Soil texture and layering effects on water and salt dynamics in the presence of a water table: a review

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Abstract:

Soil texture and its vertical spatial heterogeneity may greatly influence soil hydraulic properties and the distribution of water and solutes in the soil profile, and are therefore of great importance for agricultural, environmental, and geo-engineering applications such as land reclamation and landfill construction. This paper reviews the following aspects on water and salt dynamics in the presence of a water table: 1) the effect of soil texture on the extent of upward movement of groundwater in homogenous soils; and 2) the impact of soil textural layering on water and salt dynamics. For a homogenous soil, the maximum height of capillary rise (h_{max}) or the evaporation characteristic length (ECL) is closely related to the soil texture. When water table is deeper than h_{max}, water will evaporate at some depth below surface and salts will be retained below the evaporation front, causing the separation of water and salt. For layered soils, flow barriers (capillary and hydraulic barriers) can make the soil hold more water than a non-layered one. A capillary barrier may work when a fine-textured layer overlies a coarse-textured layer during infiltration or a coarse-textured layer overlies a fine-textured layer during evaporation, and a hydraulic barrier may occur when a poorly permeable layer exists in the soil profile. The extra water held by flow barriers may improve water availability to plants, and may at the same time increase salinization and other environmental risks. Under some special conditions, such as in seasonally frozen soils with a shallow water table, there is an additional soil salinization incentive caused by freeze-thaw cycles. Above all, further research is needed to understand the complex effects of soil texture and

layering on water and salt dynamics, especially in artificial soils such as reclaimed soils with contrasting properties.

Key words

Salinity, soil textural layering, evaporation characteristic length, capillary break, freeze-thaw effect

1. Introduction

Soil secondary salinization is usually caused by improper soil and water managements due to the lack of understanding of the effects of soil properties, groundwater depth and evaporation on soil water and salt dynamics. In arid and semiarid agricultural regions of the world, soil salinity can markedly decrease crop production and make water unsuitable for drinking. It is a serious problem that hinders the development of agriculture and society (Nulsen 1981; Bezborodov et al. 2010; Ben-Hur et al. 2001; Karpachevskii et al. 2008; Liu et al. 2009; Purdy et al. 2005). In geotechnical engineering, water and solute ponding or leaching also threats the success of soil cover rebuilding for the decommissioning of landfills or tailings piles which contain high concentrations of salts and other hazardous materials (Oddie and Bailey 1988; Sadegh-Zadeh et al. 2009). For example, in oil sands reclamation projects in Canada, soil salinity is considered one of the major issues that substantially impact the success of land reclamation (Renault et al. 1998; Purdy et al. 2005).

When salt migrates into the clean cover soil from below, the clean cover soil becomes salinized and the surface salinity negatively affects the growth of vegetation. The effects of soil salts on plant growth were quantified according to their EC values by USDA (Richards 1954). Soils with EC below 2dS/m are suitable for most plants, for even salt sensitive plants such as beans; but only a few salt tolerant plants such as barley can live in soils with EC up to 16 dS/m without much yield reduction (Rhoades and Loveday 1990).

Soil texture can dramatically vary within a soil profile, from clay to sand, or even gravel. Soil texture strongly determines soil hydraulic properties, such as water permeability (saturated (K_s) and unsaturated water conductivities (K_θ)), and soil water retention characteristics such as soil water retention curve (SWRC), water holding capacity and air entry pressure value (Wösten et al. 2001). Soil texture is thought to be a major factor impacting water recharge to groundwater, especially when the soil is not covered by vegetation (Scanlon et al. 2006). Usually, the finer the soil texture, the greater the porosity and water retention capacity, but the slower the permeability. Saturated water conductivity can vary from 1×10^{-2} m/s to less than 1×10^{-9} m/s, and the porosity can vary from less than 0.30 to over 0.60, when the texture changes from clean sands to homogenous clays (Tietje and Hennings 1996). In heterogeneous soils, the textural configuration of soil layers with contrasting properties also strongly influences water and salt dynamics, and other important processes in soils (Shokri et al. 2010; Ma et al. 2011; Meiers et al. 2011). With diverse soil hydraulic properties, the hydraulic connection between surface soil and the groundwater may markedly vary even with the water table depth and other conditions the same. Consequently, the distribution of soil solutes in the profile is also affected because they usually transport with soil water.

Understanding water and solute movement in soils is relevant to understanding water and energy balances, terrestrial ecosystem functioning, and microbiological activities in the vadose zone (Shokri and Salvucci 2011). Because of the complexity of water and solute dynamics in heterogeneous porous media, early studies were mostly carried out on the hypothesis that the soil was homogeneous. However, it is not easy to find an ideal homogenous soil in real studies because of the heterogeneous geological soil forming environment, and hydrological, meteorological, and biological disturbances (Phillips 2001). Furthermore, when soils are artificially constructed, such as in land reclamation in the Canadian oil sands (Fung and Macyk 2000; Purdy et al. 2005; Meiers et al. 2011) and in surface mined sites around the world (Oddie and Bailey 1988; Bowen et al. 2005; Fulton and Wells 2005; Sadegh-Zadeh et al. 2009), soil layering with contrasting textures is very common. In a soil with different textural layers, water and solute dynamics become more complex due to the discontinuity of the soil properties. Thus, it is of great importance to further understand the effect of soil texture and textural layering on soil water and salt dynamics. Improved understanding of the effects of soil texture and textural layering on evaporation, salt migration, and water and salt distribution in the presence of a water table will help address salt migration and soil secondary salinization problems.

This paper provides a synthesis of studies on soil water and salt dynamics in different soils and focuses on the following questions:

- How does soil texture affect the extent of water and salt movement (especially during the evaporation process with upward water movement) in texturally homogeneous soils?
- How does soil layer configuration affect water and salt movement during both evaporation and infiltration processes, considering different configuration scenarios such as when a coarse-textured soil layer is placed above a fine-textured soil layer or in a reverse configuration?

- What are the mechanisms that determine the extent of vertical water and salt movement in both homogenous and texturally layered soils?
- What are the characteristics of water and salt dynamics in seasonally frozen soils (such as those in the boreal zone in the northern region) as compared with soils that do not freeze?

The sections that follow will address each of the above questions.

2. Effects of soil texture on soil water dynamics in homogeneous soils

In homogenous soils, the evaporation process that moves water and salt upward and extends the impact of saline groundwater brings the risk of salinization into the root zone. During the evaporation process, in the capillary zone of a homogenous soil profile, the soil near the water table is close to be saturated when an equilibrium is reached (Fig. 1); and as the distance to the water table increases, soil water content gradually decreases until it becomes negligible at the upper limit of the capillary rise (if the extent of groundwater doesn't reach the soil surface). Consequently, the soil profile can be separated into wet and dry zones. The distance from the water table to the interface between the wet and dry zones is the maximum capillary rising height (h_{max}), and the interface is termed the **evaporation front** (Fig. 1; Shokri and Salvucci 2011). With the rise of the water table, there is a critical depth at which the maximum height of capillary water rise equals the water table depth. Under this situation, the evaporation front is at the soil surface, and this is the maximum depth that the water table is hydraulically connected to the soil surface. This depth is termed **evaporation**

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characteristic length (ECL) (Fig. 1; Lehmann et al. 2008) .

Hydraulically, the ECL corresponds to the pressure head (in terms of length) on the SWRC close to residual water content under hydrostatic conditions (Shokri and Salvucci 2011). When the water table is lower than the ECL, the soil above the ECL becomes too dry to permit appreciable liquid form of water to be transported; and consequently, vapor diffusion becomes the dominant form of water movement through the overlying dry layer. The matric potential above the evaporation front decreases significantly without an appreciable change in the water content (when it is close to the residual water content); and the evaporation rate significantly decreases due to the loss of soil hydraulic connectivity (Steenhuis et al. 2005; Lehmann et al. 2008; Shokri and Salvucci 2011). ECL is a critical term describing the textural influence on soil water and salt dynamics, and it may widely vary in differently textured soils with contrasting hydraulic properties.

