Hydrophysical Aspects of Salinization Processes from the Groundwater

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Water as reactant, solvent and transporting agent plays a decisive role in the formation of salt affected soils. Accumulation of Na^+ ions in the liquid and/or solid phases of soils, i.e. salinization and alkalinization processes are directly or indirectly related to soil water since the soluble salts move with the water and the ion composition of soil solution governs the ion composition of the soil absorption complex.

According to 'SIGMOND [15] the evolution of salt affected soils "begins when a territory of arid or semi-arid climate becomes periodically exposed to an excess of water and the subsoil prevents the salts dissolved in the water from filtering through. Later the salt solution becomes more and more concentrated because of evaporation, and in turn the chemical and physical composition of the soil undergoes transformation in the above mentioned manner. Although conditions have varied from place to place we might name the following as periods or steps in the evolution of this soil type:

1. The formation of an impermeable subsoil layer which hinders the filtration and drainage of the water and salt solutions which accumulate on the surface.
2. Surface evaporation of the water and the concentration of the soil solution combined with an intensive mineral and organic decomposition.
3. An exchange of bases produced by the sodium salts reacting with the soil zeolites and humates which resulted in different salt affected soil types, according to the quality and quantity of the different sodium salts and the original composition of the soil material.

These periods or steps of evolution were probably repeated many times and resulted in a great variety of salt affected soil formations. In this connection, the original composition of the soil and water which accumulated on and in the soil, is of primary importance. Wherever such conditions have existed or now exist there were and will be formed sooner or later salt affected soils". Under arid or semi-arid conditions and in regions of poor natural drainage there exist possibilities and a real hazard of Na-salt accumulation in soils from the saline or brackish irrigation water, or from the shallow, saline groundwater.

Groundwater is an important factor of salt accumulation processes, because it accumulates the soluble weathering products of large areas, extensive watersheds in relatively small depressed lowlands; transports (transmits) the salt content of deep subsurface waters, geological deposits and/or of soil layers into the top horizons; impedes the leaching of salts derived from local weathering and/or irrigation water. The groundwater is the main secondary salt source first of all in low-lying, poorly or very poorly drained areas (closed basins, lowlands, low alluvial terraces and delta areas, etc.), where its horizontal flow is very slow (small slope gradient, hydraulic gradient and generally very low hydraulic conductivity) and this "stagnant" character affords the possibility of its gradual concentration; the water table is near the surface and so the capillary flow can transport relatively large quantities of water and soluble salts from the groundwater to the overlying horizons, to the active root zone.

Under such conditions salt affected soils may develop not only in arid regions but also under relatively humid climate. An appropriate example of this is the Hungarian Plain where the extension of salt affected soils exceeds 20—25% of the total arable land and there exists a potential hazard of salinization and/or alkalinization almost in the whole region [15, 19, 20, 21]. The main causes of the actual and potential salt affected soil formation processes here are: the closed character of the Carpathian Basin; thick salty Tertiary and Quaternary layers in the geological profile; stagnant, salty groundwaters with shallow (or, due to very poor natural drainage, easily and rapidly rising) water table; improper vertical and horizontal drainage and very limited possibilities of leaching [15, 19, 21]. The importance of hydrological factors (especially that of subsurface waters) in the formation of Hungarian salt affected soils were emphasized by 'SIGMOND [14—17] and eminent scientists of his school: MADOS [8], HERKE [5] and ARANY [1]. The recent scientific results obtained by SZABOLCS [18—20], SZABOLCS et al. [21], VÁRALLYAY [23] and others support this and prove the profound significance of water in salinization and alkalinization processes.

The amelioration and intensive agricultural utilization of salt affected soils (especially those with heavy texture, high swelling clay content, alkalinity and sodium saturation) are rather difficult. Their improvement necessitates expensive, complex technical measures. Therefore maximum efforts have to be exerted to prevent salinization and alkalinization, to limit their further vertical and horizontal extension and/or deterioration.

An efficient salinity and alkalinity control is possible only if the salt sources are determined; vertical and horizontal water and salt movements are exactly and quantiatively described; the influencing factors, their mechanisms and relationships are quantitatively analysed; salinization and alkalinization processes are predicted. In regions where the groundwater is the main salt source and subsurface waters play a decisive role in the development of salt affected soils, the salt accumulation processes from the groundwater have primary importance. For the exact description and prediction of these processes various aspects have to be taken into consideration:

1. Meteorological aspect: time- and territorial distribution of precipitation, potential and actual evapotranspiration, as the primary causes of suction gradient and water movement in soil.
2. Hydrogeological aspect: depth and fluctuation of water table, horizontal flow of groundwater, etc.

3. Hydrophysical aspect: time- and territorial distribution of moisture and suction profiles, water movement in the unsaturated soil layers and soil properties influencing these phenomena.
4. Chemical and physico-chemical aspect: time- and territorial distribution of concentration and chemical composition of groundwater and soil solution; the direct or indirect effects of these factors on the hydrophysical parameters; interactions between the solid and liquid phase of soils; etc.

