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SOIL PHYSICS

An Improved Model for Predicting Soil Thermal Conductivity from Water Content at Room Temperature

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The thermal conductivity and water content relationship is required for quantitative study of heat and water transfer processes in saturated and unsaturated soils. In this study, we developed an improved model that describes the relationship between thermal conductivity and volumetric water content of soils. With our new model, soil thermal conductivity can be estimated using soil bulk density, sand (or quartz) fraction, and water content. The new model was first calibrated using measured thermal conductivity from eight soils. As a first step in validation, predicted thermal conductivity with the calibrated model was compared with measured thermal conductivity on four additional soils. Except for the sand, the root mean square error (RMSE) of the new model ranged from 0.040 to 0.079 W m^{-1} K^{-1}, considerably less than that of the Johansen model (0.073–0.203 W m^{-1} K^{-1}) or the Côté and Konrad model (0.100–0.174 W m^{-1} K^{-1}). A second validation test was performed by comparing the three models with literature data that were mostly used by Johansen and Côté and Konrad to establish their models. The RMSEs of the new model, the Johansen model, and the Côté and Konrad model were 0.176, 0.176, and 0.177 W m^{-1} K^{-1}, respectively. The results show that the new model provided accurate approximations of soil thermal conductivity for a wide range of soils. All of the models tested demonstrated sensitivity to the quartz fraction of coarse-textured soils.

The thermal conductivity of soil (λ) is an important heat transfer property that is connected to the ability of soil to conduct heat (Bristow, 2002). Soil thermal conductivity has been widely used in energy engineering, soil science, and agricultural meteorology. Among the many factors that influence λ, e.g., soil water content (θ), mineral composition, temperature, bulk density (ρ_b), and porosity (n), θ is the one that varies the most under field conditions. In situ monitoring of λ as a function of θ can represent a significant challenge. Consequently, much effort has been made to develop $\lambda(\theta)$ models based on easily measurable soil parameters (Kersten, 1949; de Vries, 1963; Johansen, 1975; Campbell, 1985; Côté and Konrad, 2005).

The available soil $\lambda(\theta)$ models can be divided into two groups: semitheoretical models and empirical models. The de Vries (1963) model is physically based and has been applied in numerous studies, including predicting the response of soil thermal conductivity to temperature change (Campbell et al., 1994). This model requires many input parameters (Bachmann et al., 2001; Tarnawski and Wagner, 1992), and proper parameter selection, such as critical water content and shape factors, is necessary to predict λ accurately (Horton and Wierenga, 1984; Ochsner et al., 2001). On the basis of a large number of laboratory measurements, Kersten (1949) developed an empirical $\lambda(\theta)$ model. This model only requires bulk density, ρ_b, as an input parameter, but it is not suitable for predicting λ at lower water contents. Campbell (1985) introduced an empirical function to describe the $\lambda(\theta)$ relationship. The empirical function has five parameters, some of which are difficult to obtain. Johansen (1975) proposed the concept of normalized thermal conductivity, and established a simple empirical $\lambda(\theta)$ model that is based on the degree of saturation (S_r) and soil mineral composition. For many soils, the Johansen (1975) model provides accurate predictions of λ (Farouki, 1981, 1982; Tarnawski and Wagner, 1992). Recently, Côté and Konrad (2005) developed an improved model based on Johansen's method by introducing an empirical relationship between the normalized thermal conductivity and S_r. Our independent tests of the Côté and Konrad (2005) correction indicate that it does not always perform well at lower water contents, especially on fine-textured soils. Thus, there is a need for a thermal conductivity model that is applicable at lower soil water contents.

The objective of this study was to develop a simple model that can describe the room-temperature $\lambda(\theta)$ relationship for the entire water content range. The performance of the new model, along with the Johansen (1975) model and Côté and Konrad (2005) model, was evaluated by comparing predicted thermal conductivity with measured thermal conductivity.

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MATERIALS AND METHODS

Soil Thermal Conductivity Measurement

Twelve soils, 10 from China (Soils 1–9 and 11) and two from Iowa (Soils 10 and 12) were used in this study. The basic soil characteristics are listed in Table 1. Soil particle-size distribution was determined by the pipette method (Gee and Or, 2002), and soil organic matter content was measured by the Walkely–Black titration method (Nelson and Sommers, 1982).

Soil samples were air dried, ground, and sieved through a 2-mm screen. Various soil water contents were obtained by adding a known amount of water to the samples and mixing thoroughly. The moist soil was then packed into a column (50.2-mm inner diameter and 50.2 mm high) to a desired bulk density. The packed soil samples were sealed and placed in a constant temperature room (20 ± 1°C) for 24 h before measurements to ensure thermal equilibration between water and soil.