There are some other terms similar to ECL that illustrate the extent of the upward movement of groundwater through capillary rise. **Extinction depth** refers to the water table depth at which the fraction of evapotranspiration from groundwater is zero (Shah et al. 2007). In theory, the extinction depth can be deeper than ECL because groundwater can still be evaporated in the vapor form if its depth equals ECL. Another difference is that extinction depth not only concerns evaporation from the soil surface but also transpiration of plants. In soil salinity studies, the term **critical water table depth** is used to indicate "the depth beyond which water rising by capillarity would not cause salinization of arable soil horizons" (Saleh and Troeh

1982). It is the maximum saline groundwater depth that causes salinization under the impact of capillary rise and evaporation, and below that the evaporation rate significantly decreases (Staley 1957; Anat et al. 1965; Nulsen 1981; Bruch 1993). This critical depth in sandy soils was thought to be approximately equivalent to the air entry water potential value (in terms of length) on the SWRC (Staley 1957), which is different from the hydraulic meaning of ECL defined by Shokri and Salvucci (2011).

2.1 Effects of soil texture on evaporation and quantitative studies of the ECL

Relationships among evaporation, soil texture and groundwater depth have been studied since early last century; and the main emphasis of those early studies was mostly on evaporation rates in homogeneous soils with the presence of water tables. Moore (1939) was one of the first to study the effect of soil hydraulic properties on upward water flow, and he found that the finer-textured soils had higher evaporation rates with shallow water tables. Staley (1957) and Anat et al. (1965) reported that the water table depth under which no significant evaporation could be detected was related to the properties of the capillary pressure-desaturation curve; and the accurate measurement of the soil hydraulic properties was important for determining the maximum evaporation rate. An experiment by Hellwig (1973) showed that the depth below which the evaporation rate was reduced to less than 10% of an open water surface evaporation amount was about 600 mm in a sandy soil (with an average particle size of 0.53 mm) (Table 1). Rasheed et al. (1989) suggested that there are two main factors affecting the evaporation process: the external meteorological conditions and the soil characteristics, and illustrated that the SWRC, especially the high suction section of the SWRC, was paramount in affecting evaporation rates. Therefore, the accurate measurement of SWRC was of vital importance for simulating the evaporation process (Wilson 1990). It was found that there were close hydraulic connections between the unsaturated and saturated zones in soils with texture varies from clay to sand for water tables within 0.5 m of the soil surface, and nearly all evapotranspiration came from groundwater (Shah et al. 2007).

Quantitative relationships between evaporation rate and soil hydraulic properties and the depth of water table were often expressed in the form of regression functions (Moore 1939; Gardner and Fireman 1958; Anat et al. 1965; Rasheed et al. 1989; Jalili et al. 2011; Shokri and Salvucci 2011). Regression functions between evaporation rate and water table depth were usually considered nonlinear. Studies by Veihmeyer and Brooks (1954) revealed that evaporation rates were significantly reduced when the water table was lowered to a certain depth, and the evaporation rate from uncropped soils with a high water table was not linearly related to the water table depth. Gardner and Fireman (1958) derived numerical solutions to the unsaturated steady state flow equation with a water table. According to the solutions, the maximum evaporation rate is determined by a function of the hydraulic conductivity and the depth to the water table; and there were nonlinear relationships between evaporation and water table depth in very differently textured soils. In fact, the decline of evapotranspiration with increasing depth to water table was better simulated by an exponential decay function than linear equations (McDonald and Harbaugh 1988; Rasheed et al. 1989;

Banta 2000; Shah et al. 2007). Table 1 summarizes the relationships between water table depths and height of capillary rise or evaporation extinction.

2.2 Theoretical considerations of soil texture effects on water movement

Capillary water movement is highly regulated by the pore size distribution in porous media (Campbell 1985). For capillary water movement above the water table, the smaller the soil pore size, the higher the upward groundwater movement. The average pore size and the pore size distribution are closely related to soil texture (Campbell 1985), with clay and loamy soils usually having relatively greater heights than sandy soils. However, if the pore size is too small (in some clay soils), the upward movement of groundwater can also be retarded because of the very slow flow rate.

The maximum height that capillary water can reach in a certain soil pore size at balance (h_{max}) can be calculated by the **Jurin** equation:

$$h_{max} = \frac{4\gamma\cos\varphi}{D\rho g} \tag{1}$$

where γ is water surface tension, φ is contact angle, D is pore diameter, ρ is water density, and g is gravitational acceleration. This equation offers a method to calculate h_{max} based on mechanical equilibrium. However, it is difficult to directly apply this method to a real soil because of the various soil pore sizes, which are associated with soil texture, soil structure, and the degree of soil compaction, and so on.

In an unsaturated soil with a water table (Fig. 1), the flow rate of water between points A and B can be calculated by Darcy's law, which is expressed as:

$$q = K_\theta \Delta \psi_{AB} = K_\theta (\frac{\Delta S_{AB}}{\Delta h} - 1) \quad (2)$$

where q is the rate of water flow (cm/h), ΔS_{AB} is the difference in matric suction between points A and B (cm), Δh is the distance between A and B (cm), and K_θ is the unsaturated hydraulic conductivity (cm/h). Equation 2 illustrates that the larger the gradient in matric suction, the more water recharge to the unsaturated soil. With the increasing water content in the unsaturated zone, the difference in soil matric suction between points A and B will decrease. Under ideal conditions, when it gets hydraulically balanced, the difference in matric suction between the two points will eventually disappear and matric suction at any point x will be equal to the gravity at that point, with $\nabla \psi_x = 0$, $q = 0$, and capillary rise reaches h_{max}.

van Genuchten (1980) developed a relationship between effective water content (θ_{eff}) and hydraulic potential related parameters α (related to the inverse of the air entry value) and n (related to the pore size distribution) as:

$$\theta_{eff} = \frac{\theta - \theta_r}{\theta_s - \theta_r} = \left[1 + (\alpha h)^n\right]^{(\frac{1}{n} - 1)} \quad (3)$$

where θ_r is the residual water content at which the matric potential tending to be a negative infinity, θ_s is the saturated water content. Based on this equation, h_{max} can be estimated at the onset of the liquid discontinuity ((Shokri and Salvucci 2011), using a linearization treatment of the SWRC (Lehmann et al. 2008):

$$h_{max} = \frac{1}{\alpha}\left(\frac{n-1}{n}\right)^{\frac{1-2n}{n}} \quad (4)$$

Equation 4 is used for predicting capillary water rise in coarse-textured soils where the viscous effect is negligible. However, in fine-textured soils such as clay soils, an extra viscous dissipation characteristic length which corresponds to the dissipation of viscous effect should be included and that would result in the reduction of the maximum height of capillary rise above the water table (Lehmann et al. 2008).

3. Effects of soil layer configuration on soil water and salt movements

3.1 Water dynamics in layered soils

Layered soils are widely distributed in the field and they may be formed by eluviation–illuviation, bioturbation and surface removal, erosion, deposition or inheritance, or a combination of several of the above mechanisms (Phillips 2001). In a layered soil, water dynamics are affected not only by the innerlayer properties and the thickness of the layers, but also by their spatial configuration. It is quite different from that in the homogeneous soils because of the discontinuity of hydraulic properties. Consequently, salt accumulation in the root zone is also tightly linked to the configuration of differently textured soil layers in saline soils.