When adequate data on these aspects are available the integrated analysis of the above mentioned factors can be carried out by multi-factor mathematical models (rationally with the use of computers) requiring well coordinated international cooperation [10, 11, 34].

In the present paper the hydrophysical aspects of salt accumulation from the groundwater are discussed. A method is introduced for the application of the unsaturated flow theory in the exact description and prediction of salinization processes from the groundwater.

Materials

Four various soils were selected for the present study from the Hungarian Plain: U — 1: Calcareous silty loam topsoil of a meadow soil.

U — 3: Highly calcareous sandy loess.

U — 8: Calcareous, heavy-textured hydromorphic fluviatile loess.

U — 9: Non-calcareous, very coarse alluvial sand.

The particle-size distribution and some other characteristic soil properties of the studied soils are given in Table 1.

Table 1.

Some characteristic data of the studied soils

Soil properties	Soil			
	U—1	U—3	U—8	U—9
pH	8.00	8.10	8.40	6.90
$CaCO_3$ (%)	21.70	30.10	13.50	0.00
EC in the saturation paste (mmhos/cm)	1.80	0.90	1.70	0.00
ESP	0.27	0.18	0.40	0.00
Bulk density (g/cm³)	1.24	1.27	1.42	1.51
Particle-size distribution				
Loss in HCl processing	25.95	33.43	14.13	0.00
1 — 0.25 mm fraction	—	4.46	0.05	50.08
0.25 — 0.05 mm fraction	6.71	22.86	9.87	49.92
0.05 — 0.01 mm fraction	25.34	24.56	24.30	—
0.01 — 0.005 mm fraction	8.37	2.48	7.80	—
0.005 — 0.001 mm fraction	10.26	2.36	16.28	—
< 0.001 mm fraction	23.37	9.85	27.57	—
Saturated hydraulic conductivity (cm/day)	3.00	32.00	1.10	2000

Theory and methods

The quantity of soluble salts entering the soil profile from the ground-water (Q_s) is determined by two factors, as follows:

$$Q_s = 10^3 \cdot C_s \cdot Q_w \tag{1}$$

where: Q_s = Quantity of soluble salts entering the soil profile from the ground-water (t/ha)

C_s = Concentration of the soil solution (g/liter or kg/m³)

Q_w = Quantity of water (solution) entering the soil profile from the groundwater (m³/ha or 0.01 cm)

C_s depends on the concentration and chemical composition of groundwater; on the solubility of the salts present; on the changes in the concentration and chemical composition of the solution during the upward capillary flow; and on the interactions between the solid and liquid phases of the soil.

Q_w depends on the direction, duration and velocity of vertical capillary flow in the unsaturated soil layers between the soil surface and the groundwater table:

$$Q_w = \frac{V \cdot T}{100} \tag{2}$$

where: Q_w = See eq. (1)

V = Velocity of the upward vertical capillary flow (cm/day)

T = Duration of vertical upward capillary flow during the reference period (day).

V — as each flux in general — is a function of the gradient of the total driving force and conductivity and may be expressed, according to the Darcy equation [7, 11, 31]:

$$V = -K \cdot \operatorname{grad} \Phi \tag{3}$$

where: V = Vector flow velocity (cm/day)

K = Hydraulic conductivity (cm/day)

$\operatorname{grad} \Phi$ = Potential gradient: the change in the total water driving force

per unit distance: $\dfrac{\Delta \Phi}{\Delta z}$ (cm water column/cm).

The total water potential (Φ) is equal with the vectorial sum of the gravitational potential (ψ_g), pressure potential (ψ_p), matric potential (ψ) and osmotic potential (ψ_s):

$$\Phi = \psi_g + \psi_p + \psi + \psi_s \tag{4}$$

therefore:

$$V = -K \cdot (\operatorname{grad} \psi_g + \operatorname{grad} \psi_p + \operatorname{grad} \psi + \operatorname{grad} \psi_s). \tag{5}$$

In the case of isothermal steady-state unsaturated flow in soils ψ_p and ψ_s can be neglected and so:

$$V = -k\lambda \cdot (\operatorname{grad} \psi_g + \operatorname{grad} \psi) \tag{6}$$

and:

$$V = -k\lambda \cdot \left(1 + \frac{\Delta\psi}{\Delta z}\right) \tag{7}$$

where: V = Velocity of vertical capillary flow (cm/day)
   k = Capillary (unsaturated) conductivity (cm/day)
   λ = dimensionless factor to account for any reduction in the cross section of water flow
   ψ = Matric potential (suction) (water column cm)

$\dfrac{\Delta\psi}{\Delta z}$ = grad ψ (cm/cm)

   1 = grad ψ_g (cm/cm)
   z = Height above a certain reference level (e.g. above the water table) (cm)

For the quantitative description of water movement in the soil profile between the soil surface and the water table the general unsaturated flow equation (eq. (7)) can be applied.