A thermo-time domain reflectometry (thermo-TDR) probe (Ren et al., 1999) was used to measure the thermal properties of the packed soil columns. The probe is able to simultaneously measure θ with the TDR technique and thermal properties with the heat-pulse method. Details on the thermo-TDR method are presented in Ren et al. (1999), Ochsner et al. (2001), and Ren et al. (2003).

To determine λ, a thermo-TDR probe was inserted carefully into each soil column, and current was applied to the heater in the middle needle for 15 s to produce a heat pulse. The magnitude of the applied current was selected to cause a maximum temperature increase in the outer thermo-TDR rods of 0.7 to 0.8°C. A data logger (CR 23X, Campbell Scientific, Logan, UT) was used to record the power input and temperature change in the outer rods with 1-s intervals. Heat-pulse measurement was repeated three times on each soil column. Soil thermal properties (thermal conductivity, thermal diffusivity, and volumetric thermal capacity) were determined with a nonlinear regression technique (Welch et al., 1996) involving the temperature increase vs. time in the outer rods. Following thermo-TDR measurements, soil columns were oven dried at 105°C to determine ρ_b and θ gravimetrically.

We also conducted heat-pulse measurements on oven-dried samples to determine the thermal conductivity of dry soil materials (oven dried at 105°C).

Model Development

Johansen Model

Johansen (1975) proposed the concept of normalized thermal conductivity, K_e (i.e., the Kersten number), and established a relationship between λ and K_e for unsaturated soils, based on λ values at dry and saturated states:

$$\lambda = \left(\lambda_{sat} - \lambda_{dry}\right)K_e + \lambda_{dry} \qquad [1]$$

where λ_{dry} and λ_{sat} are the thermal conductivity of dry and saturated soils (W m^{-1} K^{-1}), respectively.

Johansen (1975) related the Kersten number to the normalized soil water content, or S_r ($S_r = \theta/\theta_s$, where θ_s is the saturated water content):

$$K_e \simeq 0.7\log S_r + 1.0 \quad (S_r > 0.05) \qquad [2]$$

for coarse-textured soils, and

$$K_e \simeq \log S_r + 1.0 \quad (S_r > 0.1) \qquad [3]$$

for fine-textured soils.

A geometric mean equation based on the thermal conductivity of water (λ_w = 0.594 W m^{-1} K^{-1} at 20°C) and effective thermal conductivity of soil solids (λ_s), was used to estimate λ_{sat}:

$$\lambda_{sat} = \lambda_s^{1-n}\lambda_w^{n} \qquad [4]$$

where n is soil porosity. The value of λ_s was determined using another geometric mean equation from the quartz content of the total solids content (q) and thermal conductivities of quartz (λ_q = 7.7 W m^{-1} K^{-1}) and other minerals (λ_o):

Table 1. Texture, particle-size distribution, organic matter (OM) content, and bulk density (ρ_b) of the soils.

Soil no.	Texture	Sand	Silt	Clay	OM	ρ_b
		%				g cm^{-3}
1	sand	94	1	5	0.09	1.60
2	sand	93	1	6	0.07	1.60
3	sandy loam	67	21	12	0.86	1.39
4	loam	40	49	11	0.49	1.30
5	silt loam	27	51	22	1.19	1.33
6	silt loam	11	70	19	0.84	1.31
7	silty clay loam	19	54	27	0.39	1.30
8	silty clay loam	8	60	32	3.02	1.30
9	clay loam	32	38	30	0.27	1.29
10	silt loam	2	73	25	4.40	1.20
11	loam	50	41	9	0.25	1.38
12	sand	92	7	1	0.60	1.58

$$\lambda_s = \lambda_q^{q}\lambda_o^{1-q} \qquad [5]$$

where λ_o was taken as 2.0 W m^{-1} K^{-1} for soils with $q > 0.2$, and 3.0 W m^{-1} K^{-1} for soils with $q \leq 0.2$.

Johansen (1975) used a semiempirical relationship to obtain λ_{dry}:

$$\lambda_{dry} = \frac{0.135\rho_b + 64.7}{2700 - 0.947\rho_b} \qquad [6]$$

where ρ_b was the bulk density of soil (kg m^{-3}) and 2700 was the density of soil solids (kg m^{-3}).

Côté and Konrad Model

To eliminate the logarithmic function in the model of Johansen (1975), Côté and Konrad (2005) related soil K_e to S_r using a soil texture dependent parameter k:

$$K_e = \frac{kS_r}{1+(k-1)S_r} \qquad [7]$$

The k values used by Côté and Konrad (2005) were 4.60, 3.55, 1.90, and 0.60 for gravel and coarse sand, medium and fine sand, silty and clayey soils, and organic fibrous soils, respectively.

