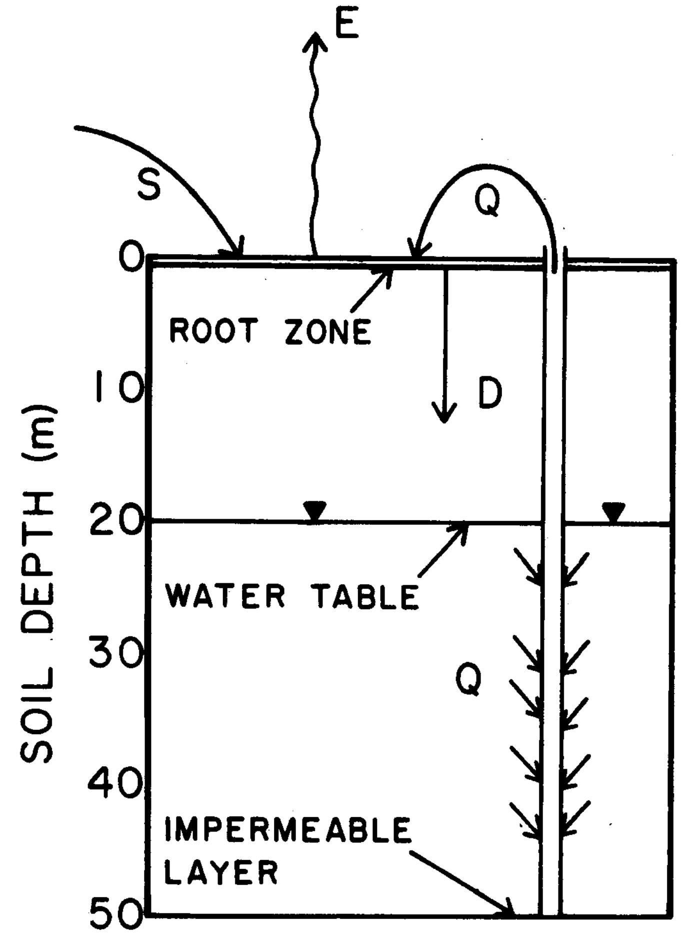

FIG. 1.—Schematic Representation of Simulated One-Dimensional Ground-Water Basin: E = Net Evapotranspiration Rate (Set at 1.00 m/yr), S = Rate at Which Water is Imported (1.00 m/yr), D = Drainage Rate below Rootzone (0.667 m/yr for L = 0.4 m/yr and 0.111 m/yr for L = 0.1), and Q = Rate of Pumping from Saturated Zone ($Q = D$)

Two different leaching fractions (L) are considered: $L = 0.4$ and $L = 0.1$. These values represent common inefficient practices (0.4) and a situation where management optimizes water efficiency (0.1). The leaching fraction, defined as that fraction of the applied water that leaches out of the rootzone, can be written for this study as

$$L = \frac{D}{S + Q} \qquad (1)$$

in which S represents the rate at which water is imported per unit surface area ($S = E$). In the present case, $D = Q$ and $S = 1.00\ m^3/m^2$ yr. One may calculate from Eq. 1 that the drainage flux for $L = 0.4$ equals 0.667 m/yr, and for $L = 0.1$ equals 0.111 m/yr. Total irrigation rates are then 1.667 m/yr for 0.4 leaching and 1.111 m/yr for 0.1 leaching.

Water and salt movement are simulated with a slightly modified version of the one-dimensional single-ion saturated-unsaturated transport model described in detail elsewhere (8). The basic transport equations however will be given here. This study analyzes the effects of different irrigation management strategies on resulting vertical salt distributions over time periods of several decades. This makes it possible to consider only steady-state water flow. Earlier studies by Wierenga (11) and Duguid and Reeves (1) suggest that to predict the quality of drainage water over long periods of time, the use of a steady state rather than a transient transport model is justified. The appropriate steady-state flow equation is

$$\frac{\partial}{\partial x}\left(K\frac{\partial h}{\partial x} - K\right) - Q_s(x) = 0 \quad \text{. . . . . (2)}$$

in which h = pressure head; K = hydraulic conductivity; and x = depth below the rootzone. The sink term $Q_s(x)$ accounts for the withdrawal of water from the saturated zone, and is given by:

$$Q_s(x) = 0, \quad 0 \le x < 20\text{ m}; \quad Q_s(x) = \frac{Q}{X_s}, \quad 20 \le x \le 50\text{ m} \quad \text{. . . . . (3)}$$

in which X_s (= 30 m) = the thickness of the saturated zone. Therefore, Q_s = zero in the unsaturated zone, and then remains constant between the water table and the bottom of the aquifer. The following relationship between the volumetric moisture content (θ) and the pressure head (h) was used for the unsaturated zone (10):

$$\theta = \theta_r + \frac{(\theta_s - \theta_r)}{[1 + (\alpha h)^n]^m} \quad \text{. . . . . (4)}$$

in which θ_r (= 0.15 m^3/m^3) = the residual soil-water content; θ_s (= 0.40 m^3/m^3) = the saturated water content; $\alpha = -0.01$; $n = 2$; and $m = 1 - 1/n$. The parameter values given here are typical for a fine sandy soil. The hydraulic conductivity of the unsaturated zone, furthermore, is described by the following predictive equation (10):

$$K = K_s\Theta^{1/2}[1 - (1 - \Theta^{1/m})^m]^2 \quad \text{. . . . . (5)}$$

in which K_s = 0.208 mm/s (or 75 cm/day) is the saturated hydraulic conductivity; and Θ = the dimensionless moisture content:

$$\Theta = \frac{\theta - \theta_r}{\theta_s - \theta_r} \quad \text{. . . . . (6)}$$