Layering in the soil can affect water movement in two directions: the downward infiltration and water redistribution processes, and the upward capillary rise during the evaporation (or evapotranspiration) process. Soils with textural layering may hinder vertical water movement and allow the soil to store more water than those without textural layering during both infiltration and evaporation processes (Bruch 1993; Zornberg et al. 2010; Huang et al. 2011). Zettl et al. (2011) compared the field capacities within 1 m depth of reclaimed soil profiles in the oil sands region in Alberta, Canada, and showed that soils with textural layering had higher field capacities than those homogenous ones and supported more productive ecosites. This kind of textural heterogeneity created by placing layers with contrasting properties is very common in reclamation practices. On a macro level, textural layering actually forms some artificial "barriers" to make the soil hold more water. The increased water holding capacity in layered soils can be explained by several mechanisms as described below.

3.1.1 Effects of soil textual layering on downward water movement

During infiltration and soil water redistribution, the downward water movement can be impeded by soil textural layering. Water stagnation and increased water storage in layered soils are caused by two phenomena: the capillary barrier and the hydraulic barrier, which are collectively referred to as flow barriers (Alfnes et al. 2004; Si et al. 2011).

The capillary barrier is a soil layer that has the capillary break effect, which is also referred to as a "wick" cover (Bruch 1993). The capillary break effect usually occurs during infiltration, often observed under unsaturated conditions, when a fine-textured layer overlies a coarse-textured layer in the soil profile. It is known that capillary water suction and water conductivity are closely related to soil texture (Campbell 1985). The layer causing the capillary break usually has a higher hydraulic conductivity than the neighboring layer when under a high moisture or saturated conditions, but has a lower hydraulic conductivity when it is relatively dry (McCartney and Zornberg 2010). If there is such a soil layer in the soil profile, when the profile is dry there would be no effective hydraulic connection in this coarse-textured soil layer: its hydraulic conductivity is much lower than that in the neighboring fine-textured soil, which is the opposite of the condition when they are close to saturation. In this case, when water flows from the overlying fine-textured soil layer into the unsaturated coarse-textured soil layer, water is not able to directly cross the interface because of the inefficient unsaturated hydraulic conductivity and the weaker water suction in the capillary barrier layer. Once the capillary break effect occurs, more water will be retained in the upper layer, which is supposed to normally drain to the deep under gravity in the case of no capillary barrier layer; and in this case the water storage capacity of the overlying layer is increased. With increasing water content in the upper layer, the difference between the unsaturated hydraulic conductivities of the fine- and coarse-textured layers at the interface decrease. Infiltrating water will get into the lower layer only if the suction at the interface drops to some degree at which the lower layer can establish an effective hydraulic conductivity; and in this case the capillary break effect vanishes (Bruch 1993; Si et al. 2011). The ratio of saturated hydraulic conductivities of the coarse- and fine-textured soil layers can be used as a criterion for selecting materials to build capillary barriers (Bruch 1993).

If the layer configuration reverses, with a coarse-textured layer overlying a fine-textured layer, water infiltration can also be impeded because of another kind of flow barrier: a hydraulic barrier. A hydraulic barrier is caused by the weak permeability of the underlying fine-textured layer (Hillel and Talpaz 1977). Due to the reduced rate of water percolating into the deeper layer, both kinds of flow barriers can result in increased lateral soil water movement, especially if the interfaces are sloped (Si et al. 2011).

Water movement in layered soils has long been studied. The capillary break effect was discovered in studying seepage from clay-blanketed reservoirs (Kisch (1959). Pavelic and Johnston (1993) used the concept of a **clogging layer** similar to a capillary barrier to describe the phenomenon of clogging caused by a layer with low permeability sitting on top of one with a higher permeability during infiltration. Furthermore, they indicated that the hydraulic processes within the subsoil below might dominate the flow rates through the profile. McCartney and Zornberg (2010) went further and found that the condition for water to break through the interface corresponded to the point on the drying-path of the SWRC of the barrier layer where it transitioned from residual to saturated conditions. When the breakthrough suction is reached, leakage rate into the coarse-textured layer was at a rate similar to the saturated hydraulic conductivity of the coarse-textured layer; and the information from the SWRC and K-function of the components of a capillary barrier can be used to predict the breakthrough suction and water storage expected in the fine-textured layer. The study of Zornberg et al. (2010) gave another explanation of the breakthrough condition: water will continue to accumulate above the interface until the suction at the interface reached a value at which the hydraulic conductivity of the coarse-textured material was no longer below that of the fine-textured material. In other words, water could break through into the coarse-textured soil only if an effective hydraulic connection is built up.

Capillary barriers are not stable because of the high water content in the overlying fine-textured layer and the high saturated hydraulic conductivity of the underlying coarse-textured layer. Under such conditions, vertical heterogeneous structures such as cracks, root channels and worm holes, and small inequalities in soil structure or other disturbances in the underlying coarse-textured layer will allow water to get into the coarse-textured layer in the form of preferential flow, which creates a hydraulic shortcut between the two layers. The preferential flow weakens the capillary break effect, and reduces the contact area of the underlying layer under the influence of the capillary barrier (Alfnes et al. 2004; Si et al. 2011).

3.1.2 Effects of soil textual layering on upward water movement

During the evaporation process, a soil interlayer can have either positive or negative effects on evaporation depending on its hydraulic properties, depth to the water table, and its thickness. The capillary break effect, which reduces the rate of upward water flow, also occurs in the evaporation process, when a coarse-textured soil layer overlies a fine-textured one (Bruch 1993; McCartney and Zornberg 2010; Shokri et al. 2010). When the capillary break effect occurs, the flow rate is lowered and the top layer becomes drier than if there is no capillary break effect. However, evaporation may also be enhanced if there is a coarse-textured layer lying near the water table, because the coarse-textured layer usually has higher hydraulic conductivity when it is close to saturation and there is usually high moisture content near the water table (Unger 1971). The thickness of the coarse-textured layer is another factor that determines whether it has a negative or positive effect on upward water flow. If the coarse-textured layer is too thick and the capillary rise in this layer is less than the thickness of the layer, water won't even get to the interface between the coarse-textured layer and the fine-textured layer above. In this case, the coarse-textured layer has negative effects on upward water flow and evaporation, regardless of where it is located. Generally, a coarse-textured interlayer close to the soil surface is more likely to have barrier effect on upward water flow and evaporation, but one close to the groundwater may have a positive effect if it is not too thick. The detailed upward water flow scenarios determined by the factors above are discussed in the next section.

It was found in the 1960's that the spatial configuration of soil layers could lead to quite different evaporation rates: the evaporation rate in a fine-textured soil overlying a coarse-textured soil was smaller than that in a homogeneous profile of a fine-textured soil, but the reversed layer configuration increased the evaporation rate (Willis (1960). Unger (1971) introduced a 2.5 cm thick gravel layer at 0, 5, 15, 25 cm depths respectively in sandy loam and clay loam soils to study the effect of the gravel layer on soil water distribution and evaporation. Some of the treatments created the capillary break effect and some had increased upward flow rates: evaporations from soils with a gravel layer at the soil surface or at 5 cm depth were slower than that from the control; but evaporations from soils with gravel at 15 or 25 cm depth were generally greater than that from the control. Computer simulations by Bloemen (1980) demonstrated that capillary rise in multi-layered soils was strongly influenced by the depth of the water table; and in a soil profile with a clay layer overlying a sandy layer over the water table, there were significant linear relationships between the thickness of the lower sandy layer and the critical groundwater depth below which capillary water supply to the upper clay layer broke down . When the water table was at the critical depth, the thickness of the sandy soil layer above water table was close to a constant, and this constant represented the critical thickness for the sandy layer to become a capillary barrier. Shokri et al. (2008) found that hydrophobic layers in a soil profile interrupted water flow and reduced evaporative water loss as compared to a uniform hydrophilic column. Further study based on alternating layers of coarse and fine sands showed that air invading an interface between fine- and coarse-sand layers resulted in a capillary pressure jump and subsequent relaxation that significantly modified water distribution as compared with evaporation from homogeneous porous media (Shokri et al. 2010).