Furthermore, Côté and Konrad (2005) proposed a new empirical equation to represent the dependence of λ_{dry} on soil porosity (n),

$$\lambda_{dry} = \chi 10^{-\eta n} \qquad [8]$$

where χ (W m^{-1} K^{-1}) and η were parameters accounting for particle shape effects. The χ and η values given by Côté and Konrad (2005) were 1.70 W^{-1} K^{-1} and 1.80 for crushed rocks, 0.75 W m^{-1} K^{-1} and 1.20 for mineral soils, and 0.30 W m^{-1} K^{-1} and 0.87 for organic fibrous soils. Côté and Konrad (2005) also applied Eq. [4] to obtain λ_{sat}. However, they calculated λ_s in a slightly different way. Instead of using quartz and other minerals (Johansen, 1975), λ_s was calculated by using the geometric mean method based on the thermal conductivity and volumetric proportion of rock-forming minerals. Therefore, a complete soil mineral composition was required.

The New Model

At room temperature, the relationship between λ and θ can be explained by the interactions between solid particles and water. At lower θ values, water molecules are tightly adsorbed to particle surfaces. In this domain, the thickness of water films grows with increasing θ, and λ

increases gradually but not substantially. With further θ increases, water bridges begin to form between soil solid particles, and λ starts to increase rapidly because of the improved contact between the particles (Sepaskhah and Boersma, 1979; Tarnawski and Gori, 2002). This process continues until most of the solid particles are connected together. Thereafter, the increase in λ with increasing θ depends largely on the displacement of air by water, and as a result, λ increase becomes slower.

Although Côté and Konrad (2005) improved the Johansen the dependence o (1975) method, their model did not always adequately account for f K_e on S_r, especially on fine-textured soils and at lower θ. In this study, we proposed the following equation relating K_e to S_r across the entire range of soil water content:

$$K_e = \exp\{\alpha[1 - S_r^{(\alpha-1.33)}]\} \qquad [9]$$

where α is a soil texture dependent parameter, and 1.33 is a shape parameter.

Furthermore, we introduced a simple linear function that describes the relationship between λ_{dry} and n for mineral soils, because the magnitude of heat transfer in dry soils is related to soil porosity as:

$$\lambda_{dry} = -an + b \qquad [10]$$

where a and b are empirical parameters.

To calculate λ_{sat}, we followed the same procedure as Johansen (1975, Eq. [4]). We did not measure the quartz contents of our soils, so we assumed that the quartz content was equal to the sand content (Peters-Lidard et al., 1998).

Model Calibration and Testing

We used measured thermal conductivity values from eight soils (Soils 1–8) to calibrate the new model. The soil texture dependent parameter α was estimated by fitting Eq. [9] to heat-pulse measured λ and θ data on the soils. We divided the eight soils into two groups and determined a specific α value for each group. The parameters a and b were obtained by fitting Eq. [10] to heat-pulse measurements on dry soil samples. We also used λ_{dry} data from Johansen (1975) for rock, gravel, and sand.

The new model was tested and compared against the Johansen model and the Côté and Konrad model using λ and θ measurements of four additional soils (Soils 9–12). The RMSE and the bias in model prediction were calculated for the three models:

$$\mathrm{RMSE} = \sqrt{\frac{\sum\left(\lambda_m - \lambda_p\right)^2}{m}} \qquad [11]$$

$$\mathrm{bias} = \frac{\sum(\lambda_m - \lambda_p)}{m} \qquad [12]$$

where m was the number of measurements, and λ_m and λ_p represented the measured and predicted values, respectively.

Finally, the new model was evaluated using data from Kersten (1949), Johansen (1975), and Farouki (1982).

RESULTS AND DISCUSSION

Analysis of the Thermal Conductivity (Water Content) Curve

Figure 1 presents the results of soil λ as a function of θ for four soils of different textures. In Fig. 1a, heat-pulse measured λ data along with standard deviations are plotted against θ values determined gravimetrically. In Fig. 1b, the normalized λ data are plotted against S_r, or the normalized θ values. Two noticeable characteristics can be observed about the influence of soil texture on the shape of the λ(θ) curve. First, the coarse-textured soils (Soils 1 and 2) have higher λ values than the fine-textured soils (Soils 5 and 8). The greatest λ values appear in the soil that has the largest sand fraction (Fig. 1a). A second feature is that, at lower water contents, the λ increase with θ is more gradual on the fine-textured soils than on the coarse-textured soils. The θ at which a sharp λ increase begins is greater for the fine-textured soils than for the coarse-textured soils (Fig. 1a). This can be explained by the fact that fine-textured soils have larger surface areas and more water is required before water bridges are formed between solid particles.