Figure 2 gives a graphical representation of Eqs. 4 and 5. It should be kept in mind that the selection of the soil-hydraulic properties will have very little effect on the results since the volumetric flux is fixed by the rate of water application and pumping rate. These rates are in turn fixed by the aquifer depth and leaching fraction. Equation 2 is solved, subject to the boundary conditions

$$\left(-K\frac{\partial h}{\partial x}+K\right)\Bigg|_{x=0}=Q \quad (7a)$$

$$\frac{\partial h}{\partial x}(X,t)=1 \quad (7b)$$

in which X (= 50 m) = the depth to the impermeable layer. Equation 7b represents a no-flow boundary condition. Note that the boundary flux at $x = 0$ is equal

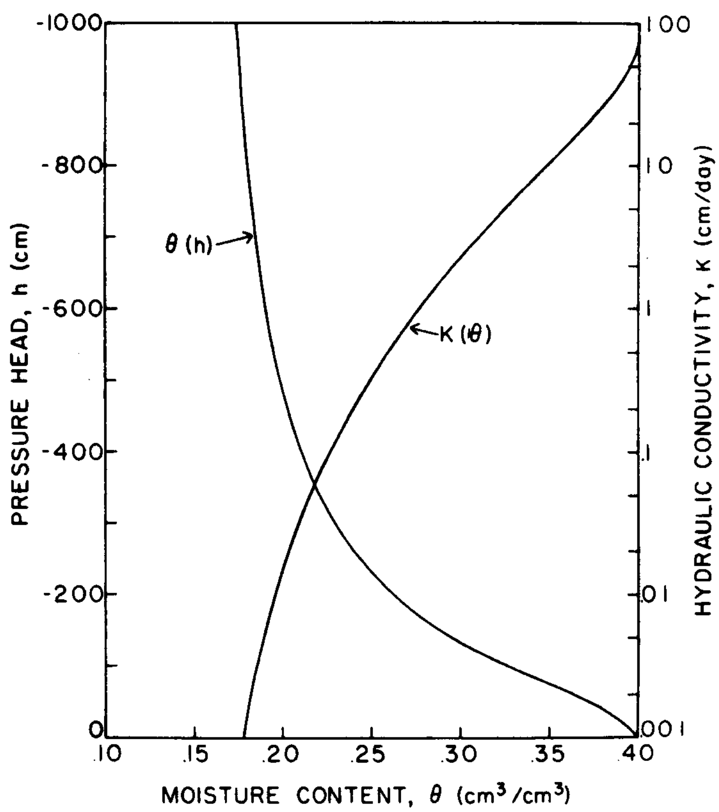

FIG. 2.—Relationship between Pressure Head (h), Volumetric Moisture Content (θ) and Hydraulic Conductivity (K) for Conditions Simulated

to the pumping rate (Q), which in turn is the same as the average drainage flux (D). The boundary at $x = 0$ corresponds to the bottom of the rootzone; the thickness of the rootzone itself is assumed to be negligible compared with the overall dimensions of the simulated system.

The governing salt-transport equation is

$$\frac{\partial \theta c}{\partial t}=\frac{\partial}{\partial x}\left(\theta\mathscr{D}\frac{\partial c}{\partial x}-qc\right)-Q_s(x)c \quad (8)$$

in which c = the total salt concentration; $\mathscr{D}$ = the dispersion coefficient; and q = the volumetric flux:

$$q = -K \frac{\partial h}{\partial x} + K \quad \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots (9)$$

The dispersion coefficient D in Eq. 8 is evaluated by

$$\mathscr{D} = a + \lambda \left| \frac{q}{\theta} \right| \quad \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots (10)$$

in which $a = 0.0116\ \mathrm{mm^2/s}$; $\lambda = 1{,}000\ \mathrm{mm}$; and where $\mathscr{D}$ and q are expressed in $\mathrm{mm^2/s}$ and mm/s, respectively. The parameter λ is often referred to as dispersivity. The 1,000-mm value chosen for λ is much higher than is generally the case (10 mm–30 mm) in most soil physics studies. The high value of λ is necessary because of the steady-state approximation of the water flow equation. This approximation results in an average downward flow of water of only a few millimeters per day. Because of temporal variations in the irrigation and evapotranspiration rates, both during the day and between days and seasons, the actual flow velocities of the water will be much higher and often change directions. These flux variations in time will result in an increased spreading of the salt, especially in the unsaturated zone and upper parts of the saturated zone. Such variations can only be included in the model by increasing the apparent dispersivity of the soil. Additional dispersion effects are caused by the fact that the basin is treated as a one-dimensional vertical system. Actual evapotranspiration and irrigation rates may not be uniform over the basin; irrigation frequency is often also not uniform while some portions of the basin may not be irrigated at all. Also, the wells may not be distributed uniformly over the basin. These nonuniformities undoubtedly will lead to some lateral flow components, especially in the saturated zone, thereby increasing the apparent dispersion of the salt along the vertical dimension. The salt concentrations, therefore, must be viewed as areal averaged quantities.

Initial and boundary conditions imposed on Eq. 8 are:

$$c(x,0) = C_o \quad \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots (11a)$$

$$\left(-\theta \mathscr{D} \frac{\partial c}{\partial x} + qc \right) \Bigg|_{x=0} = QC_d \quad \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots (11b)$$

$$\frac{\partial c}{\partial x}(X,t) = 0 \quad \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots (11c)$$

in which C_d = the concentration of the drainage water leaving the rootzone.

The subsurface material is taken to be a fine sandy soil with no ion exchange potential. Two different water types are considered in this study. In each case the initial salt concentration in the soil profile (C_o) is taken to be the same as the concentration of the imported water. In the first example, C_o is taken to be 8.52 meq/L with a composition resembling that of Colorado River or Rio Grande River water equilibrated with $CaCO_3$ at 0.03 atm CO_2 pressure. Thus, this water is initially saturated with $CaCO_3$ ("Type 2" water). Its composition in milliequivalents per liter is as follows: Ca, 6.87; Mg, 0.60; Na, 1.05; Cl, 0.21; alkalinity, 7.00; and SO_4, 1.31.