There are a wide range of applications of both capillary and hydraulic barriers in geo-engineering, agriculture, and other fields (Frind et al. 1976; Johnson et al. 1983; Badv and Mahooti 2005; Salem et al. 2010). Bruch (1993) indicated that construction of multi-layered soils might lead to water ponding or water scarcity in the upper soil layer by including a capillary barrier. For example, Stormont and colleagues carried out many studies on the application of capillary barrier materials in geo-engineering, including the constant anisotropy effect on capillary performance, the estimation method to evaluate the water storage capacity of capillary barriers, the breakthrough head of the capillary barriers and the influence of soil texture on the efficiency of capillary barrier, and the application of geosynthetic materials to buildup capillary barriers (Stormont 1995; Stormont and Morris 1998; Stormont and Anderson 1999; Stormont and Ramos 2004). A perched water table caused by the capillary break effect can periodically cause high moisture content in the soil even when the permanent water table is far deeper. This additional water supply makes the dynamics of water and solutes in a layered soil very different from that in homogenous soils. It can provide extra water for plant growth (McCartney and Allen 2008) or cause intense evaporation and an increased risk of salinization (Puengpan et al. 1990; Funakawa et al. 2000; Kim 2001; Tejedor et al. 2003; Kanzari et al. 2012).

3.2 Theoretical considerations of the effects of layering on water movement

Whether during infiltration or evaporation, for water to penetrate the interface of different soil layers there must be effective water connectivity between layers with the water entering layer having a lower water potential when there is an effective hydraulic connection. Even if a neighboring layer has a lower water potential, but if there is very low effective water conductivity, water cannot break through the interface because of a break in the hydraulic connection. The effective hydraulic connection between the neighboring layers is especially important when the soil moisture content is low.

Two SWRCs in two different textured soils are shown in Fig. 2, describing the relationship between volumetric soil moisture content and water potential. In the figure, soil 1 is coarse-textured, and soil 2 is fine-textured. The figure shows that when they have the same matric potential, the water content is much lower in the coarse-textured soil (soil 1) than in the fine-textured soil (soil 2); and at the same soil water content, the coarse-textured soil has a lower matric suction than the fine-textured soil. The capillary break effect may occur when water comes from soil 2 to the interface: in downward flow during infiltration (in this case, soil 1 underlies soil 2) or in upward flow during evaporation (soil 1 overlies soil 2).

During infiltration, whether water would stagnate at the interface depends on the SWRC of the coarse-textured soil; and water can get into the coarse layer only after the coarse-textured layer has formed effective hydraulic connections inside itself and with the neighboring layer (McCartney and Zornberg 2010). If soil 2 overlies soil 1, infiltration water can only get into soil 1 when the matric suction is lower than s_1. This means that the capillary break effect occurs only when water suction in soil 1 is greater than s_1 during infiltration, which is the low moisture (high water suction) part of the SWRC of the underlying coarse-textured soil.

During the evaporation process, things get more complicated: as mentioned above, the water movement is influenced not only by the hydraulic properties of both soils but also the thickness of the lower fine-textured soil layer, because the direction of the resultant force (mainly gravitational and matric potentials) varies with the variation of soil properties and the underlying layer. Taking soil 1 overlying soil 2 in Fig. 2 as an example:

a) When the thickness of soil 2 is greater than its h_{max} (should be much greater than s_1), capillary water only exists in the underlying soil 2, and cannot get into soil 1.

b) When the thickness of soil 2 is s_2 that is larger than s_1 but smaller than the h_{max} of soil 2, capillary water can reach the interface. However, when it is equilibrated, the minimum water suction at the interface would be s_2, still larger than the maximum water suction in soil 1 that is needed to establish an effective water conductivity. In this case, water still cannot break through the interface.

c) When the thickness of soil 2 equals s_3, which is less than s_1, capillary water can still reach the interface; but water movement will be impeded under the interface and cause the capillary break effect. With more water moved to the interface, the suction at the interface drops to s_1, and keeps decreasing to s_3. In this process, water eventually breaks through the interface and gets into the upper layer. Fig.3 illustrates the soil water potential and soil water distributions at equilibrium in a layered soil profile when capillary water crosses the sand interlayer during evaporation. Note that the water suction is continuous in the whole soil profile but the water content is not, and the distribution of matric suctions in this layered soil is similar to a profile without layering. However, the water content distribution maybe not continuous in the profile, and there are jumps at the loam-sand and sand-loam interfaces.

From the analysis above, it can be seen that it is of crucial importance to determine the turning point at which matric suction transits from residual to saturated conditions or effective water conductivity is established on the SWRC of the soil layer in the downstream direction. However, in real soils, if the flow rate is too low to reach equilibrium, some pseudo-equilibrium states can be formed. These pseudo-equilibrium states may make the observed results different from theoretical predictions (Hillel and Talpaz 1977; Yang et al. 2006).

4. Soil texture and their spatial configuration affect salt dynamics

Salt transport to the soil surface is mainly attributed to convection due to upward water movement in response to evapotranspiration, diffusion due to a salinity gradient with depth, and restricted drainage flow caused by the flow barrier effect (Kessler et al. 2010). Salt accumulation at the soil surface is not only related to evaporation but also to properties of soils in the unsaturated zone. Both the soil texture and the layering of differently textured soils affect salt migration. Continuous water supply, intense evaporation, and the existence of potential sources of salts in the profile or input of salts with water are considered prerequisites for salinization to occur.

4.1 Effects of soil texture on salt migration and distribution in homogenous soils

In homogenous soils, salt dynamics are highly influenced by both the evaporation demand and the height of capillary water rise, which are highly controlled by soil texture and groundwater depth. With a shallow water table, the groundwater may reach the soil surface through upward capillary water flow, which would maintain effective hydraulic connection in the soil profile above the water table. In this situation, groundwater can reach and be evaporated at the soil surface at a relatively high rate, which can bring salts up from deeper soil horizons (Nulsen 1981; Rasheed et al. 1989; Jalili et al. 2011). Once there is intense upward water flow, the risk of salinization at the soil surface will increase, even if the salt content in groundwater is low and there is no saline layer in the profile.

As described above, a hydraulic discontinuity in the soil profile can be caused by the evaporative demand greater than the rate of soil water supply to the soil surface when the water table is deeper than h_{max} (Rose et al. 2005); and the soil profile with an evaporation front can be divided into three zones from the water table up: liquid flow zone, mixed liquid-vapor flow transition zone, and vapor flow zone (Rose et al. 2005; Gowing et al. 2006). In this case, salt is detained in the transition zone below the evaporation front and little salt can migrate into the upper layer by diffusion (Merrill et al. 1983; Badv and Mahooti 2005). The critical depth at which salinity inhibits plant growth is often in the 1-2 m range (Talsma 1963); and under dryland conditions, critical depth was generally within the 1-6 m range which was greater than that under irrigation (Peck 1978). Nulsen (1981) illustrated that the critical depth of a saline water table in the wheat belt of Western Australia (soil textures range from loamy sand to sandy clay) for agricultural production (barley and wheat) is 1.5-1.8 m. Zhou and Li (2011) indicated that the critical water depth for alleviation of salinization in the lower reaches of the Tarim River of China (inland arid region with alkalized desert soil or salinized meadow soil) was about 6 m.