In s a t u r a t e d soil layers all of the pores are filled with water, consequently, as a first approximation (assuming that the solid phase, porosity and pore-size distribution are constants) the saturated hydraulic conductivity can be characterized by a single value (K = cm/day). Under natural conditions this is true only for soils of good structure and aggregate stability, for structureless coarse sands, etc. In other cases during the filtration of the soil solution a mechanical compaction (structure destruction, aggregate failure, clogging of macropores by particle movement, etc.) or solid-liquid phase interactions (sodium saturation, peptization and dispersion of primary soil particles, swelling, etc.) take place and the hydraulic conductivity changes, sometimes to a great extent [7, 11, 25, 31, 34].

The hydraulic conductivity of u n s a t u r a t e d soils (capillary conductivity, k = cm/day) can not be characterized by a single value but only as a function of suction or moisture content. In unsaturated layers only the pores filled with water under the given suction can conduct water, therefore capillary conductivity is always lower than saturated hydraulic conductivity ($k < K$). The water-filled pore volume at a certain suction is reflected by the water retention (pF) curves. The pF-curves, i.e. the moisture content (Θ = vol. % of moisture) — suction (cm) relationships for the studied soils are given in Fig. 1. It is obvious that with an increase in suction there is a decrease in moisture content, in water-filled pore volume and, consequently, in capillary conductivity, as well.

When data on the saturated hydraulic conductivity (Table 1) and on the moisture content — suction relationship (Fig. 1) are available the relation between capillary conductivity and suction can be determined by field and laboratory methods. WIND [32] and TALSMA [22] calculate this $k = f(\psi)$ or $k = f(\Theta)$ relationship from the field measurement of the changes in the moisture (or suction) profiles of soils, recorded by tensiometers installed at various depth. GARDNER, MILLER and ELRICK, RIJTEMA and others calculate capillary conductivity from pressure-membrane outflow data [9]. After STAPLE and LEHANE many authors computed the nonsteady-state and steady-state capil-

lary flow from the moisture profile redistribution of disturbed or undisturbed soil columns and from the equilibrium moisture profile as a result of a certain constant suction gradient, respectively. In this relation the use of γ-radiation techniques offers new possibilities [9 – 11]. RICHARDS, RICHARDS and MOORE, KLUTE and others measured capillary conductivity as a function of suction in a special apparatus [9].

Based on CHILDS – COLLIS-GEORGE [2] and WESSELING – WIT [29, 30] laboratory infiltration method, apparatuses were constructed for the determination of the $k = f(\psi)$ relationship at steady-state conditions [26]. The elaborated system can be used first of all on light-textured soils in the low suction range ($\psi = 0 - 800$ cm). During the measurements a constant capillary flux ($V =$ cm/day) is established through a vertical undisturbed soil column by precisely controlled and stabilized infiltration using various accurate feeding systems. The existing suction gradient $\left(\dfrac{\Delta\psi}{\Delta z}\right)$ in the soil column is continuously measured registrating the suction ($\psi =$ cm) at 10 cm intervals ($\Delta z = 10$ cm) by small, sintered glass tensiometers placed in the soil column. At equilibrium condition capillary conductivity is computed with equation (7) which solving for k gives:

$$k = \frac{V}{\dfrac{\Delta\psi}{\Delta z} - 1} \ . \tag{8}$$

Repeating the same procedure at various flow velocities and suctions (precisely controlled by suction plates) adequate amount of data can be obtained for the quantitative characterization of the $k - \psi$ relationship [24, 26].

The measured k values may be plotted against suction on log-log paper and the function $k = f(\psi)$ can be described mathematically by using one of the following equations:

$$k = \frac{a}{\psi^n} \ (\text{WIND [32, 33], VISSER [27, 28]}) \tag{9}$$

where: $k =$ Capillary conductivity (cm/day)
$\psi =$ Suction (water column cm)
$a =$ Experimentally determined constant (according to VISSER it is related to the conductivity of nearly saturated soil)
$n =$ Experimentally determined exponent ($1.5 - 2.0$ in clay soils, higher values in sandy soils).

$$k = \frac{a}{b + \psi^n} \ (\text{GARDNER [3, 4]}) \tag{10}$$

where: k, ψ, n: See equation (9)
$a, b =$ Experimentally determined constants
$a/b =$ Saturated conductivity ($\psi = 0$ cm) (cm/day)

$$k = \frac{K}{e^{\alpha \ (\psi - \psi_a)}} \ (\text{RIJTEMA [12, 13]}) \tag{11}$$

where: $k =$ See equation (9)
$K =$ Saturated conductivity (cm/day)
$\psi_a =$ Suction at the air entry point (water column cm)
$\alpha =$ Experimentally determined constant.

The RIJTEMA equation (eq. (11)) holds until ψ_{max} which is quite different in various soils (80 – 300 cm). In the higher suction range equations (9) and (10) are more suitable.