Chemical equilibrium calculations, based on a procedure similar to that of

Ref. 7, showed that the salinity of the drainage water (C_d) can be linearly related to the salinity of the irrigation water (C_i) given the foregoing. This generalization is valid only for waters where Ca $\simeq$ HCO_3, such as in the water composition selected here. In addition, when Ca is not equal to HCO_3 (in milliequivalents per liter), no generalized relationship can be proposed between the concentration of the irrigation water and the concentration of the drainage water. This is because waters of the same salinity but with different Ca concentrations can be derived by mixing various proportions of different drainage waters generated from the same original water. In that case, the flow equation and the chemical equilibria model would need to be combined into a multi-ion model and solved for each of the ions in the model. Again, this was not the case for the water chosen in this paper, since Ca $\simeq$ HCO_3.

For 0.40 leaching, the linear relationship is

$$C_d = 2.50C_i - 10.3 \quad (12)$$

For 0.10 leaching, the relationship is

$$C_d = 10C_i - 61.8 \quad (13)$$

These equations imply that the amount of $CaCO_3$ precipitated is a linear function of the concentration, with constant chemical precipitation for each leaching fraction. Although there was a slight interrelationship between chemical precipitation and concentration, it can be neglected for our present analysis.

For 0.40 leaching we have

$$C_i = 0.4C_Q + 0.6\,C_o \quad (L = 0.4) \quad (14)$$

and for 0.10 leaching

$$C_i = 0.1C_Q + 0.9\,C_o \quad (L = 0.1) \quad (15)$$

in which C_o (= 8.52 meq/L) = the concentration of the imported water; and C_Q = the average concentration of the pumped groundwater (Fig. 1):

$$C_Q = \frac{1}{30}\int_{20}^{50} c(x,t)\,dx \quad (16)$$

Substitution of Eq. 14 into Eq. 12 allows C_d to be expressed in terms of C_Q and C_o only. This expression is valid for any value of C_o chosen as long as C_d varies linearly with C_i.

This analysis assumes that no chemical reactions occur below the bottom of the rootzone. The partial pressure of CO_2 in an aquifer below an irrigated area with no recharge other than drainage water will be approximately the same as that of the bottom of the rootzone, if no $CaCO_3$ precipitation or dissolution occurs when mixing waters of different composition. Initially, calcite-saturated solutions can be supersaturated or undersaturated when mixed. This effect is explained in detail by Wigley and Plummer (12). As shown in their analysis, for fixed P_{CO_2} and temperature, this effect is largest when chemically dissimilar waters are mixed, e.g., water where Ca > HCO_3 mixed with water where Ca < HCO_3, and results in supersaturation. This effect is not important in our simulation. The other important potential cause of nonequilibrium is due to the nonlinearity of activity coefficients with ionic strength. This nonlinear effect causes an undersaturation in mixtures of solutions with different ionic strengths. The effects, however, are not substantial in the present example because subsurface mixing occurs only with waters that differ at the most one order of magnitude in concentration, not between waters with orders of magnitude differences in concentration. We assumed, therefore, that no substantial chemical precipitation or dissolution reactions take place below the rootzone in the simulation described here. The ionic strength effect is more important when groundwater is mixed directly with dilute surface water; however, this is properly accounted for by precipitation in the rootzone (Eq. 12).

The other water type considered is that of a water containing only $CaSO_4$. The ground-water concentration at $t = 0$ is taken now as $C_o = 5$ meq/L, the same as the concentration of the imported water at all times. Again, from chemical equilibrium calculations, it was found that the concentration of the drainage water leaving the soil rootzone is given by

$$C_d = 2.5\,C_i \quad (L = 0.4) \qquad (17a)$$

$$C_d = 10.0\,C_i \quad (L = 0.1) \qquad (17b)$$

in which C_i, as before, is given by Eq. 13 and Eq. 16 for 0.40 and 0.10 leaching,

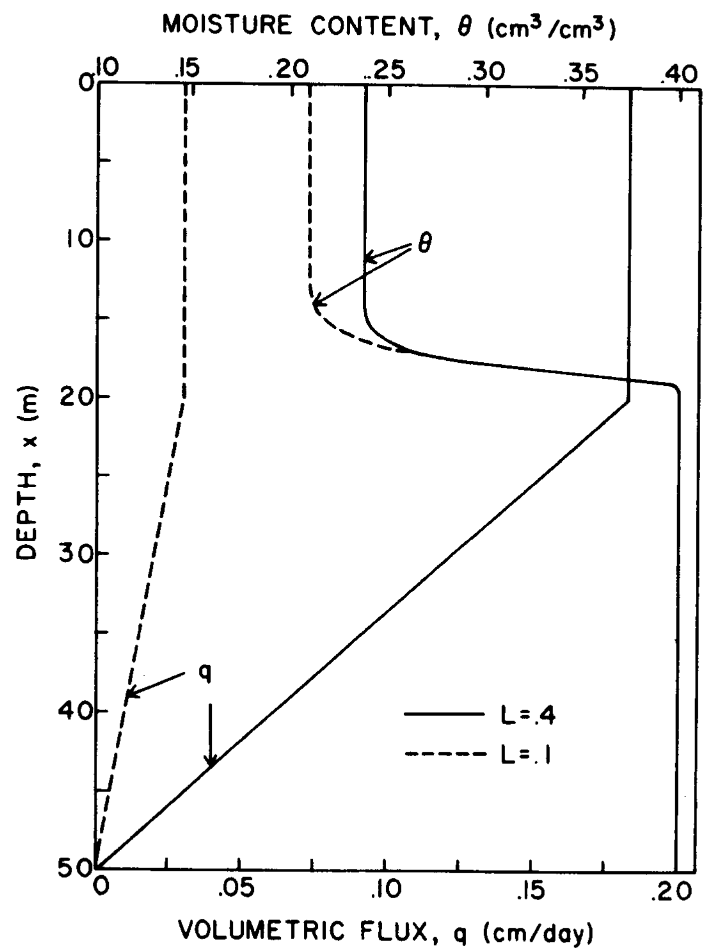

FIG. 3.—Moisture Content (θ) and Volumetric Flux (q) with Depth for 0.1 and 0.4 Leaching Fractions