4.2 Effects of soil layering on salt dynamics

Soil layering affects salt transport in both convection and diffusion. Soil layering changes salt transport by convection as a result of the variation of soil moisture

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condition; and enhances salt diffusion due to the salinity gradient between salt-free and salt-rich layers. For example, in some artificially constructed soils such as in reclaimed soils with salt-rich underlying soil layers in the oil sands region of Canada, diffusion could be the major form of salt transport due to strong salinity gradients between layers, the low drainage rate due to flow barriers, and the weak evaporative demand in the boreal area; and diffusion could drive the migration of salt 20- 40 cm towards the clean cover soil layer above the saline layer within 4 years after reclamation (Kessler et al. 2010). Salt accumulation in a soil profile varies with evaporation rate, the position of the evaporation front, and the flushing effect of drainage water (Price 2005). The layering effect on salt accumulation in the root zone has been studied in the past, and various kinds of materials have been used to build artificial layers at or below the surface for relieving the salinization stress. Another effect of soil layering on salt dynamics is that the dilution on salt concentration because of the increased water holding capacity and increased water content in layered soils (Sadegh-Zadeh et al. 2009); furthermore, the salt can be eluted to deeper soil layers when the capillary barrier vanishes during the rainy season (Tejedor et al. 2003; Guo et al. 2006).

Mulching the soil surface is one of the most frequently used methods for water conservation and prevention of salt accumulation in the surface soil in arid and semi-arid regions; the type of mulching that can be used includes wheat straw (Bezborodov et al. 2010), volcanic mulch (mulched with tephra) (Tejedor et al. 2003), layered mulch (comprises of a light-colored mineral, farmyard manure, and common reed) (Sadegh-Zadeh et al. 2009), as well as hydrophobic polymers that can be sprayed at the soil surface as an aqueous solution (Al-Kalbani et al. 2003). It has been demonstrated that mulching is an effective way to reduce evaporation and salt accumulation, and increase moisture content in the root zone, which are all beneficial for crop survival and yield enhancement. Materials with a capillary break effect have also been used beneath the soil surface as an interlayer for salinity control. Guo et al. (2006) suggested that cutting off capillarity using a coarse-textured layer in the subsoil can prevent the dissolved salt in groundwater migrating into the surface soil even with a shallow water table, and the salts can be leached to deeper layers in the rainy season. Yin et al. (2012) used slag as a salt isolating layer below the root zone and demonstrated that adding a 20 cm slag layer underlying a 10 cm straw layer and a 10 cm farmyard manure layer at 100 cm depth increased water retention in the root zone, decreased EC, and slowed down the infiltration process that was important for the plants to survive.

4.3 Water and salt dynamics in seasonally frozen soils

Seasonally frozen soils are widely distributed in frigid and cold temperate zones, such as northern Canada, Siberia, and northern Europe. In seasonally frozen soils, the freeze-thaw effect can be another important risk for soil salinization, especially when the water table is shallow and there are sources of salt in the soil (Zhang and Wang 2001; Lopez et al. 2007; Liu et al. 2009). In these soils, water and salt dynamics are highly influenced by the freeze-thaw cycles, which is very different from the salinization process in arid or semiarid regions caused by intense evaporation. In seasonally frozen soils, salinization can occur even in soils where the groundwater is much deeper than the critical water depth in soils that do not seasonally freeze and when the soil texture is conducive for water movement (Zhang and Wang 2001; Liu et al. 2009; Luo et al. 2011).

In seasonally frozen soils, the redistribution of water and salt caused by freeze and thaw cycles are closely linked to temperature changes in the soil (Bing and He 2007). When the soil freezes in the presence of a shallow water table, soil water, along with dissolved salts, has the tendency to migrate to the freezing front because of the decreasing water content and water potential there due to water becoming ice. As a result, groundwater with dissolved salts transports into the frozen zone by intense upward capillarity and gets frozen there, causing the enrichment of water and salt in the frozen zone, when the soil textural composition is conducive for soil water movement (Kozlowski 2003; Li et al. 2008; Luo et al. 2011). This phenomenon can also lower the phreatic water table, similar to that caused by intense evaporation. However, salt migration during the freezing process is affected by soil texture, the initial water content, and salt concentration and composition, and other factors. Bing and He (2007) reported that the direction of salt movement in the soil when it freezes depends on the soil permeability: salts will migrate from the freezing zone to the unfrozen zone in soils with high permeability but in the opposite direction in soils with low permeability.

During the thawing process, the frozen layer thaws from the top and bottom at the same time. Before the frozen layer is completed thawed, the unthawed layer forms a seasonal hydraulic barrier that prevents the thawed water from percolating to deeper layers. Water accumulates above the unthawed layer which leads to intense evaporation, leaving the salts in the surface soil. As a result, the perched water caused by freeze-thaw effects above the unthawed layer is an important reason causing salinization rather than the critical depth of the real groundwater. (Zhang and Wang 2001).

Salinization caused by the freeze-thaw effect can be minimized by altering the hydraulic condition in seasonally frozen soils. Mulching with plastic films, snow or other materials is thought to be an effective way to decrease the intensity of upward water movement by reducing evaporation and maintaining higher surface soil temperatures to prevent the soil from becoming frozen or reduce the thickness of the frozen soil layer. Breaking up the capillary pore channels between the topsoil and subsoil by plowing before the ground freezes also helps to alleviate salt accumulation at the soil surface ((Luo et al. 2011).

Salts contained in the soil also have some reactive effects on soil water dynamics. For example, salt precipitation in soil pores and the formation of salt crust have apparent effects on evaporation (Chen 1992; Nachshon et al. 2011). Those phenomena should not be ignored when studying highly saline soils. A pseudo steady-state model that considers the evaporation difference between non-saline and saline water showed that under salt-saturated conditions the evaporation front was deeper by a factor of 1.11 as compared with non-saline conditions, regardless of the soil type (Gowing et al.

2006). Nachshon et al. (2011) reported that salts preferably precipitate out in fine-textured layers, clogging small pores which artificially increases the average size of the pores remaining in the soil and forces vapor transport via large pores, further affecting water and salt movement in the soil. And they also pointed out that evaporation in homogeneous soils was more strongly suppressed by salinity compared with heterogeneous soils.

5. Conclusions

(1) In homogenous soils, h_{max} is considered equivalent to the water suction on the SWRC close to residual water content under hydrostatic conditions. When the water table is lower than h_{max}, water will evaporate at the depth that capillary water can reach, and salt will be left below this evaporation front, causing the separation of water and salt in the soil, a condition that will minimize the risk for upward salt migration to the soil surface.

(2) In layered soils, water holding capacity in the whole soil profile is increased by capillary and hydraulic barriers. During infiltration, a capillary barrier usually occurs when a fine-textured layer overlies a coarse-textured layer. In this case, the occurrence of the capillary break effect depends on the hydraulic properties of the coarse-textured soil, or rather the soil water potential of the coarse-textured soil at which the water content in this soil transitions from residual to saturated conditions. A hydraulic barrier usually occurs when a relatively coarse-textured soil overlies a soil with low permeability (fine-textured). With much infiltration water, the salt

concentration in layered soils can be diluted and eluted to deeper soil horizons.

(3) During evaporation, a capillary barrier occurs when a fine-textured layer underlies a coarse-textured layer. In this case, the occurrence of the capillary break effect is determined by the SWRCs of both soils and the thickness of the lower fine-textured soil. When it reaches equilibrium in the soil profile, groundwater can get into the coarse-textured soil only when the thickness of the lower fine-textured layer is less than the suction at the residual point of the coarse-textured soil. Moreover, whether water can pass a coarse-textured interlayer is also related to the h_{max} of the interlayer. When there is intensive evaporation demand, the common ways to prevent salt accumulation in the root zone are: mulching to reduce evaporation, and cutting the capillarity using artificial soil layers.

