Determining In-situ Unsaturated Soil Hydraulic Conductivity at a Fine Depth Scale with Heat Pulse and Water Potential Sensors

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Keywords
In-situ unsaturated soil hydraulic conductivity, heat pulse, water potential gradient

Disciplines
Agricultural Science | Hydrology | Soil Science

Comments

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has potential to reveal the influences of natural soil conditions on hydraulic properties as they change with depth and time.

**Keywords:** In-situ unsaturated soil hydraulic conductivity, heat pulse, water potential gradient.

**Highlights:**

- A novel method is used to monitor in-situ unsaturated soil hydraulic conductivity.
- The method is based on heat pulse and water potential measurements.
- The method gives reasonable unsaturated soil hydraulic conductivity estimates.
1. Introduction

Unsaturated soil hydraulic conductivity \((K)\) plays a significant role in processes important for ecological, agricultural, and hydrological applications. In general, \(K\) is measured in the laboratory. Wind (1966) developed a method to quantify \(K\) from evaporation experiments on vertical soil columns. Multiple matric potential \((\psi)\) measurements at different depths were required in Wind’s method. Schindler (1980) simplified Wind’s method for determination of \(K\) by using \(\psi\) measurements at only two depths, while total column weight was recorded at several times. Schindler’s method is particularly attractive due to its simplicity, and commercial devices based on this method have been developed and widely used (Schindler et al., 2010; Schwen et al., 2014; Peters et al., 2015; Brunetti et al., 2017). The one-step (Kool et al., 1985) and multistep outflow methods (Hopmans et al., 2002) have also been used to determine \(K\) by numerical inversion of the Richards equation from controlled transient flow measurements. The numerical inversion method is relatively complicated because it requires an outflow experiment and a numerical computation. In the field, approaches have been developed to estimate \(K\) through infiltration experiments (Ankeny et al., 1991; Kosugi and Nakayama, 1997; Angulo-Jaramillo et al., 2000). The tension disc infiltrometer and pressure ring infiltrometer are two of the primary devices that can be used for determining \(K\) in the field (Angulo-Jaramillo et al., 2000). More commonly, \(K\) is estimated using measured soil water characteristic curves and saturated hydraulic conductivity \((K_s)\), based on relative hydraulic conductivity models (Burdine, 1953; Mualem, 1976; van Genuchten, 1980; Assouline, 2001). However, none of the mentioned approaches are capable of continuously tracking in-situ \(K\) dynamics under natural conditions.

Monitoring in-situ dynamics of soil physical properties at fine depth and temporal scales is challenging due to the lack of reliable measurement techniques. Recent advances have begun to
make these measurements more feasible. Heat pulse sensor and time domain reflectometry (TDR) arrays have been used to determine near surface soil water content (θ) at a depth resolution of about 1 cm (Sheng et al., 2017; Zhang et al., 2017). The thermo-TDR method can be used to monitor temporal variation of soil bulk density at the centimeter scale (Lu et al., 2017; Tian et al., 2018). The thermo-TDR method also has the capability to determine ice content at fine depth scale (Tian et al., 2016, 2017). The sensible heat balance method is capable of determining subsurface soil water evaporation (E) dynamics at a sub-cm scale (Zhang et al., 2012), and it also has been used to estimate soil freezing and thawing rates of near-surface layers (Kojima et al., 2014). Near surface soil also undergoes significant changes in K due to transient environmental conditions (Chen et al., 2014; Ghysels et al., 2018). Unfortunately, it remains difficult to determine real-time in-situ K variations at a fine depth.

The objective of this study is to present a novel method to estimate in-situ K dynamics at fine depth scale using heat pulse and matric potential sensors. The method derives K values at a fine depth by inverting the Buckingham-Darcy equation using ψ measurements and water flux densities estimated with a simple heat and water balance method. We conduct a field experiment to test the performance of this approach for monitoring in-situ K dynamics. Laboratory K measurements made on samples collected from the same site and Mualem-van Genuchten (1980) model predictions are used to evaluate the accuracy of the field values.