respectively. The value of C_d cannot exceed 31.15 meq/L, because gypsum will precipitate at 31.15 meq/L in a pure $CaSO_4$ system (25° C). Substituting Eq. 15 into Eq. 17*b* shows that the limiting concentration occurs immediately after the start of the simulation experiment, when the leaching fraction equals 0.1.

Results

Steady-state distributions of the soil water content and the volumetric flux for both leaching fractions are shown in Fig. 3. The volumetric flux in the

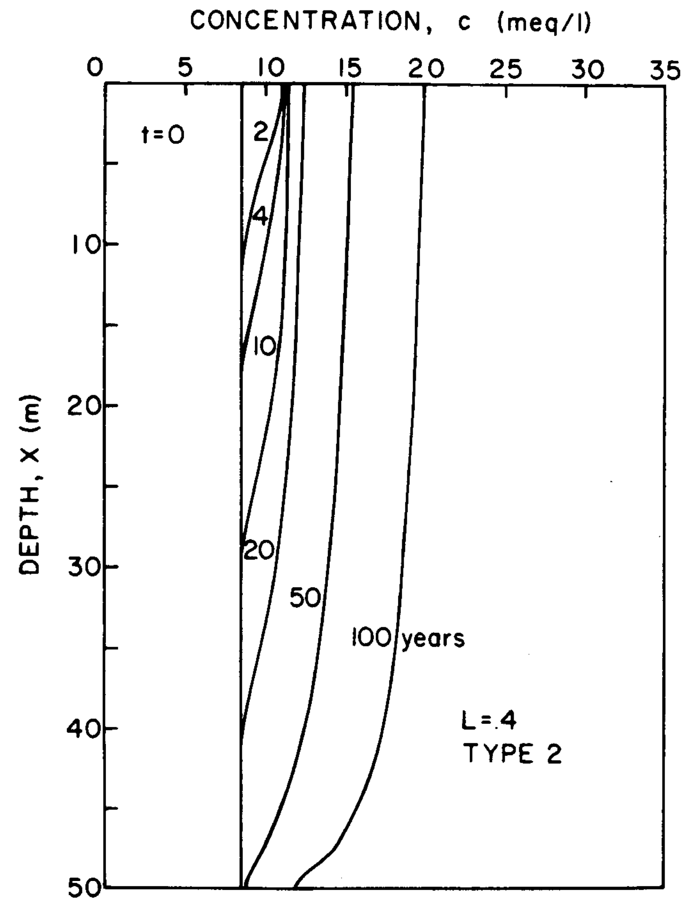

FIG. 4.—Salinity Profiles (Concentration Versus Depth) Obtained after Irrigating for 2 yr, 4 yr, 10 yr, 20 yr, 50 yr, and 100 yr with Type 2 Water ($CaCO_3$ Saturated) with Initial Solute Concentration of 8.52 meq/L and Leaching Fraction of 0.4

unsaturated zone equals the rate at which water is pumped out of the saturated zone: 0.667 m/yr for 0.40 leaching, and one-sixth of this amount or 0.111 m/yr for 0.10 leaching. As shown in Fig. 3, soil-water content in the unsaturated zone is 0.238 for 0.40 leaching and 0.210 for 0.10 leaching for the given fluxes. The volumetric flux is constant in the unsaturated zone and decreases linearly from the water table to the bottom of the aquifer where it equals zero; this results from no water being pumped from the unsaturated zone and from uniform pumping of the saturated zone.

The mean residence (or travel) time of the salt in the unsaturated zone, τ, is given by the relationship

$$\tau = \frac{\theta X_u}{q} \qquad (18)$$

in which X_u (= 20 m) = the thickness of the unsaturated zone. From Eq. 18 it follows that the mean residence time in the unsaturated zone is about 7 yr for 0.40 leaching, and 38 yr for 0.10 leaching. Thus, it will take more than five times longer for the salt to travel from the rootzone (assumed to