(4) In cold regions, some special soil salinization risk should be paid more attention to. Freeze-thaw cycle is a special phenomenon in seasonally frozen soils that can change soil water and salt dynamics especially in soils with a shallow groundwater. The freeze-thaw effect has significant implications for salt migration in northern regions, such as the boreal region in North America and northern Eurasian regions. Much more research is needed to further understand the effects of freeze-thaw cycles on soil water and salt movement in layered or non-layered soils.

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References

Al-Kalbani, M. S., P. Cookson, and H. A. Rahman. 2003. Uses of hydrophobic siloxane polymer (Guilspare®) for soil water management. Water Int. **28**(2): 217-223.

Alfnes, E., W. Kinzelbach, and P. Aagaard. 2004. Investigation of hydrogeologic processes in a dipping layer structure: 1. The flow barrier effect. J. Contam. Hydrol. **69**(3-4): 157-172.

Anat, A., H. R. Duke, and A. T. Corey. 1965. Steady upward flow from water tables. Fort Collins, Colorado: Colorado State University.

Badv, K., and A. A. Mahooti. 2005. Chloride transport in layered soil systems with hydraulic trap effect. Environ. Technol. **26**(8): 885-897.

Banta, E. R. 2000. MODFLOW-2000, the U.S. geological survey modular ground-water model—documentation of packages for simulating evapotranspiration with a segmented function (ETS1) and drains with return flow (DRT1). In USGS Open-File Report 00-466. Reston, Virginia: USGS.

Ben-Hur, M., F. H. Li, R. Keren, I. Ravina, and G. Shalit. 2001. Water and salt distribution in a field irrigated with marginal water under high water table conditions. Soil Sci. Soc. Am. J. **65**(1): 191-198.

Bezborodov, G. A., D. K. Shadmanov, R. T. Mirhashimov, T. Yuldashev, A. S. Qureshi, A. D. Noble, and M. Qadir. 2010. Mulching and water quality effects on soil salinity and sodicity dynamics and cotton productivity in Central Asia. Agr. Ecosyst. Environ. **138**(1-2): 95-102.

Bing, H., and P. He. 2007. Experimental study of water and salt redistribution of clay soil in an opening system with constant temperature. Environ. Geol. **55**(4): 717-721.

Bloemen, G. W. 1980. Calculation of steady-state capillary rise from the groundwater table in multi-layered soil profiles. Zeitschrift Fur Pflanzenernahrung Und Bodenkunde **143**(6): 701-719.

Bowen, C. K., G. E. Schuman, R. A. Olson, and L. J. Ingram. 2005. Influence of topsoil depth on plant and soil attributes of 24-year old reclaimed mined lands. Arid Land. Res. Manag. **19**(3): 267-284.

Bruch, P. G. 1993. A Laboratory study of evaporative fluxes in homogeneous and layered soils, Civil. Eng., University of Saskatchewan.

Campbell, G. S. 1985. Soil physics with Basic, transport models for soil-plant systems: Developments in soil sciences 14. Amsterdam, The Netherlands: Elsevier.

Chen, X. Y. 1992. Evaporation from a salt-encrusted sediment surface - field and laboratory studies. Aust. J. Soil Res. **30**(4): 429-442.

Frind, E. O., R. W. Gillham, and J. F. Pickens. 1976. Applications of unsaturated flow properties in the design of geologic environments for radioactive waste storage facilities. Paper read at International Conference on Finite Elementsin Water Resources, at PrincetonUniversity, Princeton, NJ.

Fulton, J. P., and L. G. Wells. 2005. Evaluation of a mechanical system for reconstructing soil on surface mined land. Appl. Eng. Agric. **21**(1): 43-51.

Funakawa, S., R. Suzuki, E. Karbozova, T. Kosaki, and N. Ishida. 2000. Salt-affected soils under rice-based irrigation agriculture in southern Kazakhstan. Geoderma **97**(1-2): 61-85.

Fung, M. Y. P., and T. M. Macyk. 2000. Reclamation of oil sands mining areas. edited by R. I. Barnhisel, R. G. Darmody and W. L. Daniels. Madison, USA: American Society of Agronomy.

Gardner, W. R., and M. Fireman. 1958. Laboratory studies of evaporation from soil columns in the presence of a water table. Soil Sci. **85**(5): 244-249.

Gowing, J. W., F. Konukcu, and D. A. Rose. 2006. Evaporative flux from a shallow watertable: the influence of a vapour–liquid phase transition. J. Hydrol. **321**(1-4): 77-89.

Guo, G., K. Araya, H. Jia, Z. Zhang, K. Ohomiya, and J. Matsuda. 2006. Improvement of salt-affected soils, part 1: interception of capillarity. Biosystems Engineering **94**(1): 139-150.

Hellwig, D. H. R. 1973. Evaporation of water from sand, 4: the influence of the depth of the water-table and the particle size distribution of the sand. J. Hydrol. **18**(3-4): 317-327.

Hillel, D., and H. Talpaz. 1977. Simulation of soil water dynamics in layered soils. Soil Sci. **123**(1): 54-62.

Huang, M., S. L. Barbour, A. Elshorbagy, J. D. Zettl, and B. C. Si. 2011. Infiltration and drainage processes in multi-layered coarse soils. Can. J. Soil Sci. **91**(2): 169-183.

Jalili, S., H. Moazed, N. S. Boroomand, and A. A. Naseri. 2011. Assessment of evaporation and salt accumulation in bare soil: constant shallow water table depth with saline ground water. Scientific Research and Essays **6**(29).

Johnson, T. M., T. H. Larson, B. L. Herzog, K. Cartwright, C. J. Stohr, and S. J. Klein. 1983. A study of trench covers to minimize infiltration at waste disposal sites, edited by U. N. R. C. (USNRC).

Kanzari, S., M. Hachicha, R. Bouhlila, and J. Battle-Sales. 2012. Characterization and modeling of water movement and salts transfer in a semi-arid region of Tunisia (Bou Hajla, Kairouan) – Salinization risk of soils and aquifers. Comput. Electron. Agr. **86**: 34-42.

Karpachevskii, L. O., L. V. Yakovleva, and A. V. Fedotova. 2008. Soil salinization of the baer mounds in the Volga River delta. Eurasian. Soil Sci. **41**(2): 135-139.

Kessler, S., S. L. Barbour, K. C. J. van Rees, and B. S. Dobchuk. 2010. Salinization of soil over saline-sodic overburden from the oil sands in Alberta. Can. J. Soil Sci. **90**(4): 637-647.

Kim, P. J. 2001. Effects of compact layer on salt accumulation in plastic film house soils. Soil Sci. Plant Nutr. **47**(1): 67-77.

Kisch, M. 1959. The theory of seepage from clay-blanketed reservoirs. Geotechnique **9**(1): 9-21.

Kozlowski, T. 2003. A comprehensive method of determining the soil unfrozen water curves: 2. stages of the phase change process in frozen soil–water system. Cold Reg. Sci. Technol. **36**(1-3): 81-92.

Lehmann, P., S. Assouline, and D. Or. 2008. Characteristic lengths affecting evaporative drying of porous media. Phys. Rev. E **77**(5).

Li, Z., X. Liu, X. Zhang, and W. Li. 2008. Infiltration of melting saline ice water in soil columns: consequences on soil moisture and salt content. Agr. Water Manage **95**(4): 498-502.

Liu, Q., B. Cui, and Z. Yang. 2009. Dynamics of the soil water and solute in the sodic saline soil in the Songnen Plain, China. Environmental Earth Sciences **59**(4): 837-845.

Lopez, C. M. L., A. Brouchkov, H. Nakayama, F. Takakai, A. N. Fedorov, and M. Fukuda. 2007. Epigenetic salt accumulation and water movement in the active layer of central Yakutia in eastern Siberia. Hydrol. Process. **21**(1): 103-109.