2. Materials and methods

2.1. Conceptual background

Latent heat flux density induced by evaporation of water within a soil layer can be estimated using the difference between sensible heat flux densities measured at upper and lower boundaries of the soil layer and the change in sensible heat storage within the soil layer (Fig. 1),
\[ LE = (H_u - H_l) - \Delta S_h \]  

where \( L \) (J m\(^{-3}\)) is the latent heat of vaporization; \( E \) is evaporation rate (m s\(^{-1}\)); \( H_u \) and \( H_l \) are soil sensible heat flux densities (W m\(^{-2}\)) at upper and lower boundaries of the soil layer, respectively; and \( \Delta S_h \) (W m\(^{-2}\)) is the change in soil sensible heat storage within the soil layer. Based on Eq. (1), Heitman et al. (2008a, 2008b) developed a method to estimate \( E \) within the soil profile using heat pulse sensors. \( H_u, H_l \), and \( \Delta S_h \) can be estimated with heat-pulse sensor measured soil temperature gradient, volumetric heat capacity \( (C) \), and thermal conductivity \( (\lambda) \). This method is referred to as the sensible heat balance method. For more details of the sensible heat balance method, please refer to Heitman et al. (2017).

For bare soil conditions without root extraction, the \( E \) in Fig. 1 can also be treated as the difference between liquid water flux densities (m s\(^{-1}\)) at upper and lower boundaries of the soil layer \( (q_u \) and \( q_l \), respectively) and the change in water storage within the soil layer, \( \Delta S_w \) (m s\(^{-1}\)),

\[ E = (q_u - q_l) - \Delta S_w \]  

(2)

The value for \( \Delta S_w \) can be estimated from the change of soil water content \( (\theta) \). Many approaches have been used to measure \( \theta \), among these the heat pulse sensor can determine \( \theta \) dynamics at a fine depth scale and near the soil surface using \( C \) measurements (Heitman et al., 2003; Zhang et al., 2017). Thus, \( \Delta S_w \) is derived from heat pulse sensor measured \( C \) values as

\[ \Delta S_w = \frac{(C_i - C_{i-1}) \Delta \tau}{C_w \Delta t} \]  

(3)

where \( C_i \) denotes the soil volumetric heat capacity (MJ m\(^{-3}\) K\(^{-1}\)) measured at time \( t' \), \( \Delta t = t' - t'^{-1} \) is the time step (s), \( C_w \) is the volumetric heat capacity of water (4.182 MJ m\(^{-3}\) K\(^{-1}\)), and \( \Delta z \) is the thickness of the soil layer (m).
For a soil profile, \( q_u \) at the soil surface is equal to 0 for the condition of no water input, while for the condition of a water input, it can be approximated as the rainfall or irrigation rate (m s\(^{-1}\)) when this rate is lower than the potential infiltration rate. Thus, \( q_l \) at a depth \( z \) (defined as positive downward, m) in the soil profile can be derived from \( q_u \) and heat pulse sensor estimated \( E \) and \( \Delta S_w \) between depths 0 and \( z \) (Eq. 2). If \( \psi \) (values for unsaturated soils are negative) in the soil profile is also measured, \( K \) at depth \( z \) can be estimated using the Buckingham–Darcy equation,

\[
K = -\frac{q_l}{(\frac{\psi_2 - \psi_1}{z_2 - z_1}) - 1}
\]

where \( \psi_1 \) and \( \psi_2 \) are \( \psi \) values (here, using water pressure head in m to unify the units) at depths \( z_1 \) and \( z_2 \), respectively. Depth \( z \) is the center of depths \( z_1 \) and \( z_2 \), i.e., \( z = (z_1 + z_2)/2 \). Note, when the \( z \)-axis is defined to be positive upward, the gravitational term (i.e., -1) in Eq. (4) changes to +1.

2.2. Field measurements

A field experiment was performed to assess the feasibility of the new approach to determine in-situ \( K \) values. The experimental site was located at the Central Crops Research Station, Clayton, NC, USA. The surface soil layer (0-20 cm) at the site has a texture of loamy sand (85% sand, 9% silt, and 6% clay). About 100 m\(^2\) area of land was leveled using a harrow, and it was maintained bare via weed control with an herbicide throughout the study. Instrumentation was done in April 2017.