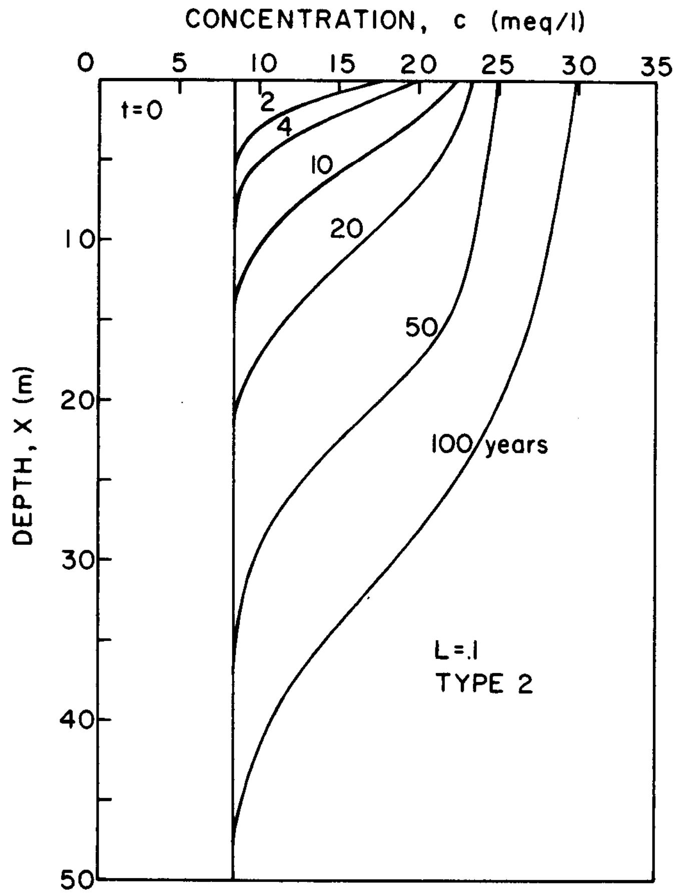

FIG. 5.—Salinity Profiles (Concentration Versus Depth) Obtained after Irrigating for 2 yr, 4 yr, 10 yr, 20 yr, 50 yr, and 100 yr with Type 2 Water ($CaCO_3$ Saturated) with Initial Solute Concentration of 8.52 meq/L and Leaching Fraction of 0.1

be of negligible thickness) to the water table for 0.10 leaching, as compared with 0.40 leaching.

Figure 4 shows salinity profiles obtained after irrigating for 2 yr, 4 yr, 10 yr, 20 yr, 50 yr, and 100 yr with Type-2 (Colorado River) water and 0.40 leaching. The concentration distributions are quite diffuse at all times, partly because of the relatively high value of the dispersivity used in the calculations, and partly because of the initially small concentration difference between groundwater and drainage water. This concentration difference will increase slowly after about 5 yr, when the concentration front reaches the water table. This occurs a few years earlier than the mean residence time (7 yr) because of the effects of dispersion on the concentration front. At that time the average concentrations in the ground water (C_Q) begin to increase, leading to an equivalent increase in the concentration of the irrigation and drainage water (Eq. 13 and Eq. 12, respectively). The concentration distributions, however, remain quite uniform versus depth, especially in the unsaturated zone (Fig. 4). Note that the concentration front reaches the bottom of the aquifer after about 50 yr.

Similar distributions for 0.10 leaching are shown in Fig. 5. While the concentration in the unsaturated zone now increases much faster than for 0.40 leaching, the concentration front itself moves much slower towards the water table. The leading edge of the front reaches the ground-water table after about 20 yr, roughly 17 yr earlier than predicted with Eq. 18 if no dispersion were present. The most striking differences between Fig. 4 and Fig. 5 are the steepness and

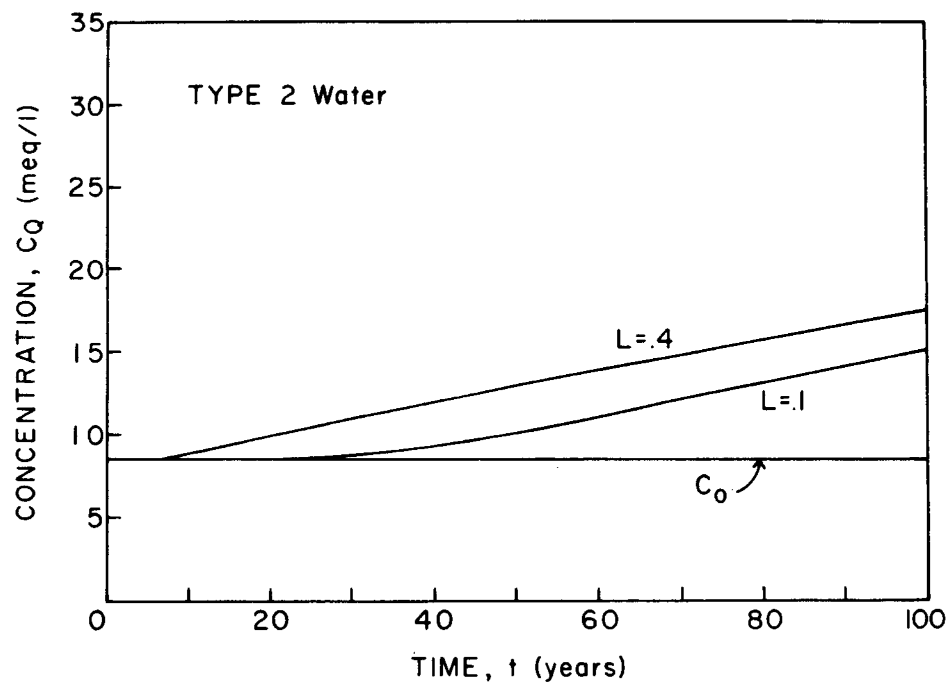

FIG. 6.—Average Ground-Water Concentration, C_Q with Time for Type 2 Water for 0.4 and 0.1 Leaching: Initial Ground-Water Concentration, C_o is 8.52 meq/L

locations of the solute fronts. Low leaching causes much more salt to be stored in the upper parts of the soil profile, leading to a slower salination of the underlying ground-water system. Especially the lower part of the saturated zone remains relatively free of salt for a much longer period of time.