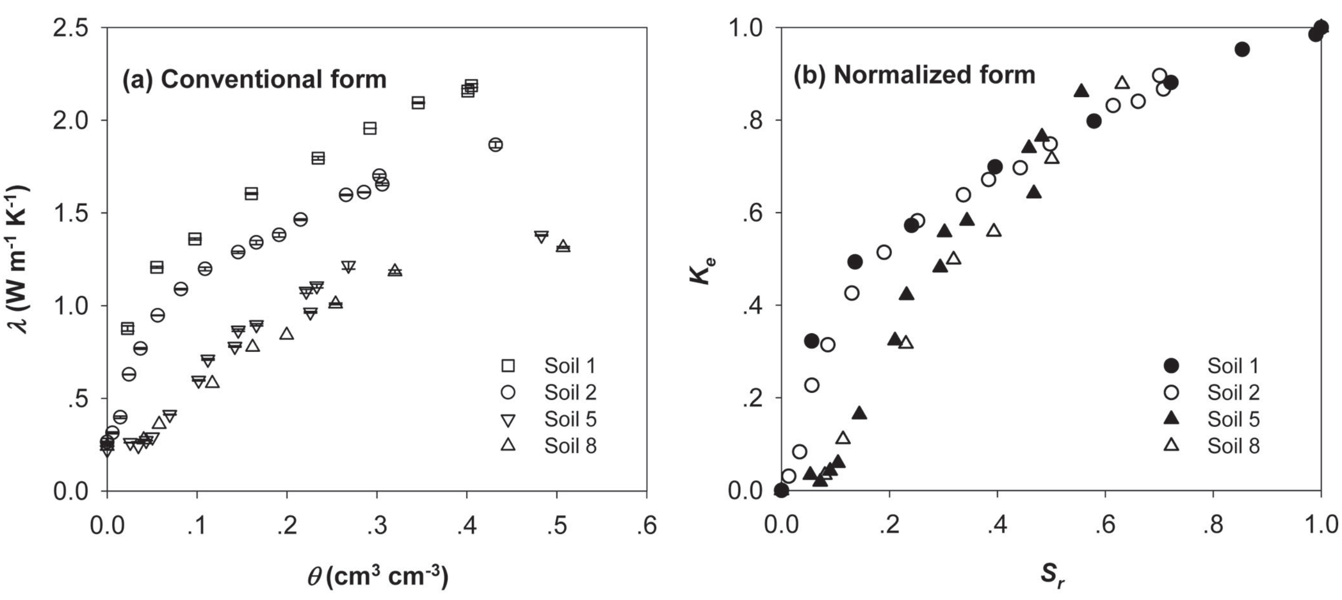

Fig. 1. Thermal conductivity as a function of water content as indicated by (a) the conventional thermal conductivity (λ) and water content (θ) relationship and (b) the normalized form (the Kersten number K_e vs. the degree of saturation S_r) for selected soils.

When the conventional $\lambda(\theta)$ curves are transformed to the normalized form, the K_e–S_r curves separate into distinct texture groups (Fig. 1b). Johansen (1975) first noticed this phenomenon and used two different equations to calculate K_e from S_r for fine-textured and coarse-textured soils (see Eq. [2] and [3]). Côté & Konrad (2005) treated this by setting different k values for gravel and coarse sand, medium and fine sand, silty and clayey soils, and organic fibrous soils (see Eq. [7]). Neither Johansen (1975) nor Côté and Konrad (2005) defined quantitative boundaries between different soil groups, however.

Our investigation indicates that the eight soils from this study can be divided into two groups: coarse-textured soils with sand fractions >0.40 (Soils 1–4), and fine-textured soils with sand fractions <0.40 (Soils 5–8). The fine-textured soils are characterized by a three-region K_e–S_r curve, with S_r = 0.13 and 0.30 as the dividing points (Fig. 1b). Within each region, K_e generally increases linearly with S_r, and the magnitude of the slopes follow the order of Region 1 (S_r < 0.13) < Region 3 (S_r > 0.30) < Region 2 (0.13 < S_r < 0.30). A two-region K_e–S_r curve is typical for the coarse-textured soils, where S_r = 0.13 is the separation point (Fig. 1b). For these soils, the slope of the K_e–S_r curve of Region 1 is greater than that of Region 2.