The sensors were installed via a 15-cm deep trench by pushing the sensor needles or heads into the undisturbed soil. For \( E \) measurements, heat pulse sensors identical to those described by Tian et al. (2015), which consisted of three parallel stainless-steel needles (4.5-cm length, 2-mm
diameter, and 8-mm needle-to-needle spacing), were used in our study. Heat pulse sensors were installed at five depths with the central needles positioned at 0.8, 2.4, 4, 7.5, and 12.5 cm, respectively (Fig. 2). The plane formed by the needles of each heat pulse sensor was oriented perpendicular to the soil surface. Ambient soil temperature values were recorded each 15 min with thermocouples positioned in each heat pulse sensor needle. Heat pulse measurements were performed each 3 h to obtain $C$ and $\lambda$ dynamics. All of the measurements were controlled and recorded with a data logger (CR3000, Campbell Scientific, Logan, UT) that was placed in a waterproof enclosure on the soil surface. Under field conditions, heat pulse temperature signals measured in the near surface soil layer can be affected by ambient temperature fluctuations. To account for the effect of ambient temperature variations, a temperature drift correction following Zhang et al. (2014) was performed. After the correction, heat pulse data were then processed using the Lu et al. (2013) method to derive thermal properties ($C$ and $\lambda$). Temperature, $C$, and $\lambda$ measurements were used together to estimate hourly LE dynamics. Fig. 3 gives the details on LE computations at various depths. During the period immediately following a rainfall event, Stage I evaporation occurs at the surface and is therefore not detectable with heat pulse sensors below the surface (Sakai et al., 2011). Xiao et al. (2014) developed a modified sensible heat balance method to capture Stage I evaporation using combined net radiation and heat pulse measurements. In the present study the Xiao et al. (2014) method was used under such conditions.

In addition to determining $E$, the heat pulse sensors were also used to estimate hourly $\Delta S_w$ (mm h$^{-1}$) dynamics at different depths in the soil profile with Eq. (3). Fig. 3 presents the detailed calculation procedure for $\Delta S_w$. Ren et al. (2003) reported that relatively large errors can occur in heat pulse sensor measured $\theta$ due to errors in determining specific heat of the soil solids. Needle
deflection is another source of error for heat pulse sensor estimated water content. Wen et al. (2015) and Liu et al. (2016) present methods for correcting measurements for needle deflections. In order to verify the accuracy of heat pulse sensor measured $\Delta S_w$, $\theta$ was also determined with TDR sensors in the present study. Five 7.5-cm long TDR sensors (Model CS645-L, Campbell Scientific Inc., Logan, UT, USA) were inserted horizontally into the undisturbed soil at the same depths as the heat pulse sensors (Fig. 2). For TDR measurements, bulk soil dielectric permittivity was recorded hourly with a time-domain reflectometer (TDR100, Campbell Scientific, Logan, UT, USA), and $\theta$ was estimated from the dielectric permittivity using the empirical Topp et al. (1980) equation. By comparing with $\theta$ measurements from gravimetric sampling, Tian et al. (2018) reported that use of the Topp et al. (1980) equation provided accurate $\theta$ values for the loamy sand used in this study. Depth-averaged $\theta$ values from TDR were used to calculate hourly $\Delta S_w$ dynamics (see Fig. 3). The average of measurements from adjacent sensors was used as the value at the mid-depth between the sensor installation depths. Both heat pulse-based and TDR-based $\Delta S_w$ estimates along with sensible heat balance method estimated $E$ values using heat pulse sensors were used to calculate $q_l$ values (referred to as the heat pulse-based $q_l$ and the combined heat pulse and TDR-based $q_l$).

Several approaches have been developed for in-situ measurement of soil $\psi$. We conducted a preliminary experiment using micro-tensiometers for determination of soil $\psi$. However, the micro-tensiometers had limited measurement range and failed to work properly in this loamy sand under relatively dry conditions. Thus, in this study, soil $\psi$ values were measured with the MPS-6 water potential sensors (Decagon Devices, Inc., Pullman, WA, USA), which were more robust than the micro-tensiometers and extended the measurement range to air-dry conditions. For sensor installation, we moistened some native soil, packed it firmly around the sensor disc,
and inserted the packed sensor horizontally into a channel at the desired depth via a trench. MPS-6 sensors were installed at 1-, 3-, 5-, 10-, and 15-cm depths (Fig. 2). Accordingly, δψ gradients at depths of 2, 4, 7.5, and 12.5 cm were estimated from MPS-6 measurements. We then calculated the corresponding accumulated $E$ and $ΔS_w$ values for the 0 to 2, 0 to 4, 0 to 7.5, and 0 to 12.5 cm soil layers using the heat pulse-based method and the combined heat pulse and TDR-based method. At the same site, a weather station measured daily rainfall, net radiation, air temperature and humidity, atmospheric pressure, and wind speed. The $q_l$ values at depths of 2, 4, 7.5, and 12.5 cm were derived from $E$, $ΔS_w$, and rainfall rate. Subsequently, $K$ values at the same depths were estimated with Eq. (4). Fig. 3 gives the detailed spatial arrangements of all the measurements and computational procedures for parameters needed for estimating $K$ at various depths or for specific soil layers.