Experimental results

Capillary conductivity — suction relationship

Capillary conductivity as a function of suction is given for the studied soils in Fig. 2. The curves can be described by the following equations:

U–1: eq. (10), where $a = 182$; $b = 150$; $n = 1.7$; $a/b = 1.21$;

U–3: eq. (10), where $a = 3.17 \cdot 10^5$; $b = 1.00 \cdot 10^4$; $n = 2.65$; $a/b = 31.7$;

U–8: eq. (10), where $a = 120$; $b = 110$; $n = 1.6$; $a/b = 1.09$;

U–9: eq. (11), where $K = 2000$ cm/day; $\alpha = 0.19$; $\psi_a = 12$ cm.

Figure 2 shows that in the low suction range ($\psi = 1 - 10$ cm) the capillary conductivity of the U–9 coarse sand is much higher than that of U–3 sandy loess or, especially, that of finer textured U–1 silty loam and U–8 fluviatile silty loess, because the large pores of sand are almost saturated in this low suction range. As it is reflected by the pF-curves (Fig. 1), the water-filled pore volume decreases with increasing suction, particularly in coarser textured soils, where beside the large pores the volume of fine pores is small, or -- as in the U–9 coarse sand — negligible. Consequently the capillary conductivity of coarse-textured soils decreases much more rapidly with increasing suction than that of finer textured soils (Fig. 2). At $\psi = 70$ cm the unsaturated hydraulic conductivity is nearly the same for U–1, U–8 and U–9 soils ($k = 1.5 \cdot 10^{-1}$ cm/day). In the high suction range k values are higher in fine-textured soils, because a considerable part of their pores is filled with water even under higher suction (Fig. 1) and can conduct water. Over pF 3.5 – 4.2 the liquid flow is very slow.
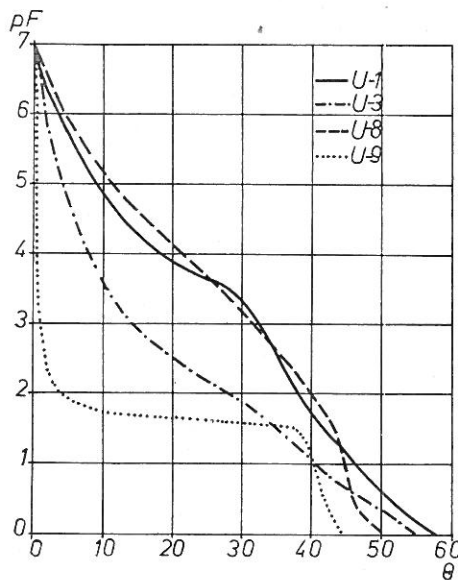

Fig. 1

Moisture retention curves of the studied soils. Vertical axis: pF; Horizontal axis: moisture content in volume percentage

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General interpretation of data

Water movement in unsaturated soil

The vertical flow of water through a homogeneous soil profile can be quantitatively characterized with the application of the general unsaturated flow equation (eq. (7)). This equation with numerical integration and solving for z gives:

$$z = \int^{\varphi} \frac{\Delta\psi}{1 + \dfrac{V}{k}} \tag{12}$$

where: z is the vertical distance above the water table
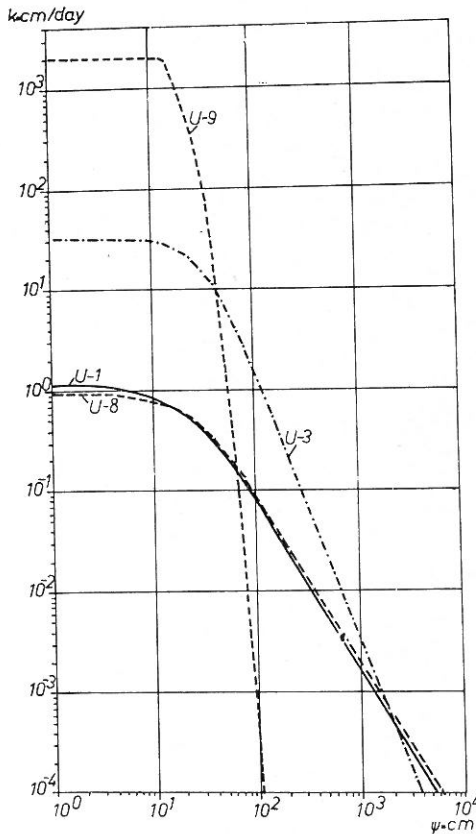
ψ, 1, V and k: See eq. (7).

Fig. 2
Relation between capillary conductivity ($k =$ cm/day) and suction ($\psi =$ cm) for the studied soils

When data on the $k = f(\psi)$ relationship are available, with the application of equation (12) a special type of set of curves can be constructed, expressing and clearly indicating the direction of vertical capillary flow and the velocity of upward capillary flow ($V =$ cm/day) as a function of the suction profile ($\psi =$ cm) and the height above the water table ($z =$ cm). These characteristic curve-sets are given for the studied soils in Fig. 3. It follows from equation (12) and can be seen from the curve-sets (Fig. 3) that as long as the suction ($\psi =$ water column centimeters) at the soil surface (or at a certain depth in the soil profile) is greater than the depth to the water table (or the height of this point above the water table: $z =$ cm), $\psi > z$, water will move upward. If $\psi = z$ (equilibrium condition) there is no capillary flow. If $\psi < z$ there is a downward capillary flow (Fig. 3).