Since the amount of imported water is just equal to the net evapotranspiration rate in both cases, and since this water is saturated with $CaCO_3$, the total amounts of salt precipitated in the rootzone must be the same for both leaching fractions. This can be seen by comparing Eq. 12 and Eq. 13. Precipitation accounts for a loss of 61.8 meq/L for 0.10 leaching and 10.3 meq/L for 0.40 leaching. After correcting for differences in drainage volumes (six times higher for 0.40 leaching), we obtained the same mass of salt precipitation in each case. Therefore, the total amounts of salt stored in the soil profile, including both unsaturated and saturated zones, must also be the same. This will essentially also be true for the average concentration in the soil profile, with small differences caused only by the different soil-water contents in the unsaturated zone and thus slightly different total volumes of water for the two leaching situations. The most important difference between high and low leaching is a more uneven distribution of salt for low leaching. Because less salt is stored in the unsaturated zone, the average ground-water salinity, C_Q, should be higher for 0.40 leaching than for 0.10 leaching. Figure 6 shows that this is indeed the case. The difference in average ground-water concentration for the $CaCO_3$ saturated (Type 2) water

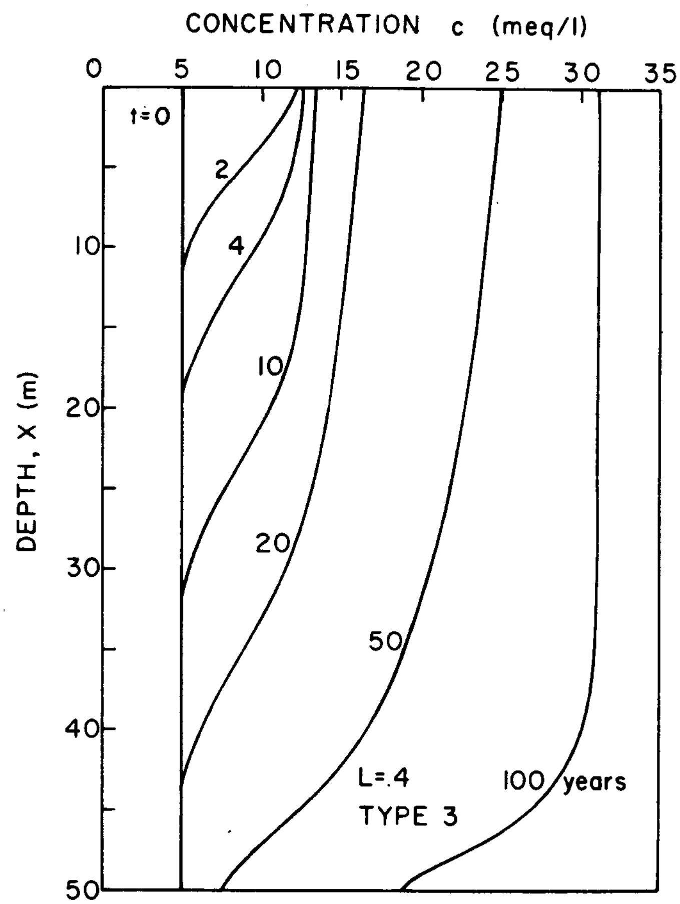

FIG. 7.—Salinity Profiles (Concentration Versus Depth) Obtained after Irrigating for 2 yr, 4 yr, 10 yr, 20 yr, 50 yr, and 100 yr with Type 3 Water ($CaSO_4$ Water) with Initial Concentration of 5.0 meq/L and Leaching Fraction of 0.4

is only about 2.5 meq/L, and remains fairly constant in time. This concentration difference, however, is expected to increase as depth to the water table increases because the mean residence time of the salt in the unsaturated zone increases (see Eq. 18).

Calculated salt distributions for Type 3 waters (those capable of precipitating gypsum) are shown in Fig. 7 and Fig. 8 for 0.40 and 0.10 leaching, respectively. The curves for 0.40 leaching (Fig. 7) are shaped similar to those for 0.40 leaching with $CaCO_3$ saturated water (Fig. 4). This was expected because the travel times in the unsaturated zone are exactly the same for both water types. The most important difference between Fig. 4 and Fig. 7 is the larger concentration gradient for most time intervals in Fig. 7. This is due to the greater amount of precipitation for the gypsum precipitating water (Fig. 7) than for the $CaCO_3$ precipitating water (Fig. 4). The concentration of a pure gypsum water, furthermore, cannot exceed the concentration of a saturated pure gypsum solution (31.15 meq/L). This upper limit on the salt concentration in the profile leads to a constant concentration between the 0 m–37 m depth after 100 yr (Fig. 7). Gypsum saturation of the drainage water was reached after 75 yr of irrigation for 0.40 leaching. This point was reached immediately for 0.10 leaching, leading

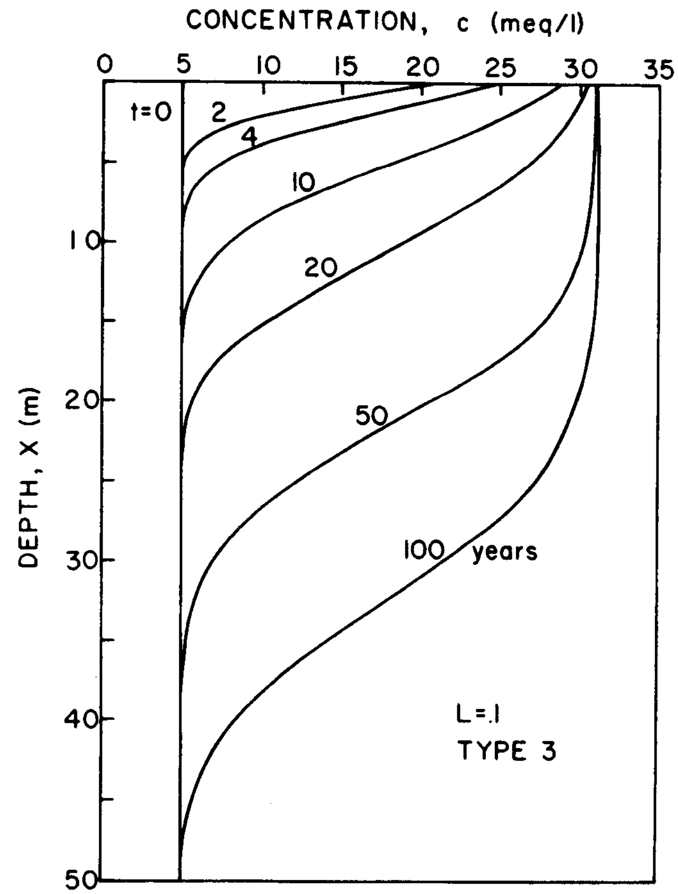

FIG. 8.—Salinity Profiles (Concentration Versus Depth) Obtained after Irrigating for 2 yr, 4 yr, 10 yr, 20 yr, 50 yr, and 100 yr with Type 3 Water ($CaSO_4$ Water) with Initial Solute Concentration of 5.0 meq/L and Leaching Fraction of 0.1