Determination of Soil Texture Dependent Parameter α

The parameter α for the two soil groups was determined by fitting Eq. [9] to the heat-pulse measurements. The fitted values of α are 0.96 and 0.27 for the coarse-textured and fine-textured soils, respectively. Figure 2 shows the measured K_e–S_r values (symbols) and the fitted curves (Eq. [9]). In general, the fitted curves well represent K_e as a function of S_r, and there is no significant deviation between the measured and fitted values. Therefore, Eq. [9] is able to represent the general trend of the K_e–S_r relationship for the entire soil water content range.

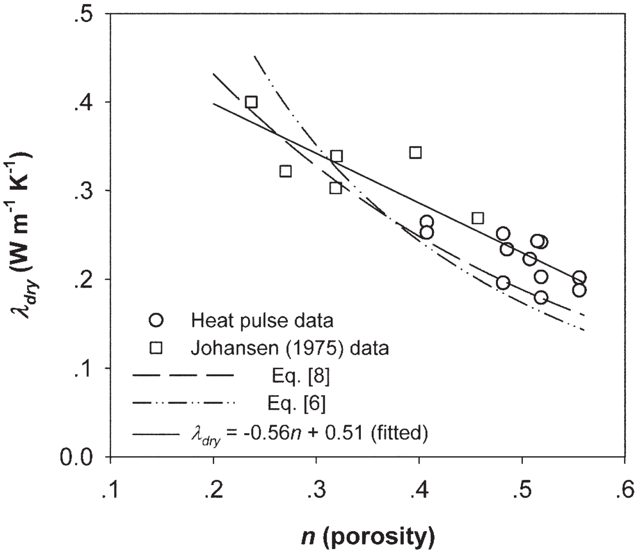

Fig. 3. The dependence of thermal conductivity of dry soils (λ_{dry}) on soil porosity (n): measurements and estimations from Eq. [6] and Eq. [8] for mineral soils. The fitted linear equation was used in the new model to predict λ_{dry} from n.

Determination of Empirical Parameters *a* and *b*

The thermal conductivity of dry soil depends on soil mineral composition, particle shape, and the volume ratio of air to solids in the soil (de Vries, 1963). Currently, thermal conductivity measurements on dry soils are limited and a general model is in demand to predict λ_{dry} from easily obtained parameters such as soil texture and porosity. Figure 3 shows our heat-pulse measured λ_{dry} results and λ_{dry} data from Johansen (1975) as a function of soil porosity. The predicted values using Eq. [6] and [8] for the mineral soils are also included. Clearly λ_{dry} decreased linearly with increasing n, and both Eq. [6] and [8] underpredicted λ_{dry} for the mineral soils from this study. Therefore, we applied a simple linear equation to