Luo, J. M., F. Yang, Y. J. Wang, Y. J. Ya, W. Deng, X. P. Zhang, and Z. Liu. 2011. Mechanism of soil sodification at the local scale in Songnen Plain, Northeast China, as affected by shallow groundwater table. Arid Land. Res. Manag. **25**(3): 234-256.

Ma, Y., S. Feng, H. Zhan, X. Liu, D. Su, S. Kang, and X. Song. 2011. Water infiltration in layered soils with air entrapment: modified Green-Ampt model and experimental validation. Journal of Hydrologic Engineering **16**(8): 628-638.

McCartney, J. S., and J. M. Allen. 2008. Use of geosynthetic capillary barriers for increasing water storage capacity of soils in agriculture applications. In GRI 21 Conference. Cancun, Mexico.

McCartney, J. S., and J. G. Zornberg. 2010. Effects of infiltration and evaporation on geosynthetic capillary barrier performance. Can. Geotech. J. **47**(11): 1201-1213.

McDonald, M. G., and A. W. Harbaugh. 1988. A modular three-dimensional finite-difference ground-water flow model. Reston, Virginia: USGS.

Meiers, G. P., S. L. Barbour, C. V. Qualizza, and B. S. Dobchuk. 2011. Evolution of the hydraulic conductivity of reclamation covers over sodic/saline mining overburden. J. Geotech. Geoenviron **137**(10): 968-976.

Merrill, S. D., E. J. Doering, J. F. Power, and F. M. Sandoval. 1983. Sodium movement in soil-minespoil profiles: diffusion and convection. Soil Sci. **136**(5): 308-316.

Moore, R. E. 1939. Water conduction from shallow water tables. Hilgardia **12**: 383-426.

Nachshon, U., N. Weisbrod, M. I. Dragila, and A. Grader. 2011. Combined evaporation and salt precipitation in homogeneous and heterogeneous porous media. Water Resour. Res. **47**.

Nulsen, R. A. 1981. Critical depth to saline groundwater in non-irrigated situations. Aust. J. Soil Res. **19**(1): 83-86.

Oddie, T. A., and A. W. Bailey. 1988. Subsoil thickness effects on yield and soil water when reclaiming sodic minespoil. J. Environ. Qual. **17**(4): 623-627.

Pavelic, P., and C. D. Johnston. 1993. A laboratory study of the flow processes through layered profiles in relation to clogging in artificial recharge basins: Division of Water Resources, CSIRO

Peck, A. J. 1978. Note on the role of a shallow aquifer in dryland salinity. Aust. J. Soil Res. **16**(2).

Phillips, J. D. 2001. Contingency and generalization in pedology, as exemplified by texture-contrast soils. Geoderma **102**(3-4): 347-370.

Price, A. C. R. 2005. Evaluation of groundwater flow and salt transport within an underdrained tailings sand dam, Department of earth and atmospheric sciences, University of Alberta, Edmonton.

Puengpan, N., T. Subhasaram, P. Patcharapreecha, K. Miura, and H. Wada. 1990. Impermeable layers in salt affected sandy soils in norteast Thailand. Thai Journal of Soils and Fertilizers **14**(1): 40-48.

Purdy, B. G., S. E. MacDonald, and V. J. Lieffers. 2005. Naturally saline boreal communities as models for reclamation of saline oil sand tailings. Restor. Ecol. **13**(4): 667-677.

Rasheed, H. R., H. Al-Anaz, and K. A. Abid. 1989. Evaporation from soil surface in presence of shallow water tables. Paper read at Proceedings of a Symposium held in Baltimore, Maryland, at Baltimore.

Renault, S., C. Lait, J. J. Zwiazek, and M. MacKinnon. 1998. Effect of high salinity tailings waters produced from gypsum treatment of oil sands tailings on plants of the boreal forest. Environ. Pollut. **102**(2-3): 177-184.

Rhoades, J. D., and J. Loveday. 1990. Salinity in irrigated agriculture. Agronomy **30**: 1089-1142.

Richards, L. A. E. 1954. Diagnosis and improvement of saline and alkali soils. Washington D. C.

Rose, D. A., F. Konukcu, and J. W. Gowing. 2005. Effect of watertable depth on evaporation and salt accumulation from saline groundwater. Aust. J. Soil Res. **43**(5): 565.

Sadegh-Zadeh, F., B. J. Seh-Bardan, A. W. Samsuri, A. Mohammadi, M. Chorom, and G. A. Yazdani. 2009. Saline soil reclamation by means of layered mulch. Arid Land. Res. Manag. **23**(2): 127-136.

Saleh, H. H., and F. R. Troeh. 1982. Salt distribution and water consumption from a water table with and without a crop. Agron. J. **74**(2): 321-324.

Salem, M. A., W. Al-Zayadneh, and A. J. Cheruth. 2010. Water conservation and management with hydrophobic encapsulation of sand. Water Resour. Manag. **24**(10): 2237-2246.

Scanlon, B. R., K. E. Keese, A. L. Flint, L. E. Flint, C. B. Gaye, W. M. Edmunds, and I. Simmers. 2006. Global synthesis of groundwater recharge in semiarid and arid regions. Hydrol. Process. **20**(15): 3335-3370.

Shah, N., M. Nachabe, and M. Ross. 2007. Extinction depth and evapotranspiration from ground water under selected land covers. Ground Water **45**(3): 329-338.

Shokri, N., P. Lehmann, and D. Or. 2008. Effects of hydrophobic layers on evaporation from porous media. Geophys. Res. Lett. **35**: L19407.

Shokri, N., P. Lehmann, and D. Or. 2010. Evaporation from layered porous media. J. Geophys. Res. **115**: B06204.

Shokri, N., and G. D. Salvucci. 2011. Evaporation from porous media in the presence of a water table. Vadose Zone J. **10**(4): 1309-1318.

Si, B., M. Dyck, and G. W. Parkin. 2011. Flow and transport in layered soils: preface. Can. J. Soil Sci. **91**(2): 127-132.

Staley, R. W. 1957. Effect of depth of water table on evaporation from fine sand, Colorado State University, Colorado State University, Fort Collins, USA.

Steenhuis, T. S., A. G. Hunt, J. Y. Parlange, and R. P. Ewing. 2005. Assessment of the application of percolation theory to a water repellent soil. Aust. J. Soil Res. **43**(3): 357.

Stormont, J. C. 1995. The effect of constant anisotropy on capillary barrier performance. Water Resour. Res. **31**(3): 783-785.

Stormont, J. C., and C. E. Anderson. 1999. Capillary barrier effect from underlying coarser soil layer. J. Geotech. Geoenviron **125**(8): 641-648.

Stormont, J. C., and C. E. Morris. 1998. Method to estimate water storage capacity of capillary barriers. J. Geotech. Geoenviron **124**(4): 297-302.

Stormont, J. C., and R. Ramos. 2004. Characterization of a fiberglass geotextile for unsaturated in-plane water transport. Geotech. Test. J. **27**(2): 214-219.

Talsma, T. 1963. The control of saline groundwater: Landbouwhogeschool.

Tejedor, M., C. C. Jimenez, and F. Diaz. 2003. Use of volcanic mulch to rehabilitate saline-sodic soils. Soil Sci. Soc. Am. J. **67**(6): 1856-1861.

Tietje, O., and V. Hennings. 1996. Accuracy of the saturated hydraulic conductivity prediction by pedo-transfer functions compared to the variability within FAO textural classes. Geoderma **69**(1-2): 71-84.

Unger, P. W. 1971. Soil profile gravel layers: I. effect on water storage, distribution, and evaporation. Soil Sci. Soc. Am. J. **35**(4): 631.

van Genuchten, M. T. 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Sci. Soc. Am. J. **44**(5): 892.

