Comment on "Analytical estimation show low depth-independent water loss due to vapor flux from deep aquifers by John S. Selker [2017]"

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Selker [2017] recently presented an analytical estimation of evaporative fluxes from deep aquifers. Regardless of the type of soil, ambient temperature and relative humidity, Selker [2017] assumed the thickness of liquid region above the water table equals to zero. Here, we illustrate the limitation of this assumption and its impact on the estimation of the evaporative fluxes from deep aquifers.
Selker [2017] considered a scenario where the water table is located 200 m below the soil surface and assumed the evaporation rate can be estimated using the Fick’s law of diffusion as:

\[ q_v = -D_s \frac{M \Delta P}{\rho RT \Delta z} \]  

(1)

where \( q_v \) (\( ms^{-1} \)) is the vapor diffusive flux above the water table, \( D_s \) (\( m^2 s \)) is the effective vapor diffusion coefficient in porous media which can be estimated as \( D = e^{4/3} D_0 \) where \( e \) and \( D_0 \) corresponds to the porosity and vapor diffusion in open air, respectively,

\( M(\text{kg mol}^{-1}) \) is the molecular mass of water, \( \rho (\text{kg m}^{-3}) \) is the density of water, \( R(\text{kg m}^2 s^{-2} K^{-1} \text{mol}^{-1}) \) is the gas constant, \( T (K) \) is the temperature, \( \Delta P (\text{Pa}) = P_{sat} - P_{sur} \) is the gradient of vapor pressure with \( P_{sat}(\text{Pa}) \) and \( P_{sur}(\text{Pa}) \) indicating the saturated vapor pressure and vapor pressure at the surface, respectively and \( \Delta z (m) \) indicates the diffusion length (note that in Selker [2017], \( \frac{M \Delta P}{\rho RT} \) was represented as \( \Delta c \) corresponding to the gradient of vapor density in \( \text{kg m}^{-3} \)).

Selker [2017] assumed that \( \Delta z \) is equal to the depth of water table (the diffusion length), \( D_{WT} \). His underlying assumption was that liquid vaporization occurs at the water table and vapor diffuses through 200 m soil profile above the water table (see Figure 1). This assumption is incorrect and does not represent the true physics controlling water evaporation in the presence of a water table. One subtle issue about this assumption pertains to neglecting the significant role of liquid flow from the water table toward the surface. We will show that depending on the type of soil as well as the evaporative demand, the vapor diffusion length could be significantly smaller than the length used in Selker [2017]. The general theoretical lines described in Kamai and Assouline [2018] are followed here with a particular attention to evaluate the validity of the assumptions made in Selker [2017] and to determine how soil
texture influences the length of the unsaturated zone above the water table under different boundary conditions.

Figure 1. Schematic of different regions formed above the water table. The evaporation front corresponds to the interface between unsaturated and gas (dry) zone where the liquid vaporization occurs.

In the presence of a deep water table, liquid water flows upward with a liquid flux of $q_L (m s^{-1})$ through the soil profile up to a level located somewhere below the soil surface at the distance of $z_{EF} (m)$ from the water table (see Figure 1). This leads to formation of an evaporation front where liquid vaporization takes place and water vapor diffuses with a diffusive flux of $q_v (m s^{-1})$ from that evaporation front through the remaining part of the soil (see Figure 1). One could estimate $z_{EF}$ using the Richard’s equation as (Gardner, 1958; Sadeghi et al., 2012):
\[ z_{EF} = \int_{0}^{-\infty} \frac{dh}{1 + q_L / K(h)} \]  

(2)

where \( K(h) \) is the unsaturated hydraulic conductivity \((\text{ms}^{-1})\) expressed as a function of matric potential \( h \) \((\text{m})\). We used van Genuchten-Mualem model (Mualem, 1976; van Genuchten, 1980) to describe the relationship between unsaturated hydraulic conductivity and matric potential (the corresponding equations are presented in the Supporting Information section). To determine the location of the evaporation front, the integral in Equation (2) can be evaluated from the water table \((h = 0)\) to the evaporation front. At the evaporation front, we assume the matric potential approaches infinite negative pressure because the location of the evaporation front \( z_{EF} \) \((\text{m})\) corresponds to the position where water saturation equals the residual saturation.