The velocity of upward capillary flow ($V =$ cm/day) depends on the capillary conductivity at the given suction ($k =$ cm/day) and on the suction gradient $\left(\dfrac{\Delta\psi}{\Delta z} \right)$, as it follows from equation (7). Because capillary conductivity is a function of suction ($\psi =$ cm), which is a function of the vertical distance above the water table ($z =$ cm), the velocity of upward capillary flow

depends to a great extent on the height above the water table, especially
in coarse-textured soils where k decreases sharply with increasing suction
(Fig. 2). In the U—9 coarse sand V is very high near the water table but sharply
decreases with increasing z and is practically independent of ψ (Fig. 3): the
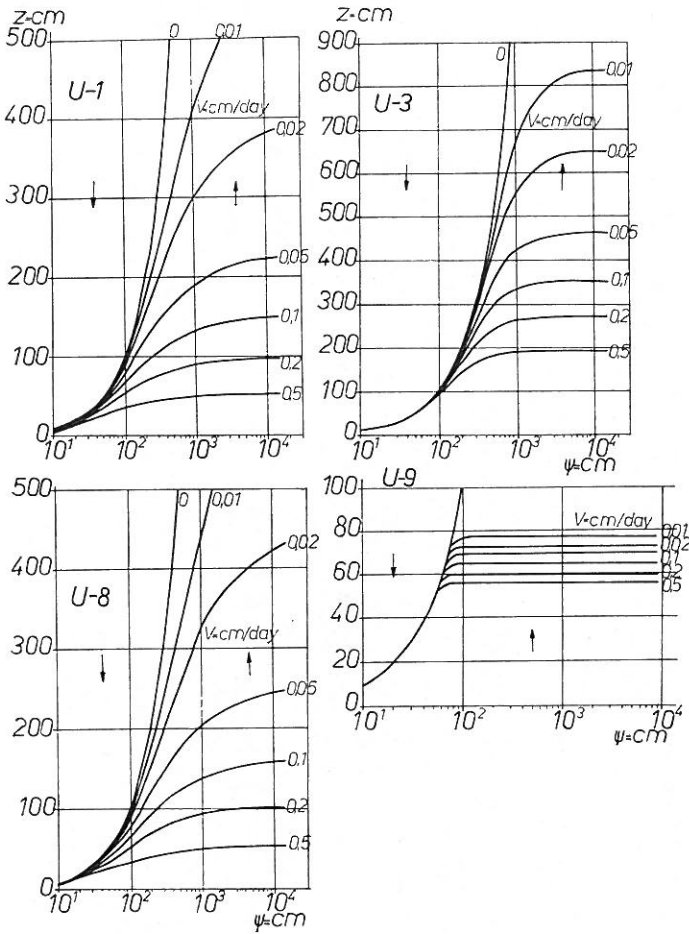

Fig. 3

The direction of vertical capillary flow and the velocity of upward capillary flow ($V =$
$=$ cm/day) as a function of the height above the water table ($z =$ cm) and suction
($\psi =$ cm) in the studied soils

capillary flow transports considerable amounts of water from a shallow ground-
water to the overlying horizons while this quantity of water is negligible in the
case of a deep water table. In the finer textured U—1 and U—8 soils V moder-
ately decreases with increasing z and depends to a high extent on the suction
profile (Fig. 3): the capillary flow transports relatively large quantities of
water (and soluble salts) even from a deep groundwater to the overlying

horizons, especially in the case of a high suction gradient. The U–3 sandy loess shows intermediate character (Fig. 3). Assuming for instance that the soil surface is dried out up to the wilting point ($\psi = 10^{4.2}$ cm) the maximum rate of upward capillary flow will be 0.5; 0.2; 0.1; 0.05; 0.02 and 0.01 cm/day in the studied soils if the depths to the water table are as follows:

U–1:	54	97	150	225	385	578 cm
U–3:	194	271	354	460	648	840 cm
U–8:	54	102	159	247	430	644 cm
U–9:	56	60	65	70	73	77 cm

If the soil surface is moist ($\psi = 200$ cm) these values are:

U–1:	42	69	94	121	150	170 cm
U–3:	144	167	181	190	195	198 cm
U–8:	40	67	93	120	152	176 cm
U–9:	56	60	65	70	73	77 cm

Water movement in layered soil profiles

For layered soils the maximum upward capillary flow velocities at a given height above the water table for various suctions can be determined only by an integrated analysis of the capillary conductivity—suction relationship (Fig. 2) or by using the characteristic set of curves (Fig. 3) for the consecutive layers.