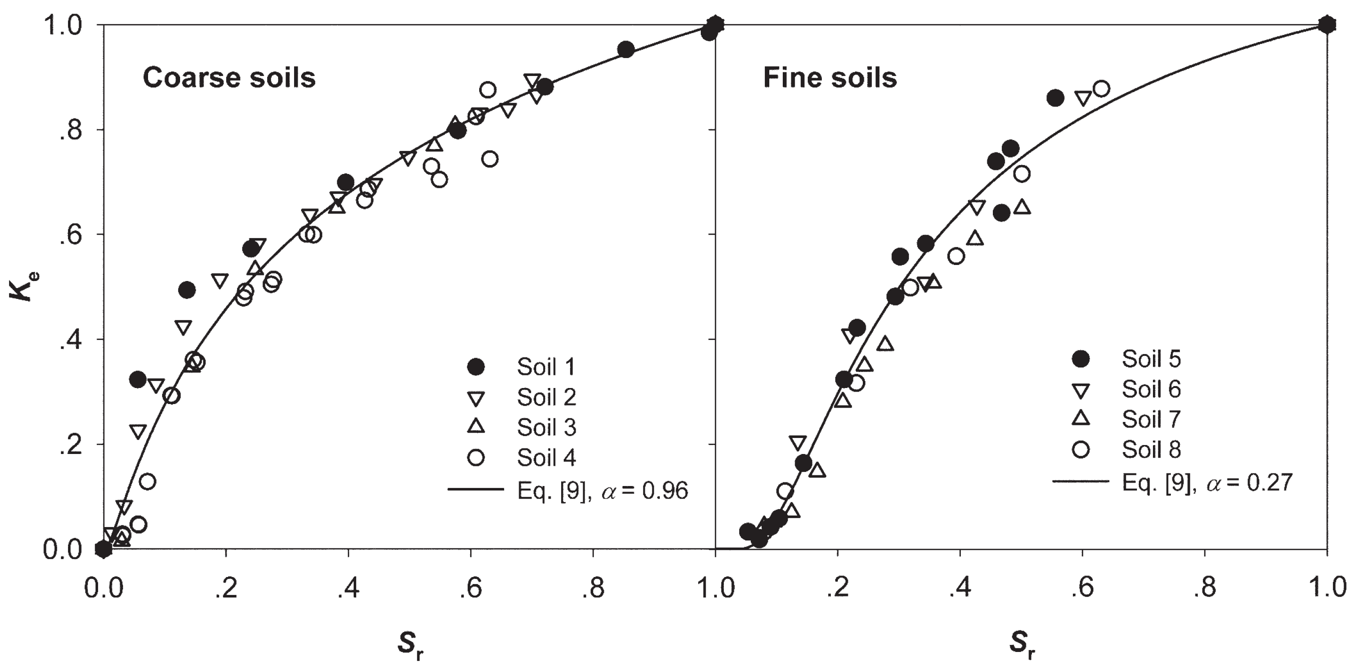

Fig. 2. The measured and fitted Kersten number (K_e) and degree of saturation (S_r) relationships for (a) coarse-textured soils with sand fraction >0.40 and (b) fine-textured soils with sand fraction <0.40.

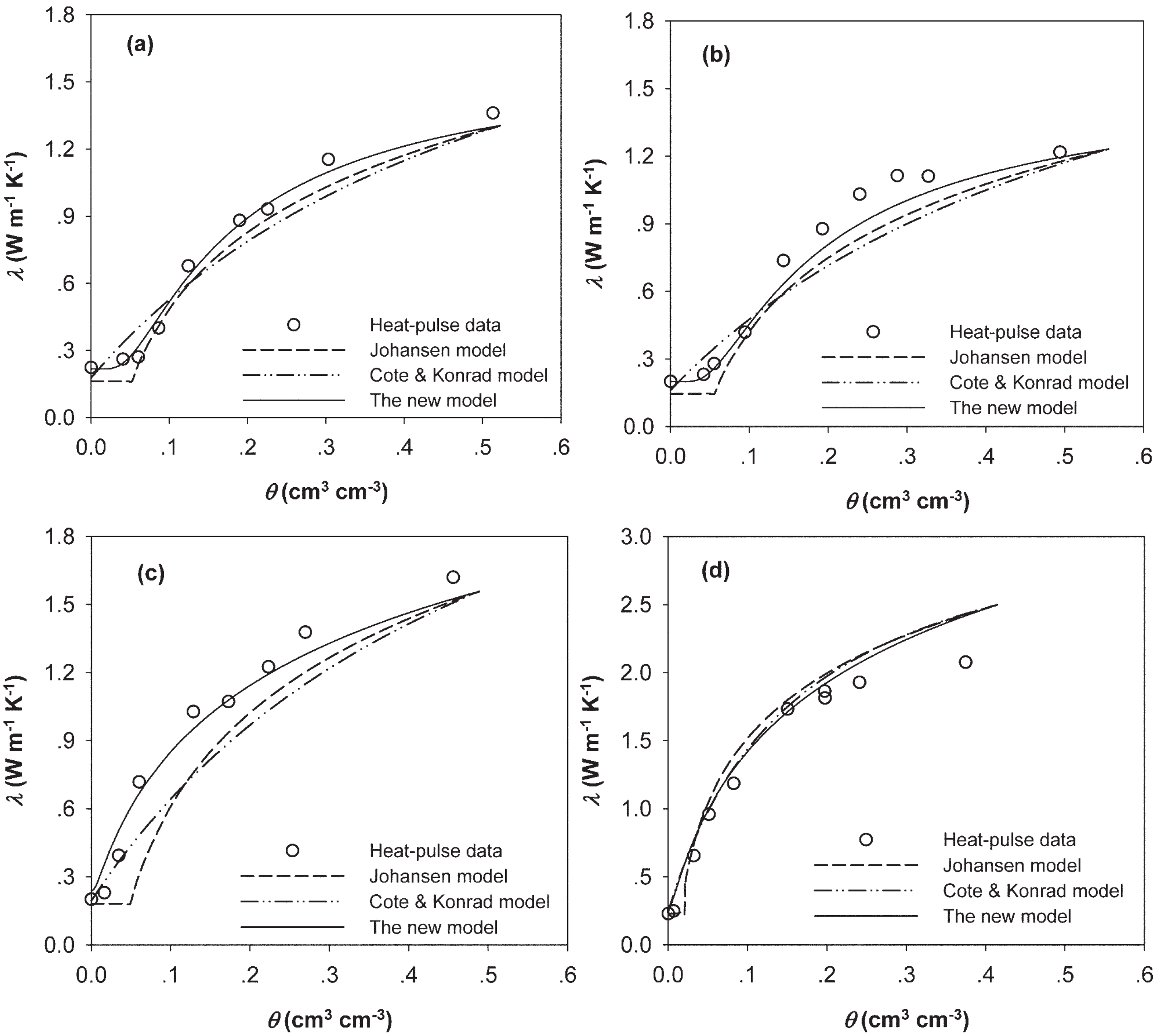

Fig. 4. Comparison of the Johansen (1975) model, the Côté and Konrad (2005) model, and the new model in predicting the soil thermal conductivity (λ) and water content (θ) relationship on (a) the clay loam soil (Soil 9), (b) the silt loam soil (Soil 10), (c) the loam soil (Soil 11) and (d) the sand (Soil 12).