2.3. Validation of in-situ $K$ estimates

In order to verify the accuracy of the in-situ $K$ estimates, soil $K(ψ)$ curves were also determined in the laboratory. Intact soil cores (8-cm diameter by 5-cm long) from depths of 0 to 5 cm, 5 to 10 cm, and 10 to 15 cm were collected from the same site. In the laboratory, soil $K$ values were determined using a HYPROP device (UMS GmbH, Munich, Germany). During the measurement, the soil column-HYPROP device assembly was placed on a balance, $ψ$ values at two depths (1.25 and 3.75 cm) within the soil cores and weight loss were recorded continuously and automatically, and then $K$ values were calculated from the measurements using the HYPROP-fit software following the Schindler (1980) evaporation method. There were two replicates for each depth. Laboratory measured $K(ψ)$ curves were compared with in-situ estimates at depths of 2, 4, 7.5, and 12.5 cm.
It is difficult to quantify the accuracy of in-situ $K$ estimates directly using laboratory measured $K(\psi)$ curves because they were determined at various $\psi$ values and ranges. In this study, we fitted the Mualem-van Genuchten relative hydraulic conductivity model (van Genuchten, 1980) to the laboratory measured $K(\psi)$ curves, and then applied the model predictions for verifying the accuracy of the in-situ estimates. The Mualem-van Genuchten model is given as follows,

$$S_e = \left[\frac{1}{1+(\alpha|\psi|)^n}\right]^{1-1/n}$$  \hspace{1cm} (5)

$$K = K_s S_e^L \left[1 - \left(1 - S_e^{n/(n-1)}\right)^{1-1/n}\right]^2$$ \hspace{1cm} (6)

where $S_e$ is the effective degree of saturation; $\alpha$ and $n$ are empirical shape parameters for the water retention curve, and $L$ is an empirical pore-connectivity parameter. Commonly an $L$ of 0.5 is used (Mualem, 1976), but several studies have indicated that $L$ varied over a wide range and could be negative (Yates et al., 1992; Schaap & Leij, 2000). In this study, we measured $K_s$ of soil samples at 0- to 15-cm depth with a constant head method (Klute and Dirksen, 1986), and obtained $\alpha$, $n$, and $L$ parameters by fitting the model to the laboratory measured $K(\psi)$ curves of soil samples from 0- to 5-, 5- to 10-, and 10- to 15-cm depths. The goodness of fit was quantified with the root mean square error (RMSE) between fitted and measured $K$ values. The calculation of RMSE is given as follows,

$$\text{RMSE} = \sqrt{\frac{\sum_{i=1}^{N} (\log_{10} K_{\text{measured}} - \log_{10} K_{\text{fitted}})^2}{N}}$$ \hspace{1cm} (7)

where $N$ is the number of data points. Logarithmic values of $K$ were used in Eq. (7) because $K$ varied over several orders of magnitude.

Finally, we calculated the RMSE between the in-situ $K$ estimates and model predictions using best-fit parameters to evaluate the performance of the new field measurement-based approach.
3. Results and discussion