As an example, some theoretical soil profile models were built up from the studied soils. With the use of simple computer programmes the heights above the water table ($z = $ cm) were calculated for upward capillary flow velocities of 0.5; 0.1 and 0.02 cm/day, respectively, assuming a suction at the soil surface of $10^{4.2}$ cm. The stratification of the soil profile models and the computed z-values are illustrated in Fig. 4. The first four columns represent the homogenous profiles of U–1, U–3, U–8 and U–9 soils, the 5–16. columns represent various layered profiles. The data show that:

- the upward capillary flow velocity depends largely on the capillary conductivity, thickness and sequence of the horizons in the soil profile,
- if the texture becomes heavier with depth (sand on clay) the upward capillary flux decreases, because both the clay in the low suction range (near the water table) and sand in the high suction range (far from the water table) have relatively low capillary conductivity,
- if the texture becomes coarser with depth (clay on sand: characteristic alluvial stratification) the upward capillary flux increases, because both the clay in the high suction range and sand in the low suction range have relatively high capillary conductivity.

Our data support the statement of several authors [4, 6, 12, 22, 24, 26, 27, 32–34] that the capillary flow transport considerable amounts of water (and soluble salts) even from a relatively deep groundwater to the overlying horizons first of all in moderately heavy-textured soils and in soil profiles where texture becomes heavier with depth (alluvial stratification). These conditions are favourable for the water supply of plants from the groundwater (in the case of

fresh groundwater and properly controlled drainage [12, 13, 27, 28]) but they are also favourable for salt accumulation in soils from the groundwater (in the case of saline, stagnant groundwater and improper drainage [4, 6, 22, 24]).
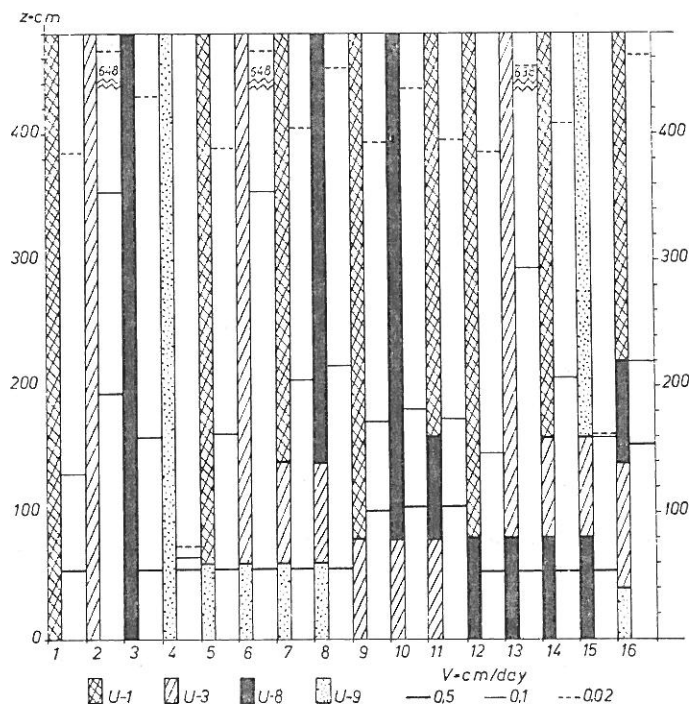

Fig. 4

Relation between the height above the water table (z = cm) and the maximum velocity of upward capillary flow (V = cm/day) in layered soil-profile models ($\psi = 10^{4.2}$ cm)

Water movement in layered soil profiles with fluctuating water table

In the case of a rising or fluctuating water table the stratification of a natural soil profile changes as a result of the change in the soil's unsaturated cross section between the soil surface and the water table. In Fig. 5, where two schematic soil profile models (N−1 and N−2) are illustrated, their stratification and the rising water table are indicated. Between the soil surface and the rising water table theoretically an infinite number of variously stratified soil profiles can be distinguished. Using the same calculation procedure as in the previous part of this paper (Fig. 4) computations were made for these profiles to determine the height above the water table where 0.5; 0.1 and 0.02 cm/day upward capillary flow velocities may exist (assuming again that on the soil surface $\psi = 10^{4.2}$ cm). With these calculations an adequate number of $z_{0.5}$ $z_{0.1}$ and $z_{0.02}$ values can be produced and from the $z_{0.5}$; $z_{0.1}$ and $z_{0.02}$ points $z_{0.5}$; $z_{0.1}$ and $z_{0.02}$ curves can be constructed. These curves are also indicated in Fig. 5. The dotted (a) lines represent the soil profile(s) illust-

rated in Fig. 4. The vertical distance between the water table and the intersection of the above mentioned z-curves and the soil surface represents the depth to the water table at which the given flow ($V = $ cm/day) may exist. These depths are for $V = 0.5$ cm/day 54 cm and 56 cm, for $V = 0.1$ cm/day 200 cm and 148 cm, for $V = 0.02$ cm/day > 250 and 310 cm in the profiles N—1 and N—2, respectively.