calculate λ_{dry} in this study. By fitting Eq. [10] to the measured data in Fig. 3, the calculated values of a and b were 0.56 and 0.51 (for $0.2 < n < 0.6$), respectively.

Model Evaluation Using Heat-Pulse Measurements

We first compared the performance of the new model against the Johansen model and the Côté and Konrad model using heat-pulse measured λ data on four soils, two from China (Soils 9 and 11) and the two from the USA (Soils 10 and 12). As discussed above, the input parameters for the new model were λ_{dry}, λ_{sat}, and S_r. For each soil, λ_{dry} was calculated using Eq. [10] (where $a = 0.56$ and $b = 0.51$) from soil porosity. The value for λ_{sat} was determined using Eq. [4] and [5] from soil porosity and texture information where the quartz fraction was taken as the sand content. The particle densities of the soils were taken as 2.70 g cm^{-3} to calculate n and S_r.

Figure 4 presents the calculated λ(θ) data from the three models and the measurements from the heat-pulse method. The RMSE of the model predictions is listed in Table 2. Except for the sand, the new model was able to provide accurate predictions in the entire water content range, while the Johansen model and the Côté and Konrad model produced relatively large errors (Fig. 4a-c). The RMSE of the new model ranges from 0.040 to 0.079 W m^{-1} K^{-1}, significantly lower than that of the other two models (Table 2). The three models agree well on the sand (RMSE ranges from 0.138 to 0.164 W m^{-1} K^{-1};

Table 2. Root mean square error of the Johansen (1975) model, the Côté and Konrad (2005) model, and the new model in predicting thermal conductivity of four soils.

Soil no.	Johansen Model	Côté and Konrad Model	New Model
9 (clay loam)	0.073	0.100	0.040
10 (silt loam)	0.129	0.145	0.077
11 (loam)	0.203	0.174	0.079
12 (sand)	0.164	0.153	0.138

Fig. 4d). At water contents >0.20 $m^3\ m^{-3}$, the predicted results from all the models tended to be larger than the measured data, which will be explained below. These results suggest that the new model is applicable to a wide texture range of mineral soils, while the Johansen model and Côté and Konrad model are more suitable for coarse-textured soils.

Further examination of the data indicates that on the fine-textured soils, the Côté and Konrad model generally overpredicted λ at lower water contents and underpredicted λ at higher water contents, and the Johansen model underpredicted λ for the entire θ range (Fig. 4a–4b). For the silt loam soil, for example, the mean value of bias for the Côté and Konrad model is −0.037 $W\ m^{-1}\ K^{-1}$ at $\theta < 0.1\ m^3\ m^{-3}$ and 0.170 $W\ m^{-1}\ K^{-1}$ at $\theta > 0.1\ m^3\ m^{-3}$; the bias of the Johansen model averages 0.074 $W\ m^{-1}\ K^{-1}$ at $\theta < 0.1\ m^3\ m^{-3}$ and 0.142 $W\ m^{-1}\ K^{-1}$ at $\theta > 0.1\ m^3\ m^{-3}$ (Table 3). This tendency is not observed for the new model. Therefore, we conclude that the new model can describe soil thermal conductivity as a function of water content better than the Johansen model or the Côté and Konrad model.

In Eq. [7], Côté and Konrad (2005) recommended a k value of 3.55 for medium and fine sand and 1.90 for silty and clayey soils, but they did not quantitatively define the boundaries between sand, silty soil, and clayey soil, which might lead to improper selection of k and thereby erroneous λ results. For the loam soil (Soil 11), for example, the Côté and Konrad model produced a RMSE of 0.174 $W\ m^{-1}\ K^{-1}$ when a k value of 1.90 was used. If k was set at 3.55, the RMSE of prediction of the model was reduced to 0.071 $W\ m^{-1}\ K^{-1}$, similar to the RMSE of prediction from the new model. This demonstrates that the Côté and Konrad model is sensitive to the selection of parameter k.

