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Estimating water movement from salt movement

Laura Constantinescu¹, Adia Grozav¹, Gh. Rogobete¹

Abstract: Three mechanisms operate to transport soluble salts in soils and aquifers: advection, diffusion and dispersion. Solute can interact with soil surfaces and the process is known as retardation. Advection often dominates the solute transport process. The rise of water in the soil from a free water surface has been termed capillary rise. The soil types which has been studied are Fluvisols, Chernozems, Phaeozems and Gleysols developed in accumulating alluvial deposits. The environmental conditions during the process of sedimentation invariably results in the stratified parent material of alluvial soils. Therefore, stratification is the major characteristics used to distinguish these soils. All types of soils have a high groundwater table and the water can be strong mineralized with soluble salts, predominantly sulphatic in the low plain Timiș-Bega. The water movement may be calculated from soil salinity with the help of moisture content from capillary fringe and the soluble salt content during a dry period. The total evaporation from the soil profile studied is 157 mm.

Keywords: soil, water, soluble salt, capillary, movement.

1. INTRODUCTION

The movement of salts in soils and groundwater aquifers (solute transport) plays a central role in global biogeochemical cycling.

Flow through the vadose zone requires study of saturated and unsaturated flow through variably saturated media. Both saturated and unsaturated flow problems may be steady-state or transient.

Pores may contain water, soil gases, and/or introduced materials such as nonaqueous-phase liquids. Three mechanisms operate to transport solutes in soils and aquifers.

“Advection” refers to the movement of chemicals with soil fluids and groundwater.

Advection often dominates the solute transport process. A fluid flowing through a porous medium at a flux q (volume of fluid per unit area of soil per unit time) with a solute concentration C (mass of solute per unit volume of fluid) carries with it a solute flux $J_a = qC$ (mass of solute per unit area of soil per unit time).

The mean velocity of the flow v (and the solute, assuming no losses, production, or interaction with the solid phase) can be computed from knowledge of the flux q and the volumetric fluid content θ via $v = q/\theta$. (Sukop M.C., 2005).

Under steady flow, $x = v \cdot t$ (where v is the magnitude of v), and the distance x traveled by an average solute molecule in time t , or time t needed for the solute to traverse a given distance x , can be calculated.

The advective solute, flux J_c can also be written in terms of the mean velocity, the concentration, and the fluid content: $J_c = v \cdot \theta \cdot C$.

In some instances, advection dominates the solute transport process and may be the only process that need be considered.

Diffusion is the omnidirectional spreading of a chemical that results from thermal (Brownian) motion of the fluid and solute molecules.

Dispersion results primarily from variations in fluid velocity that occurs at all scales.

Dispersion is anisotropic and enhanced in the overall direction of flow. Solutes can interact strongly with soil surfaces by physical or chemical adsorption and their transport can be appreciably slowed in a process known as retardation.

The rise of water in the soil from a free-water surface has been termed capillary rise. This term derives from the “capillary model”, which regards the soil as analogous to a bundle of capillary tubes predominantly wide in the case of a sandy soil. Accordingly, the equation relating the equilibrium height of capillary rise h_c to the radii of the pores is:

$$h_c = \frac{2 \gamma \cdot \cos \alpha}{r \cdot \rho_w \cdot g} \quad (1)$$

where γ is the surface tension, r the capillary radius, ρ_w the water density, g the gravitational acceleration and α the wetting angle, normally taken as zero.

This equation predicts that water will rise higher (though less rapidly) in a clay than in a sand. However, soil pores are not capillary uniform or constant radii, hence the height of capillary rise will differ in different pores (Hillel D., 1971).

Above the water table, matric suction will generally increase with height and the number of water-filled pores will decrease in each soil.

Generally, for a given soil, and given boundary and initial conditions, the water flow is governed by the soil hydraulic properties.

¹ Faculty of Civil Engineering, Department of Hydrotechnical Engineering, George Enescu Street No. 1/A, 300022, Timișoara, E-mail: adiagroav@yahoo.com

Thus, the effect of the soil-matrix-soil-solution interactions is to reduce the water flow relative to that of a “stable” reference soil.