With the application of this step by step approach, using the above summarized relationships and the directly measured or computed data for the
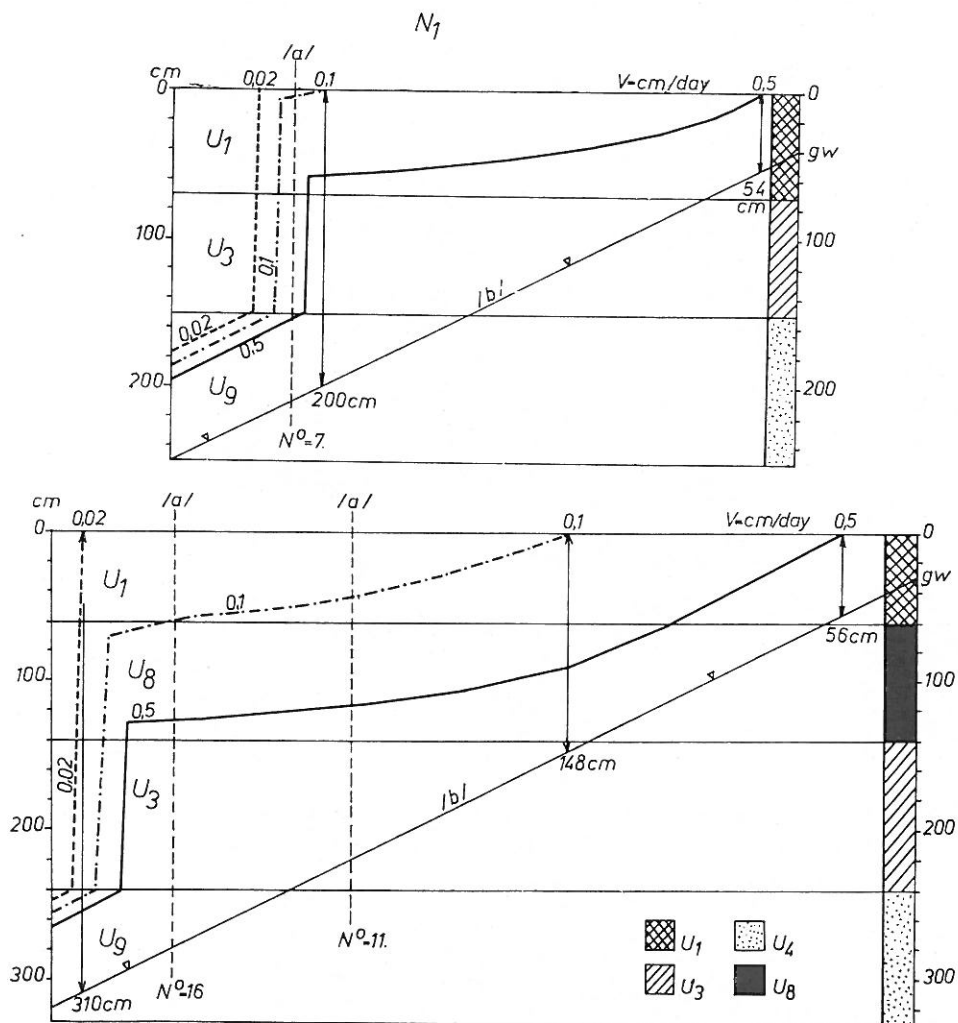

Fig. 5

Relation between the depth to the water table ($z = $ cm) and the maximum velocity of upward capillary flow in layered soil profile models in the case of rising water table ($\psi = 10^{4.2}$ cm). (a) Cross section and number of profile(s) illustrated in Fig. 4. (b) Water table

time- and territorial distribution of suction (or moisture) profiles of soils, the direction and velocity of vertical capillary flow in the unsaturated soil layers between the soil surface and the water table can be exactly determined and interpreted for soil profiles, mapping units or territories. The suction (or moisture) profile can be directly measured by tensiometers, gypsum blocks, nylon elements, neutron scattering or γ-radiation techniques. If direct measurements are lacking the suction (or moisture) profile dynamics can be computed as a function of the depth to the water table and the resultant of water movement through the soil surface (precipitation + infiltrated surface waters + irrigation water − evapotranspiration), which factors can be measured or estimated in a relatively simple way. For the time- and work-consuming calculations relatively simple computer programmes can be written and properly used. By using predicted values instead of measured ones (from the meteorological and geohydrological prognosis and irrigation plans) a more or less accurate p r o g n o s i s can be given for the water movement in layered unsaturated soil profiles with fluctuating water table and the quantity of water entering the soil profile from the groundwater can be predicted, as well.

Interpretation of data for salinity prognosis

Salt accumulation from the groundwater takes place if a larger quantity of water soluble salts enters the soil profile from the groundwater during a certain period than can be leached out from it into the groundwater and removed by horizontal groundwater flow from the area.