to considerably more gypsum precipitation in the rootzone. The concentration in the upper parts of the unsaturated zone does not reach saturation immediately because of the diffusion-dispersion effects on the salt distribution (Fig. 8). Increased precipitation from high to low leaching causes the differences in salinity distributions between the two leaching fractions to become more pronounced for Type 3 (gypsum precipitating) water. This is more clearly shown in Fig. 9, where the average ground-water salt concentrations are plotted versus time. The average ground-water salt concentration for 0.40 leaching increases nearly linearly from 5 meq/L (the initial concentration) after about 6 yr, to nearly

27 meq/L after 80 yr. The increase in concentration slows down after 75 yr, since it can never exceed the gypsum saturation value of 31.15 meq/L. In contrast, the average ground-water concentration for 0.10 leaching does not appreciably increase during the first 30 yr of irrigation. After 30 yr, the rate of increase in ground-water salinity is also much slower than that for 0.40 leaching. The average concentration reached 25 meq/L after 75 yr for 0.40 leaching (Fig. 9), whereas this point was reached only after 200 yr for 0.10 leaching (not shown in figure). The maximum difference in ground-water salinity between high and low leaching was reached after 90 yr (28 meq/L versus 14 meq/L). Eventually, both cases will result in gypsum-saturated ground water;

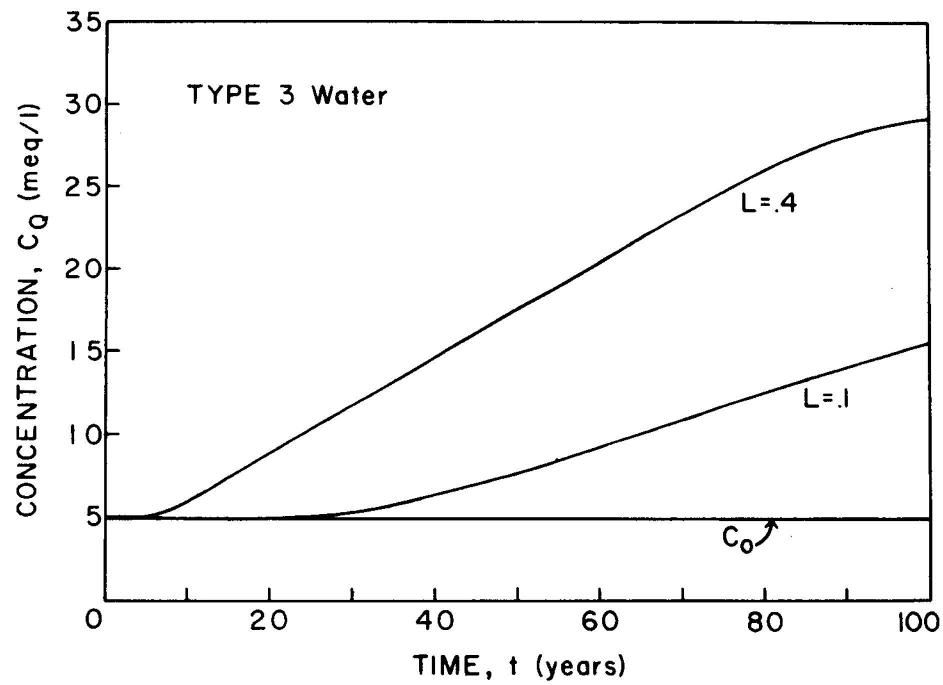

FIG. 9.—Average Ground-Water Concentration, C_Q with Time for Type 3 Water, and for 0.4 and 0.1 Leaching Fractions: Initial Ground-Water Concentration, C_o, is 5.0 meq/L

the steady-state value obviously will be reached much later for the lower leaching fractions.

Analysis

Any analysis of the type given in this paper is highly site specific and depends upon the unique conditions prevalent in each ground-water basin considered. For the closed, shallow ground-water system considered here, it is apparent that the benefit of low versus high leaching depends very much upon the type of water used for irrigation. Reduced leaching with Type 2 ($CaCO_3$ saturated) water has only a small effect on the calculated average ground-water salinity. The effect of low leaching would have been more pronounced if the depth to the ground-water table were greater than the 20 m assumed in the present calculations. The degradation of the ground water would be delayed in time and, additionally, a larger quantity of dissolved salts would then be stored in the unsaturated, rather than the saturated, zone. The main effect of increased irrigation efficiency in the case of Type 2 water is the increased travel time in the unsaturated zone. This can be of considerable benefit whenever the depth to water table is large or if the irrigation water salinity is greater than the ground-water salinity. The considerably slower travel times for low leaching means that the saline drainage front takes much longer to mix with and degrade the higher quality ground water.

From Figs. 7, 8, and 9 it is evident that low leaching can be of considerable benefit in reducing salinity when Type 3 waters are used for irrigation. The degree to which the benefit occurs is proportional to the amount of gypsum precipitated in and below the rootzone, and, therefore, depends upon the composition of the irrigation water. The greater the percentage of precipitable salts in the applied water, the greater this benefit will be. For example, a predominantly NaCl type water with lesser amounts of Ca and SO_4 will not result in much gypsum precipitation, and, thus, not reduce salinity to any great extent. The present example, that of a pure $CaSO_4$ system, represents a case of maximum benefit. The main reason for this is the increased precipitation with reduced leaching. As the proportion of soluble salts (or imbalance in Ca, HCO_3, or Ca, SO_4, or both) in the irrigation water increase, the benefits of an improved irrigation efficiency will decrease.