On the other hand, the solute transport is governed by the water flow as well as by cation exchange and anion exclusion processes. The retardation effect caused by the reductions in the soil hydraulic conductivity and, concurrently, in water velocity, increases with increasing SAR and decreasing C and h, especially in relatively fine-textured soils (Rogobete, 1994).

2. MATERIAL AND METHODS

As part of the complex soil survey effectuated in the subsidence plain Timiș-Bega in the year 2008 have had achieved a soil map at large scale.

A great area is occupied by Fluvisols, Chernozems and Gleysols with intense phenomena of salinization and gleysation.

The ground water level is at 1.5-2.0 m depth with soluble salts and because of capillary rise there is solute transport in the topsoil. Soil samples have been taken of each soil layer of 30 cm and were made soil analysis, at the beginning and end of a dry period, respectively in May and in September.

The most important analysis was the soil moisture content and soil salt content. The EC of each layer has been measured and the amount of salt which is present in each layer has been calculated.

3. RESULTS AND DISCUSSION

Chernozems (table 1) and **Phaeozems** (table 2) have a deep humus rich mollic horizon with a well developed crumb structure resulting from a high annual biomass production and a very high biological activity in the soil. Concentrations of soft powdery lime occur in the lower part of the soil profile as a main diagnostic property, separating the Chernozems from the Phaeozems. Downward percolation in spring leaches nutrients from the topsoil and lime accumulates in the subsoil. In the wetter areas at the boundary of steppe and deciduous forest, clay may also accumulate to form an argic horizon in the subsoil.

Table 1 - Hyposalic-Hyposodic Chernozems

Horizon	Ap	Atp	Am	ACK	Cca
Depth, cm	0-20	-34	-61	-72	-120
Coarse Sand, %	6,6	7,1	4,9	1,6	4,3
Fine Sand	47,5	35,6	36,0	45,4	59,3
Silt	15,8	22,5	17,8	18,0	16,2
Clay	30,1	34,8	41,3	35,0	20,2
Air Porosity, %	-2,51	3,59	-1,29		
Hydraulic Conductivity, mm/h	0,9	1,2	0,7	0,4	
pH	6,67	8,30	9,19	9,40	8,89
Humus, %	3,35	3,28	2,73		
ESP, %	-	10,08	13,29	8,85	8,50
CECs, me/100g	31,20	30,45	32,20	28,32	21,76
Soluble Salt, mg/100g		102,3	180,2	247,4	

Table 2 - Gleyic Phaeozems

Horizon	Ap	Am	AC	Cg ₂
Depth, cm	0-21	-46	-87	-
Coarse Sand, %	4,8	3,6	2,2	1,1
Fine Sand	33,9	38,3	47,7	50,9
Silt	27,0	22,8	22,2	22,3
Clay	34,3	35,3	27,9	25,7
Air Porosity, %	34,43	23,58	14,00	-
Hydraulic Conductivity, mm/h	22	8	8	-
pH	7,09	7,55	8,29	8,17
Humus, %	3,28	3,04	1,82	-
ESP, %				
CECs, me/100g	30,2	29,3	33,07	26,4
Soluble Salt, mg/100g	-	-	21,99	

The favorable physical and chemical properties, especially the high porosity and available water capacity, the high levels of organic matter and nutrients make these soils very fertile.

Gleysols typically occur in humid or semihumid environments in footslopes and low-lying situations on the landscape.

By definition, they do not occur in recent alluvial materials.

Hydric soils are soils that support or are capable of supporting wetland ecosystems, i.e., soil modifications are not needed to maintain or restore a wetland.

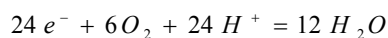
A reduced soil is one in which redox reactions have caused reduced forms of O, N, Mn, Fe or S to be present in soil solution.

Common reduced forms found in Hydric soils include: H₂O, N₂, Mn²⁺, Fe²⁺ and H₂S; their oxidized forms are: O₂, NO₃⁻, Mn⁴⁺, Fe³⁺ and SO₄²⁻, respectively.

Oxidation occurs when bacteria and other micro- and macroorganisms decompose organic matter and produce electrons (e⁻) and protons (H⁺).