3.1. Components of heat and water balance and matric potential gradient

Fig. 4 presents an example of heat and water balance components and \( \psi \) measurements required for estimating \( K \) at the 2-cm depth during a 10-day period. Within these days, a rainfall of 10.5 mm occurred on April 12, and two smaller rainfalls (both were 0.3 mm) occurred on April 13 and 17, respectively (Fig. 4-c). For all three rain events, the intensity was much smaller than the \( K_s \) (53 mm h\(^{-1}\)), and thus was lower than the potential infiltration rate. Fig. 4-a gives the sensible heat balance method estimated hourly \( E \) rate in the 0- to 2-cm soil layer. For this near surface soil layer, \( E \) and net radiation showed consistent trends in their diurnal variations. The \( E \) estimates in the other layers were much lower than those in the 0- to 2-cm soil layer (data not shown). Previous studies made at the same field site showed that the estimated accumulated \( E \) by using the sensible heat balance method agreed well with measurements from micro-lysimeter and micro-Bowen ratio methods (Holland et al., 2013; Deol et al., 2014). We note that the sensible heat balance method ignores the heat convection from the liquid water flux at the drying front, because it is negligibly small compared to the sensible heat flux by conduction (Sakai et al., 2011). However, rainfall induced infiltration may have a larger effect, because the associated water fluxes are relatively large, thus \( E \) should be assumed to be 0 mm h\(^{-1}\) during rainfall when computing the water balance. In our study, the rainfall events mainly occurred during the nighttime when evaporation rates were very small (Fig. 4).

Figs. 4-b and 4-c present the hourly \( \Delta S_w \) dynamics in the 0- to 2-cm soil layer and cumulative \( \Delta S_w \) values (i.e., net change in water storage) over time from the heat pulse-based and TDR-based methods during the study period. Compared to the heat pulse-based method, the TDR-based method estimated \( \Delta S_w \) showed greater temporal fluctuations (Fig. 4-b). This might be
because hourly $\theta$ values measured with TDR were used to calculate $\Delta S_w$, but $\Delta S_w$ values estimated with heat pulse sensors were obtained using $C$ values measured each three hours. The cumulative $\Delta S_w$ estimates from both heat pulse-based and TDR-based methods were in good agreement with a mean bias (heat pulse minus TDR) of -0.07 mm within the 10-day period (Fig. 4-c). Heitman et al. (2003) reported that the heat pulse method was more appropriate for determining change in $\theta$ than for determining its absolute value, which was consistent with the approach used for the calculation of $\Delta S_w$ herein (i.e., change in water storage rather than absolute water storage). Thus, by comparison to TDR and from observations reported previously, the heat pulse method was shown to be capable of providing accurate $\Delta S_w$ estimates. The cumulative $\Delta S_w$ values estimated with the heat pulse-based method did, however, show a diurnal variation that was more evident than those estimated with the TDR-based method. The diurnal pattern in cumulative $\Delta S_w$ is consistent with the diurnal cycling in near-surface $\theta$ reported by others (Jackson 1978; McInnes et al., 1986; Cahill and Parlange, 1998; Heitman et al., 2008a). Diurnal variations in $\psi$ at the 1-cm depth were also observed in our study (Fig. 4-e). The difference between heat-pulse and TDR data might also result from the differences in measurement volumes: The three-rod TDR sensor has a larger sensing volume (about 50 cm$^3$) compared to the heat pulse sensor which has a sensing volume of about 20 cm$^3$ (Schwartz et al., 2013; Knight et al., 2007), thus, the TDR is not as likely to capture diurnal variation in $\theta$ at shallow depths.

The hourly $\Delta S_w$ estimates from both heat pulse-based and TDR-based methods were used to calculate $q_l$ at the 2-cm depth (Fig. 4-d). $q_l$ estimates using $\Delta S_w$ from the two methods matched closely with each other. Linear regression analysis indicated that a strong correlation existed between $q_l$ estimates from the two methods ($r^2 = 0.85$ and RMSE = 0.26 mm h$^{-1}$). On most days, negative $q_l$ values (upward flow of liquid water) were observed at the 2-cm depth, and the
magnitude of $q_l$ generally increased from mid-morning through mid-afternoon and then decreased from late-afternoon through mid-night. The diurnal pattern in $q_l$ was driven by evaporation. Most of the evaporation in the 0- to 2-cm soil layer occurred in the daytime which resulted in water moving upwards from below. Large positive values of $q_l$ (downward flow of liquid water) observed on April 12 were caused by infiltrating rainwater that moved past the 2-cm soil depth on that day. In general, the pattern and magnitude of the subsurface $q_l$ dynamics appeared reasonable and was consistent between the two methods used to estimate $\Delta S_w$.

Fig. 4-e shows the $\psi$ dynamics at depths of 1 and 3 cm within the 10-day period. The $\psi$ at the 1-cm depth showed a significant diurnal variation caused by the evaporation of water and condensation of water vapor. For the 3-cm depth, however, the diurnal variation only occurred on days with relatively dry conditions. The $\psi$ measurements at 1- and 3-cm depths were used to estimate $\psi$ gradients at the 2-cm depth. $K$ values at the 2-cm depth were derived from the $\psi$ gradients and $q_l$ estimates using Eq. (4).