In any actual analysis, the effects of ion exchange should also be taken into account. Ion exchange decreases the rate at which the salt front of the adsorbing ion moves through the soil profile, and also may result in the precipitation of additional calcium carbonate and gypsum. Unfortunately, the effects of ion exchange cannot be generalized, since they depend greatly on the exact water composition, the existing exchangeable-ion composition, and the cation exchange capacity of the soil. For example, the exchange of Na or Mg for Ca does not necessarily induce precipitation. One may expect that the quantity of exchangeable Ca precipitated will be larger for reduced leaching when $HCO_3 \gg Ca$ or $SO_4 \gg Ca$. If, on the other hand, $Ca \gg HCO_3$ and $Ca \gg SO_4$, then relatively less of the exchanged calcium will precipitate. In a recent study, Jury et al. (2) showed that the exchange of Na and Mg for Ca doubled the predicted quantity of salts precipitated in the top 150 cm of a particular soil profile that was initially high in exchangeable Ca. They also showed the importance of the water-uptake distribution in the rootzone on the time necessary to achieve steady-state salt flow in the soil rootzone. These latter effects are probably less important for the type of analysis given in this paper, since the rootzone generally comprises only a small portion of the unsaturated zone. In fact, in the present analysis, the spatial effect of the rootzone was omitted altogether. If the rootzone were included in the simulation model, it would serve only to increase the travel time in the unsaturated zone and would not significantly affect the calculated salinity of the ground water.

This paper considers a ground-water basin which is a closed hydrological unit, and in which the ground-water table remained at its initial position. If the hydrological system were to be altered by not introducing water, or by significantly reducing the amount of imported water from outside the basin, then overdrafting would be simulated, provided of course that the entire basin is irrigated with the same amount of water. Then the salinity front would never reach the ground-water table, since the decline of the water table would exceed the rate of propagation of the solute front through the unsaturated zone. In this instance, salinity buildup in the ground water is not a factor. Maximum utilization of the available groundwater, however, would still require that high efficiency irrigation practices be applied.

In many cases, the ground-water basin will not be hydrologically isolated, but rather be connected hydrologically with another basin, a river, or other surface or subsurface flow system. The analysis will then become much more complex and probably not lend itself to a one-dimensional analysis of the type given in this paper. Nevertheless, we can still make some qualitative remarks, partly based upon the foregoing analysis. For example, if the applied irrigation water is imported *surface* water, then high leaching will probably result in a lower ground-water salinity because a greater volume of water and, thus, salts are being discharged from the basin.

Many basins are underlain by saline ground waters, with better quality waters being located at much greater depths. The objective in this case would be to cause as little displacement of the saline water as possible. Mixing of the saline ground water with the better quality ground water could result in a rapid degradation of a valuable water resource. The important variable now is not the drainage water composition but rather its volume. Low leaching here would be of considerable benefit.

An optimal irrigation management strategy also requires judgment as to which water resource is the most valuable. For example, in regions with inadequate surface water storage, but with a potential for good quality ground water, priority should be given to maintaining or improving the ground-water resources. If, however, the ground water is unsuitable or only marginally suitable for use, then the downstream river salinity would probably by the most important consideration in deciding which irrigation practice to adopt.

Conclusions

A one-dimensional numerical analysis of water and salt movement in a 50 m deep unsaturated-saturated soil profile showed the importance of leaching fraction and water type on ground-water salinity. For $CaCO_3$ saturated water, the effects of reduced leaching are relatively minor, especially if the depth to the water table is small. With increasing thickness of the unsaturated zone, the mean residence time of the salt in the unsaturated zone increases, leading to more salt storage in the upper part of the soil profile and less in the ground water. The average ground-water concentration in that case will be smaller with reduced leaching.

For waters approaching gypsum saturation, improved leaching efficiency will significantly improve ground-water salinity, even in shallow aquifer systems. For the case examined herein, salinity levels were up to 14 meq/L lower for 0.10 leaching as compared with those for 0.40 leaching. The long-term benefits of reduced leaching are the increased storage of precipitated gypsum and soluble salts in the unsaturated zone, and a lower salinity of the underlying groundwater system.

Appendix I.—References

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Appendix II.—Notation

The following symbols are used in this paper:

a = soil characteristic parameter used in Eq. 10, in square millimeters per second;
c = total salt concentration;
C_d = concentration of drainage water leaching rootzone, in milliequivalents per liter;
C_o = initial groundwater concentration, in milliequivalents per liter;
C_Q = average concentration of pumped groundwater, in milliequivalents per liter;
$\mathcal{D}$ = dispersion coefficient, in square millimeters per second;
D = drainage;
E = evaporation per unit surface area, in cubic meters per square meter year;
h = pressure head;
K = hydraulic conductivity, in millimeters per second;
K_s = saturated hydraulic conductivity, in millimeters per second;

L = fraction of applied irrigation water that leaches out of rootzone;
m = soil characteristics parameter used in Eq. 4;
n = soil characteristics parameter used in Eq. 4;
q = volumetric flux, in millimeters per second;
Q = rate of pumping per unit surface area, in cubic meters per square meter year;
S = rate at which water is imported into basin per unit surface area, in cubic meters per square meter year;
t = time;
x = depth below rootzone;
X = depth of impermeable layer;
X_s = thickness of saturated zone;
X_u = thickness of unsaturated zone;
α = soil characteristic parameter used in Eq. 4;
θ = volumetric moisture content;
θ_r = residual moisture content;
θ_s = saturated moisture content;
Θ = dimensionless moisture content;
λ = dispersivity, in millimeters; and
τ = mean residence or travel time.