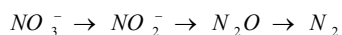
Oxidation is the norm; however, when a soil starts to become wet and eventually saturated, reduction reactions begin at microsites, if the soil is wet enough for a long period (some weeks) spread through the soil.

As long as air is present in soil water, all electrons produced by organic matter decomposition are consumed by O to make water:

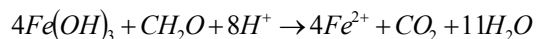
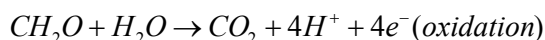
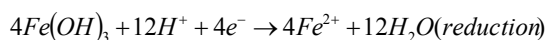


When most of the O has been reduced, a soil is said to be anaerobic. Anaerobiosis results in the reduction of nitrates (NO₃⁻) to form N (N₂) followed by Mn from Mn⁴⁺ to Mn²⁺, Fe from ferric (Fe³⁺) to ferrous (Fe²⁺), sulfates (SO₄²⁻) to form hydrogen sulfide (H₂S) and carbon dioxide (CO₂) to form methane (CH₄).

Nitrate in soil can be destroyed by the action of denitrifying bacteria when oxygen is in short supply. Under anaerobic conditions the nitrate ion can be used instead of O₂ as the acceptor of free electrons produced during respiration:



In soils, organic matter (CH₂O) is the primary source of electrons.



Soils that are oxidized most all of the time are usually red in color, and soils that are reduced most all the time are usually grayish in color.

Soils, with intermittent wetness can be a mixture of reds and grays. Redox concentrations are zones (reddish) where Fe/Mn is concentrated.

There are three types:

- (1) concretions;
- (2) masses, which are soft bodies with in the soil matrix;
- (3) pore linings, which are coatings on a natural soil surface (ped face, air pore, root pore).

Pore linings are the most reliable evidence of contemporary wetness; concretions reflect prior wetness.

For most of the world, wet soils are microbial active when air temperature is above approximately 5°C.

If microbes are active, if Fe/Mn oxides are present in the soil, if the soil is saturated, if the water is stagnant, if there is a source of usable organic water in the soil, and if O and N have been reduced then the electrons will reduce and translocate the Fe/Mn present and redox features will result.

The redox potential (E_h) of soil and water is a measure of electrochemical potential or electron availability within these systems.

$$E_h(mV) = E^0 - \frac{59}{n} \log \left(\frac{(Rd)}{(Ox)} \right) + 59 \frac{m}{n} pH$$

Typically in wetland soils, E_h values range from -300 to 700 mV with a total range of approximately 1000 mV. (Rogobete, Grozav, 2007)

In aerobic soils where the O₂/H₂O redox couple functions, the E_h range is between 300 and 700 mV.

Table 3 Mollic-Vertic Gleysols

	Apg	AzGr	ACzGr	CnGr
Clay	38,1	40,7	39,2	40,4
Silt	28,0	27,9	28,7	25,5
Bulk density	1,55	1,60	1,58	-
Field capacity	24,48	23,23	23,0	-
Ksat mm/h	0,65	0,45	0,50	-
pH	7,54	8,10	8,16	8,25
Humus	2,82	2,04	1,38	-
Pmobile,ppm	128,34	14,90	7,91	-
Kmobile,ppm	107	96	92	-

The soil type which has been studied is Fluvisols, and in accordance with WRB, it has a fluvic soil material within 25 cm of the surface and no horizons other than a histic, mollic, ochric, umbric, salic, histic or a sulphuric horizon.

Fluvisols developed in accumulating alluvial deposits.

Soil formation occurs upon layered alluvial sediment (which can normally be identified lower in the soil profile), each with a different organic matter content.

Fluvisols have a high groundwater table (1.5-2.0m) and exhibit gleyic properties or have a salic horizon.

Fluvisols are young soils and therefore show weak horizon differentiation.

Chemically, Fluvisols usually are rich, with a near neutral soil reaction.

Physically, Fluvisols may be wet through the presence of groundwater or flood water.

Freshly deposited Fluvisols usually are unripe and have a low bearing capacity (table 4).