Under the steady-state condition, the water flux through the soil profile is constant. This means that there exists a unique position above the water table where the liquid and vapor flux are the same and equal to the evaporation rate \( E \) \((\text{ms}^{-1})\) such that:

\[ q_L = q_v = E \]  

(3)

This position indicates the location of the evaporation front \( z_{EF} \) (note that the position of the water table is considered as \( z = 0)\). We developed a code in MATLAB to solve equation (3) numerically with the aim of determining the variation of \( z_{EF} \) \((\text{m})\) under various soil textures and evaporative demands. To do so, following Selker [2017] we assumed a linear thermal gradient of 25 \(^\circ\text{C}/\text{km}\) across the soil profile and assumed the soil surface temperature of 25 \(^\circ\text{C}\) and ambient relative humidity \( RH \) of 30\% \((P_{\text{sur}} = RH \times P_s)\). We expressed the temperature dependency of saturated vapor pressure \( P_s \) and the diffusion coefficient \( D_0 \) using the empirical equations \( P_s = 610 \exp \left( \frac{17.27T}{T+237.3} \right) \) and \( D_0 = 2.23 \times 10^{-5} \left( \frac{T+273.13}{273.13} \right) \) where the units of \( P_s, D_0 \) and \( T \) are \( \text{Pa}, \text{m}^2\text{s}^{-1} \) and \( ^\circ\text{C} \), respectively (Ben Neriah et al., 2014).
We followed the methodology used in Kamai and Assouline [2018] to calculate $z_{EF}(m)$. Briefly speaking, an initial guess for $z_{EF}$ is used in equation (1) to calculate $q_v$. Since $q_v$ should be equal to $q_L$, the computed $q_v$ is substituted in equation (2) to calculate $z_{EF}(m)$. The obtained value of $z_{EF}(m)$ is then compared with the initial guess and the same procedure is repeated until the difference between these two values of $z_{EF}(m)$ is less than $2.2204 \times 10^{-16}$. At that point, the numerical iteration is halted and the corresponding value is considered as $z_{EF}(m)$.

Regardless of the type of soil and the evaporative demand, $z_{EF}(m)$ was assumed to be equal to zero in Selker [2017]. To evaluate this assumption, we used literature data covering a wide range of soil type with known van Genuchten parameters, porosity and saturated hydraulic conductivity and calculated the corresponding $z_{EF}(m)$ for each soil. The data used for the computation is presented in Table S1 of the Supporting Information section. Figure 2 shows the computed $z_{EF}(m)$ as a function of van Genuchten parameter ‘$n$’ which corresponds to the soil textural class.
Figure 2. The location of the evaporation front above the water table $z_{EF}$ (m) as a function of the soil textural classes represented by the van Genuchten parameter ‘n’. The legend indicates the type of soil and the numbers indicate the source of data as: 1: Carsel and Parrish [1988], 2: Tuller and Or [2001], 3: van Genuchten [1980] and Ghezzehei et al. [2007], 4: Yang and Yanful [2002], 5: Watson [1967], 6: Elrick and Bowman [1964], 7: Philip [1957], 8: Nemes et al. [1999], 9: Jesen and Hanks [1967], 10: Gowing et al. [2006].

Although in some cases $z_{EF}$ (m) could be small and negligible, Figure 2 clearly illustrates it is not always the case. Under the boundary conditions applied in the present calculation, depending on the type of soil, $z_{EF}$ (m) could be even more than 100 m. Therefore, ignoring this liquid region and its contribution to the evaporation process is an unwarranted assumption. We have calculated the corresponding evaporation rates given the calculated values of $z_{EF}$ and the resulting evaporative fluxes range from 0.012 to 0.059 mm/yr. Note that the values of $z_{EF}$ presented in Figure 2 with the calculated range of the evaporation fluxes correspond to the external conditions of $T_{sur} = 30$ °C and $RH = 30\%$ and these values would change if the computation were performed under different external conditions. For example, if $T_{sur}$ and RH were 50 °C and 10% respectively, the evaporative fluxes would range from 0.056 to 0.212 mm/yr (which is ~20 times more than the flux estimated in Selker [2017]).

Finally, although we acknowledge the usefulness of offering simple (yet effective) analytical calculation to describe water evaporation from deep aquifers, the true physics controlling the process should not be compromised for the sake of providing analytical estimation.
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References