3.2. $K$ estimates at different depths

Fig. 5 presents the estimated in situ $K$ dynamics at depths of 2, 4, 7.5, and 12.5 cm. The heat pulse-based and the combined heat pulse and TDR-based $K$ values agreed well with each other most of the time. The $\theta$ dynamics from the TDR method are also included in Fig. 5, in which $\theta$ values at the 2-cm depth were estimated from measurements made at depths of 0.8 and 2.4 cm using linear interpolation. Generally, $\theta$ increased with increasing soil depth and $K$ varied with depth and time following changes in $\theta$. At the 2-cm depth, $K$ varied over a range of $10^{-4}$ to 10 mm h$^{-1}$ due to significant changes in $\theta$. At the 4- and 7.5-cm depths, $K$ varied over a smaller range ($10^{-3}$ to 1 mm h$^{-1}$) than it did at the 2-cm depth. At the 12.5-cm depth, $K$ remained relatively constant compared to the upper depths during the study period.
Both heat pulse-based and combined heat pulse and TDR-based $K$ values showed diurnal variations at all four depths (Fig. 5). The diurnal pattern in $K$ seemed to conflict with $\psi$ and $\theta$ values, which did not show substantial diurnal variations at depth. This might be because of the time and depth approximations we used in the Buckingham-Darcy equation, which created an artifact that propagated downward from the diurnally-varying surface flux. The diurnal variations were more significant in the heat pulse-based $K$ estimates than in those from the combined heat pulse and TDR method, and this phenomenon was related to the difference in $q_l$ estimates (Fig. 4-d). As indicated above, the TDR sensor has a relatively large sensing volume for measuring $\theta$ compared to the heat pulse method, thus, it may not have the capability to capture fine-scale variations in $\theta$ at shallow depths. On the other hand, the TDR method might give more accurate $\Delta S_w$ estimates in the 5- to 15-cm soil layers than the heat pulse method, because for each method only two sensors were installed at these depths, and the heat pulse sensor has a relatively small sensing volume (about half the sensing volume of TDR sensor). In general, the trends observed in $K$ dynamics over depth and time from both methods appeared to be reasonable.

Fig. 6-a illustrates the relationship between $K$ from the heat pulse-based method and $\theta$ at various depths. The $K$-$\theta$ relationship seemed to be influenced by soil depth. For the same $K$, $\theta$ at 2- and 4-cm depths were lower than those at 7.5- and 12.5-cm, which was likely due to differences in soil bulk density (Fig. 6-b). Previous studies have reported that soil bulk density has a considerable impact on the soil water retention curve (Assouline, 2006; Zhang et al., 2018). From Fig. 6-c, we can see that the field measured water retention curve varied with soil depth as did the soil bulk density. For near surface soil with a smaller bulk density (0- to 5-cm depth), smaller $\theta$ values were observed at the same $\psi$ values compared to the soil at deeper depths which had larger bulk density (5- to 15-cm depth). This happened because the near surface soil with
smaller bulk density had larger pores, and thus, shallow soil held less water at relatively large \( \psi \) values than did the deeper soil. Since \( K \) values were directly estimated from \( \psi \) gradient, it had a similar dependence on depth/bulk density as did the soil water retention curve.

3.3. Evaluation of in-situ \( K \) estimates

Fig. 7 shows a comparison of estimated \( K(\psi) \) in-situ values and laboratory measured \( K(\psi) \) curves. \( K(\psi) \) curves measured with the HYPROP device for soil samples collected from 0- to 5-, 5- to 10-, and 10- to 15-cm soil layers were not exactly the same. They varied within a range in \( \log_{10} (K, \text{mm h}^{-1}) \) of \( \pm 0.5 \) of the average at the same \( \psi \) values. This happened because soil bulk density and pore system varied with soil depth. The in-situ estimated \( K(\psi) \) values had larger variation (mostly, in \( \log_{10} (K, \text{mm h}^{-1}) \) of \( \pm 1 \) of the average) than the laboratory measurements at the same \( \psi \) values. Wetting and drying cycles have a substantial effect on soil hydraulic properties, and hysteresis phenomena are most apparent in-situ (Zhang et al., 2018). In the present study, several wetting and drying cycles occurred in the field experiment, which might contribute to the relatively large variation in \( K(\psi) \) estimates as compared to laboratory measurements, which were subjected to only one drying cycle. Due to the measurement range limitation, the HYPROP device only gave \( K \) measurements at potentials greater than -100 kPa. The MPS-6 sensor measurement range can include very dry conditions (−100,000 kPa), which has the potential to extend the measurement range of \( K \) to these same relatively dry conditions.