Because of a content of soluble salt (546.7 mg/100g soil) and an ESP of 7.15%, this Fluvisols are considered moderate salinized and weak sodic.

Salic - sodic Fluvisols has a coarse texture below 36 cm, and in the topsoil – loam.

It can be noticed that the Salic - Sodic Fluvisols has a great value of hydraulic conductivity.

Table 4 Salic –Sodic Fluvisols

Horizon	Ap _{sc}	Ao _k	I C _k	II C _k
<i>Depth, cm</i>	0-24	-36	-47	-57
Coarse sand, %	0.2	0.2	0.5	0.5
Fine Sand, %	47.2	44.5	75.5	91.5
Silt, %	21.8	24.0	14.3	5.3
Clay, %	30.8	31.3	0.7	2.7
Bulk density, g/cm ³	1.5	1.4	1.3	-
Air Porosity, %	13.36	13.77	6.81	-
Hydraulic Conductivity, mmh ⁻¹	3.8	40	13	40
pH	8.55	8.62	8.54	8.59
Humus, %	2.73	2.42	0.62	-
ESP, %	7.15	0.56	-	-
CECs, me/100g	31.33	28.70	-	-
Soluble Salt, mg/100g	546.7	48.3	-	-

With the help of the following two equations the water movement may be calculated from soil salinity (Rogobete, 1994):

$$C = \frac{1000}{A} \cdot B \quad (2)$$

and

$$Z = \frac{D \cdot \delta \cdot B}{100} \quad (3)$$

where C=salt concentration in g per liter;
A=soil moisture content (percentage on dry weight basis);
B=soil salt content (percentage on dry weight basis);
Z=amount of salt in grams, present in the rootzone of D cm thickness and having a surface of area of 1 cm²;

δ =bulk density.

The equation can best be used to calculate the capillary rise.

When there is a downward water movement, for instance during a rainy period problems of soil cracking, leaching efficiency are complicating the calculations considerably.

The basis of the following calculation is that the salt concentration in the deeper soil layers remains practically constant over the period under consideration.

Adding the salt contents of all layers yields the total amount of salt in the upper 1.20m of soil: Z_1 (beginning of period) is 0.493 gram, Z_2 (end of period) is 0.547 gram.

The concentration in the layer 120-150 cm did not change much:

$$C_1=8.2\text{g/liter}$$

$$C_2=8.6\text{ g/liter}$$

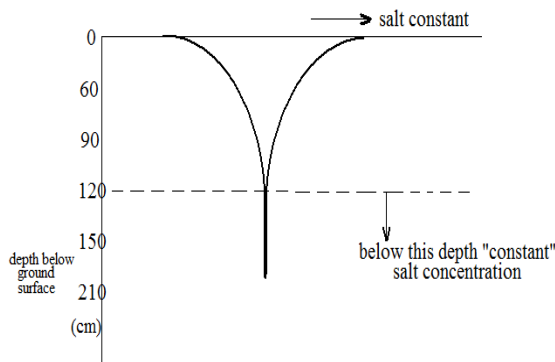
The concentration in the 150-180 cm layer remained the same. The bulk density $\delta=1.5$ the moisture content in the upper 1.20m of the profile decreased by 97mm.

It appears that the salt content increases by $0.547-0.493=0.054$ grams.

The average salt concentration in the 120-150 cm layer is 8.4 g/liter.

To effect the salt increase $0.054/8.4=0.006$ l/cm²=60mm of water was needed.

The total evaporation, therefore, is $60+97=157$ mm.



4. CONCLUSIONS

Fluvisols developed in the subsidence plain in accumulating alluvial deposits, have a high groundwater table.

Fluvisols are young soils and show weak horizon differentiation. In the dry period the movement of soil water consist predominantly in capillary rise, which has been transported the soluble salts in the topsoil.

Soil solution-soil matrix interactions can alter the geometry of the soil pores and thus may affect the soil hydraulic conductivity and water retentivity.

With the help of two equations, the water movement may be calculated from soil salinity, considered that the salt concentration in the deeper soil layers remains practically constant over the studied period.

The total evaporation is 157 mm.

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