Veihmeyer, F. J., and F. A. Brooks. 1954. Measurements of cumulative evaporation from bare soil. Am. Geophys. Union Trans. **35**(4): 601-607.

Wösten, J. H. M., Y. A. Pachepsky, and W. J. Rawls. 2001. Pedotransfer functions: bridging the gap between available basic soil data and missing soil hydraulic characteristics. J. Hydrol. **251**(3-4): 123-150.

Willis, W. O. 1960. Evaporation from layered soils in the presence of a water table. Soil Sci. Soc. Am. J. **24**(4): 239.

Wilson, G. W. 1990. Soil evaporative fluxes for geotechnical engineering problems, Department of Civil Engineering, Universityof Saskatchewan, Saskatoon.

Yang, H., H. Rahardjo, and E. C. Leong. 2006. Behavior of unsaturated layered soil columns during infiltration. Journal of Hydrologic Engineering **11**(4): 329-337.

Yin, X. L., G. D. Ding, and Y. Y. Gao. 2012. Response characteristic of soil moisture and salt to salt isolated layer in seashore saline soil. Applied Mechanics and Materials **195-196**: 1222-1226.

Zettl, J. D., S. L. Barbour, M. Huang, B. C. Si, and L. A. Leskiw. 2011. Influence of textural layering on field capacity of coarse soils. Can. J. Soil Sci. **91**(2): 133-147.

Zhang, D., and S. Wang. 2001. Mechanism of freeze-thaw action in the process of soil salinization in northeast China. Environ. Geol. **41**(1-2): 96-100.

Zhou, H., and W. Li. 2011. Effect of water resource on soil salinization of oasis in the lower reaches of Tarim River, China. Procedia Environmental Sciences **11**: 925-933.

Zornberg, J. G., A. Bouazza, and J. S. McCartney. 2010. Geosynthetic capillary barriers: current state of knowledge. Geosynth. Int. **17**(5): 273-300.

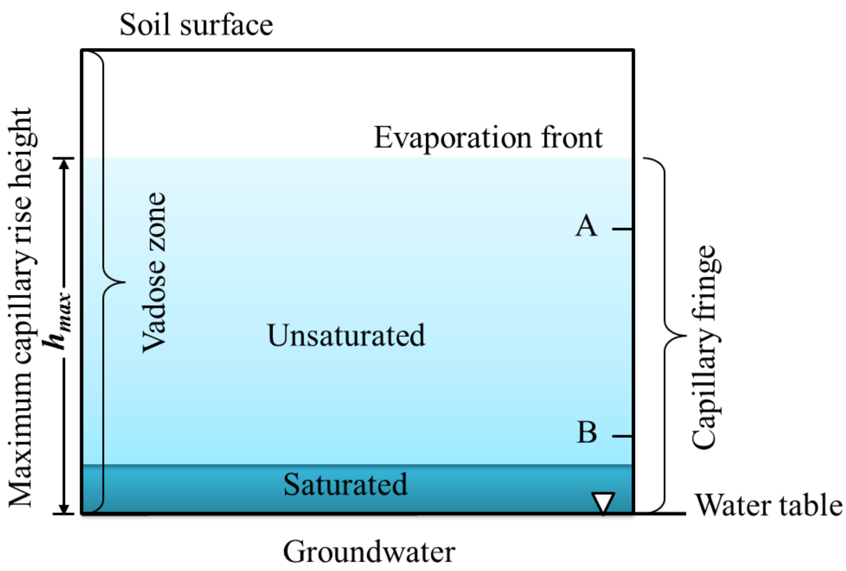

Fig. 1 Diagram of a soil profile with the presence of a water table

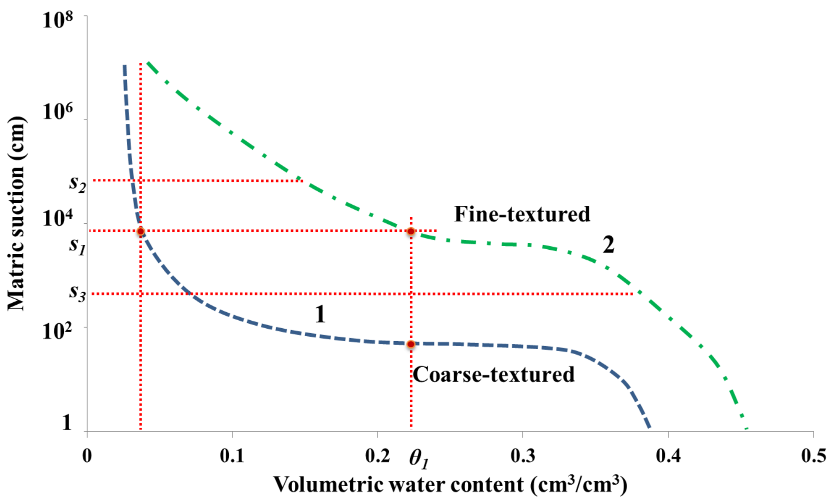

Fig. 2. The SWRCs of differently textured soils (The point on the SWRC of soil 1 at s_1 represents the transition of the matric suction from residual to saturated conditions, assuming it is also the turning point at which soil 1 establishes its effective water conductivity)

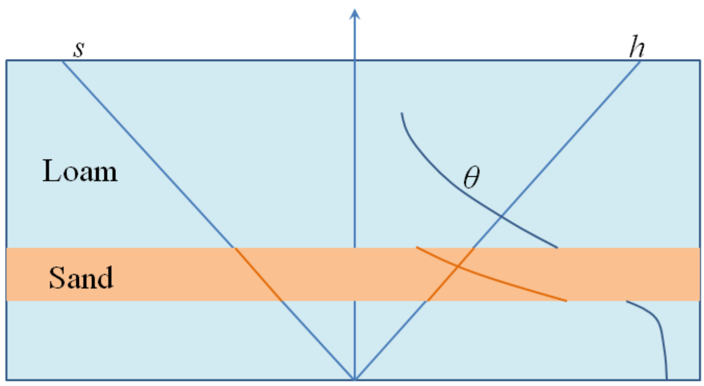

Fig. 3. Soil water potential and soil water distributions at equilibrium during evaporation when capillary water crosses the sand interlayer (s is the water suction in terms of length, h is the gravitational potential, and θ is the volumetric water content)

Table 1. Critical depth related to the reaches of capillary water or evaporation extinction in homogeneous soils

Method	Soil type	Critical depth (cm)			Remarks on the critical depth	Reference
Column experiment	Fine sand	76			No significant evaporation when the water table was under this depth	Staley (1957)
Numerical simulation	Fine sandy loam	150-200			Evaporations reduced to less than 0.1 cm/day	Gardner and Fireman (1958)
	Clay	200-250				
Column experiment	sand soil (average particle size = 0.53 mm)	60			Evaporation rate reduced to less than 10% of that from an open water surface	Hellwig (1973)
Column experiment	Loamy sand Silt loam Sandy loam Silt loam	150- more than 200			Evaporation rate was less than 0.1 cm/day (estimated values). These critical water table depths were 2-3 times longer than the observed capillary rising height (evaporation front).	Rasheed et al. (1989)
Numerical simulation		Bare	Grass	Forest	Extinction depth, where the contribution of groundwater to total evapotranspiration is less than 0.5%. Grass (rooting depth is1 m), Forest (rooting depth is 2 m).	Shah et al. (2007)
	Sand	50	145	250		
	Loamy sand	70	170	270		
	Sandy loam	130	230	330		
	Sandy clay loam	200	300	400		
	Sandy clay	210	310	410		
	Loam	265	370	470		
	Silty clay	335	430	530		
	Clay loam	405	505	610		
	Silt loam	420	515	615		
	Silt	430	530	630		
	Silt clay loam	450	550	655		
	Clay	620	715	820		