Both heat pulse-based and combined heat pulse and TDR-based in-situ \( K(\psi) \) curves (see Figs. 7-a and 7-b, respectively) showed trends similar to the laboratory measurements. At \( \psi \) values \( \leq -20 \) kPa, in-situ estimated \( K \) values (mainly from 2- and 4-cm depths) from both methods agreed very well with the laboratory measurements. The in-situ estimated \( K \) values scattered randomly along the laboratory measured \( K(\psi) \) curves within a \( \pm 0.5 \) (value of \( \log_{10} (K, \text{mm h}^{-1}) \)) range. At \( \psi \)
values > -20 kPa, however, some of the in-situ estimated \( K \) values were much larger than the laboratory measurements. At large \( \psi \) values close to 0, the absolute accuracy of \( \psi \) measurements have a greater impact on \( K \) estimates compared to smaller \( \psi \) ranges when using both our method and the HYPROP device. The HYPROP device determined \( \psi \) using micro-tensiometers, while the in-situ approach measured \( \psi \) with MPS-6 sensors. The two sensors are based on different physical principles. The micro-tensiometers measure \( \psi \) with an accuracy of about ±0.5 kPa, while the MPS-6 sensors have errors about ±2 kPa in the wet range (Decagon Devices, 2017). Thus, the deviations between in-situ estimated and laboratory measured \( K \) values at \( \psi \) values > -20 kPa stem in part from the less accurate \( \psi \) measurements by MPS-6 sensors.

We fitted the Mualem-van Genuchten model to laboratory measured \( K(\psi) \) curves to obtain best-fit model parameters (Fig. 8), then used model predictions to quantify the accuracy of the in-situ \( K \) estimates. Fig. 9 compares the heat pulse-based and the combined heat pulse and TDR-based in-situ \( K \) estimates to \( K \) values predicted by the Mualem-van Genuchten model. The \( K \) values estimated with the heat pulse-based method were in good agreement with model predictions with an average RMSE (in \( \log_{10} (K, \text{mm h}^{-1}) \)) of 0.57 and an average bias of 0.17. The combined heat pulse and TDR-based method provided \( K \) estimates that were slightly greater than the model predictions, with an average RMSE in \( \log_{10} (K, \text{mm h}^{-1}) \) of 0.54 and an average bias of 0.40. On the whole, both methods gave reasonable in-situ \( K \) estimates compared to model predictions. The combined heat pulse and TDR-based method is more costly than the heat pulse-based method, however, because in addition to TDR measurements, it also still requires heat pulse measurements for estimating \( E \). Overall, our results indicate that combining the heat pulse technique (with or without supplemental TDR measurements) with MPS-6 sensors provides a promising approach for in-situ estimation of \( K \) at a fine depth scale.
4. Conclusion

Soil $K$ is a function of soil $\theta$ or $\psi$. Under field conditions, $K(\theta)$ and $K(\psi)$ relationships vary in time and space due to changes in soil bulk density, pore size distribution, and many other transient environmental conditions. Few measurement approaches are currently available for determining in-situ $K$ variations. Combining heat pulse and $\psi$ gradient measurements provides a means to determine in-situ $K$ dynamics at a fine depth scale. We performed a field experiment to test the feasibility of this new approach for estimating $K$ in-situ. The results showed that the new approach provided reasonable in-situ $K$ estimates at fine depth scale which were in good agreement with those measured in the laboratory using a HYPROP device and those predicted using the van Genuchten–Mualem model. The field experiment showed that the estimated $K(\theta)$ relationship varied with depth due to variations in bulk density. Thus, the proposed approach is a promising method to estimate $K$ in-situ at a fine depth scale, and it is capable of determining transient $K$ dynamics that are influenced by variable environmental conditions. In this study, the new approach was only tested in a loamy sand soil during a short period with simple weather condition. Further studies on soils with various textures and under changing atmospheric conditions are needed in the future.
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References