This situation may occur in the following circumstances:

a) The resultant of the water movement through the soil profile is upward (upward flow > downward flow);

b) The resultant of the water movement through the soil profile is downward but the excess quantity of water filtrating downward is not enough for the leaching of accumulated salts, because the solubility of salts entering the profile from the groundwater changes;

c) More or less irreversible changes take place in the soil's solid phase (sodium saturation, etc.) under the effect of filtrating solutions and so the downward movement of water (and salts) is limited to a certain extent.

Phenomenon *a)* is characteristic of the neutral salt salinization of coarse-textured soils; phenomenon *c)* of the soda salinization of heavy-textured swelling soils.

SZABOLCS, DARAB and VÁRALLYAY [21] elaborated a system for the description and prediction of salinization and alkalinization processes and presented a rather simple and practical calculation method for the determination of the "critical depth" to the water table. The prognosis system has been successfully used in planning irrigation systems and has afforded practical possibilities for the prevention of secondary salinization and alkalinization.

With the application of the unsaturated flow theory for the quantitative description of salt accumulation processes from the groundwater, the hydrophysical aspect of this prognosis system can be approached more exactly, as it was done by TALSMA [22], NERPIN [10] and others [6, 11].

Disregarding the chemical and physico-chemical aspects, as a first approximation, salt movement in the unsaturated soil profile and salt accumulation in the soil from the groundwater can be described and/or predicted

on the basis of water movement (discussed above) and the measured, calculated or predicted concentration and chemical composition of the permeating fluid:

1. On the basis of exact, long-term salt balances (calculated on the basis of field measurements or laboratory model experiments) the maximum quantity of salts, which may enter the soil profile from the groundwater without causing salt accumulation, has to be determined. This quantity ($Q_{sm} = t$/ha) can be defined as *"critical quantity"* (salt regime coefficient, natural leaching potential of the soils). If a higher amount of salts enters the profile the salt balance will be positive (salt accumulation). If an equal or lower quantity of water soluble salts enters the profile, the salt balance will be stable (no changes in the salt content) or negative (desalinization).

2. Knowing the Q_{sm}, the average concentration of the upward permeating soil solution (C_s = g/lit or kg/m³) and the duration of vertical upward capillary flow during the reference period (T = day), the maximum permissible velocity of vertical upward capillary flow from the groundwater, which may be defined as *"critical upward capillary flow velocity"* (V_m = cm/day), can be computed with the application of equations (1) and (2). It gives:

$$V_m = \frac{10 \cdot Q_{sm}}{C_s \cdot T} \tag{13}$$

T can be computed from the time-distribution of suction profiles and groundwater levels as it was discussed in the previous section of the present paper.

3. Knowing V_m and in possession of data on the time-distribution of suction profiles, the minimum permissible depth to the water table which prevents a higher flux as the "critical" one can be either calculated with equation (12) or simply read from the curve-sets similar to those shown in Fig. 3. This depth can be defined as *"critical depth"*. Assuming for instance that the critical flow velocity for the soils studied is 0.05 cm/day and the soil surface is dried out up to the wilting point ($\psi = 10^{4.2}$ cm) the critical depth of water table is 70 cm for the U—9 very coarse sand, 460 cm for the U—3 sandy loess, 247 cm for the U—8 silty loess and 225 cm for the U—1 silty loam topsoil (Fig. 3). The calculations for layered soils and for layered soils with rising or fluctuating water table are similar to that which was followed when constructing Fig. 4 and 5, respectively.

4. The critical depth in this interpretation is not a single value but a function of the suction gradient, so it has a characteristic seasonal variation, consequently it is better to define it as a *"critical groundwater regime"*. In dry, warm seasons the suction gradient is generally higher than in cold, humid periods: the same critical flow velocity (V_m) means deeper minimum permissible depth to the water table as it follows from the curve-sets in Fig. 3. In this respect the selection of the observation period has a great importance because the calculations, made for longer periods (i.e. one or more years), may give false results and lead to false conclusions. It is possible for instance that the annual salt balance is stable but salinization during the dry season results in more or less irreversible changes in the soil properties, consequently in the water and salt regime, as well. The use of computer techniques broadens the perspectives of these calculations because thus short reference periods (one month, week or day) can be selected and so the calculation gives more accurate results.

5. The other preventive measures (keeping the soil surface moist, improving the infiltration rate and internal drainage of soils, etc.) can be evaluated using the same procedure. For instance keeping the soil suction at field capacity ($\psi = 300$ cm) on the surface of the silty loam topsoil (U – 1) the critical depth to the water table for $V_m = 0.05$ cm/day is 143 cm. On this basis the optimum combination of preventive and ameliorative measures can be selected.

The presented method with the application of the unsaturated flow theory offers possibilities for a better approach to the hydrophysical aspects of the exact description and prediction of the salt accumulation processes from the groundwater, thus providing a good basis for an accurate and quantitative integrated analysis of the salinization and/or alkalinization of soils.

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