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Fig. 1. Conceptual model of the heat and water balance for a soil layer (see Eqs. 1 and 3). $H_u$ and $H_l$ are sensible heat flux densities at upper and lower boundaries, respectively; $q_u$ and $q_l$ are liquid water flux densities at upper and lower boundaries, respectively; $\Delta S_h$ and $\Delta S_w$ are changes in sensible heat storage and water storage in the soil layer, respectively; $L$ is the latent heat of vaporization; and $E$ is evaporation rate.
Fig. 2. Arrangements of heat pulse, TDR, and MPS-6 water potential sensors in a vertical soil profile. Dimensions are not drawn to scale.
Fig. 3. A sketch depicting the spatial arrangements of soil temperature ($T$), volumetric heat capacity ($C$), and thermal conductivity ($\lambda$) measured with the heat-pulse sensors, water content ($\theta$) measured with TDR sensors, and matric potential ($\psi$) measured with MPS-6 sensors in the soil profile. Also shown are how the measurements are applied to calculate sensible heat flux densities ($H$), changes in...
sensible heat storage ($\Delta S_h$), latent heat flux densities ($LE$), changes in water storage ($\Delta S_w$), evaporate rate ($E$), liquid water flux density ($q$), and hydraulic conductivity ($K$) at various soil depths or in different soil layers. The subscripts of soil parameters indicate soil depths or soil layers (mm).
Fig. 4. Measured and estimated components required for estimating unsaturated soil hydraulic conductivity at the 2-cm depth within the study period: (a) soil evaporation ($E$) estimates in the 0- to 2-cm layer and measured net radiation, (b) heat pulse-based and TDR-based water storage change ($\Delta S_w$) estimates in the 0- to 2-cm layer, (c) cumulative $\Delta S_w$ values over time in the 0- to 2-cm layer and rainfall rate, (d) liquid water flux densities ($q_l$) at the 2-cm depth obtained by using heat pulse-based and TDR-based $\Delta S_w$ values, and (e) matric potential ($\psi$) measurements at depths of 1 and 3 cm.
Fig. 5. Unsaturated soil hydraulic conductivity ($K$) estimates and water content ($\theta$) measurements at different soil depths during the study period. Triangles and circles represent $K$ values estimated using the heat pulse-based and the combined heat pulse and TDR-based methods, respectively. Each point represents an average of $K$ estimates over six hours.
Fig. 6. (a) The relationship between unsaturated soil hydraulic conductivity ($K$) estimates and water contents ($\theta$) at depths of 2, 4, 7.5, and 12.5 cm, (b) soil bulk densities ($\rho_b$) in the 0- to 2.5-, 2.5- to 5-, 5- to 10-, and 10- to 15-cm soil layers and (c) The relationship between matric potentials ($\psi$) and water contents ($\theta$) at depths of 2, 4, 7.5, and 12.5 cm.
Fig. 7. The relationship between unsaturated soil hydraulic conductivity ($K$) and soil matric potential ($\psi$). Lines are laboratory measured $K(\psi)$ curves made on samples collected from 0- to 5-, 5- to 10-, and 10- to 15-cm layers. The symbols are in-situ $K$ estimates at depths of 2, 4, 7.5, and 12.5 cm obtained using the heat pulse-based (a) and combined heat pulse and TDR-based (b) methods.
Fig. 8. Laboratory measured unsaturated soil hydraulic conductivity ($K$)-matric potential ($\psi$) curves from 0- to 5-, 5- to 10-, and 10- to 15-cm soil layers and fitted $K$-$\psi$ curves using the Mualem-van Genuchten model. $\alpha$, $n$, and $L$ are best-fit model parameters.
Fig. 9. The heat pulse-based (a) and the combined heat pulse and TDR-based (a) in-situ unsaturated soil hydraulic conductivity ($K$) estimates at depths of 2, 4, 7.5, and 12.5 cm versus Mualem-van Genuchten model estimated $K$ values.
Highlights:

- A novel method is used to monitor in-situ unsaturated soil hydraulic conductivity.
- The method is based on heat pulse and water potential measurements.
- The method gives reasonable unsaturated soil hydraulic conductivity estimates.