For the sand, the large discrepancies between the measured and predicted λ values at $\theta > 0.2\ m^3\ m^{-3}$ are not clear. We hypothesize that it results from taking the sand content as the quartz fraction in calculating λ_{sat}. As indicated by Eq. [5], a small error in quartz content can cause a substantial error in λ_s, which is then transferred to λ_{sat}. This is partially supported by Fig. 5, where the predicted λ_{sat} values are compared to heat-pulse measurements on the 12 soils. When the quartz fraction is taken as the sand content from soil texture analysis, Eq. [4] significantly overpredicts λ_{sat} for the three sands, but the calculated values agree well with the measured data for the other soils. Similar observations were also reported by Bristow (1998), who stated that the accuracies of the common thermal conductivity models (e.g., de Vries, 1963; Campbell, 1985) were influenced by the information on the quartz fraction of the soil materials. Bristow (1998) demonstrated that substitution of sand content for quartz fraction could result in overprediction of soil thermal conductivity. Therefore, to predict λ_{sat} (and therefore λ) accurately, it is important to know the quartz content of soil samples.

Model Evaluation Using Literature Data

The new model was also tested using 12 data sets from Kersten (1949), Johansen (1975), and Farouki (1982). Actual quartz content measurements reported in the literature were used in the calculation. The results are shown in Fig. 6a to 6c. It is evident that the predicted λ values of the new model agree well with the λ measurements on all 12 media, as indicated by the random distribution of the data along the 1:1 line. Linear regression analysis with the pooled data produced an equation having an intercept of 0.13, a slope of 0.96, and an R^2 of 0.937, similar to the results from the Johansen (1975) and Côté and Konrad (2005) models. The RMSE of the new model is 0.176 $W\ m^{-1}\ K^{-1}$, similar to the Johansen (1975) model (0.176 $W\ m^{-1}\ K^{-1}$) and Côté and Konrad (2005) model (0.177 $W\ m^{-1}\ K^{-1}$). It is important to point out that the development of the Johansen (1975) and Côté and Konrad (2005) models was dependent on these data sets, while the development of the new model was independent of these data sets. Therefore, we conclude that the new model is capable of providing accurate λ for a wide range of soil textures.

Table 3. Bias of the Johansen (1975) model, the Côté and Konrad (2005) model, and the new model in predicting soil thermal conductivity of four soils at specified water content (θ) range.

Soil no.	θ	Johansen Model	Côté and Konrad Model	New Model
	$m^3\ m^{-3}$			
9 (clay loam)	<0.10	0.046	−0.061	−0.018
	>0.10	0.077	0.101	0.031
10 (silt loam)	<0.10	0.074	−0.037	0.005
	>0.10	0.142	0.170	0.091
11 (loam)	<0.04	0.095	−0.005	−0.090
	>0.04	0.206	0.206	0.058
12 (sand)	<0.01	0.007	−0.064	−0.076
	>0.01	−0.162	−0.133	−0.111

CONCLUSIONS

In this study, we developed an improved model for predicting soil thermal conductivity from its water content. A new functional relationship between the normalized thermal conductivity (i.e., K_e) and S_r was established for both coarse-textured and fine-textured soils. A simple linear relationship was applied to calculate the λ_{dry}

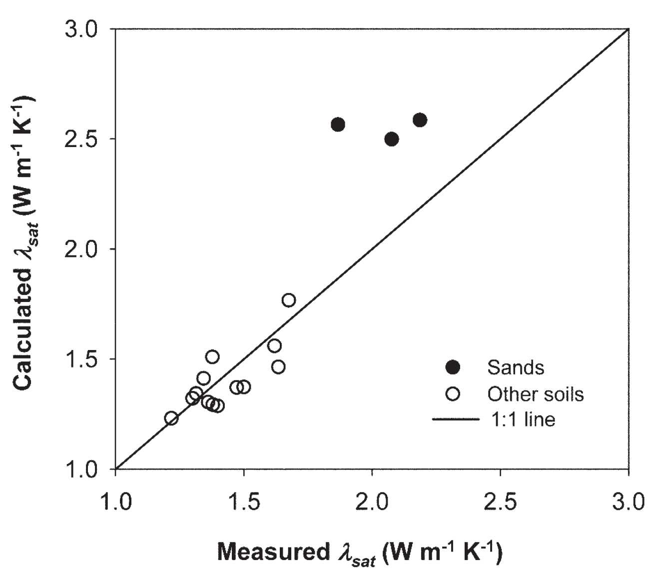

Fig. 5. Predicted saturated soil thermal conductivity (λ_{sat}) values from Eq. [4] vs. measured values on saturated soils. Quartz content is taken as the fraction of sand.

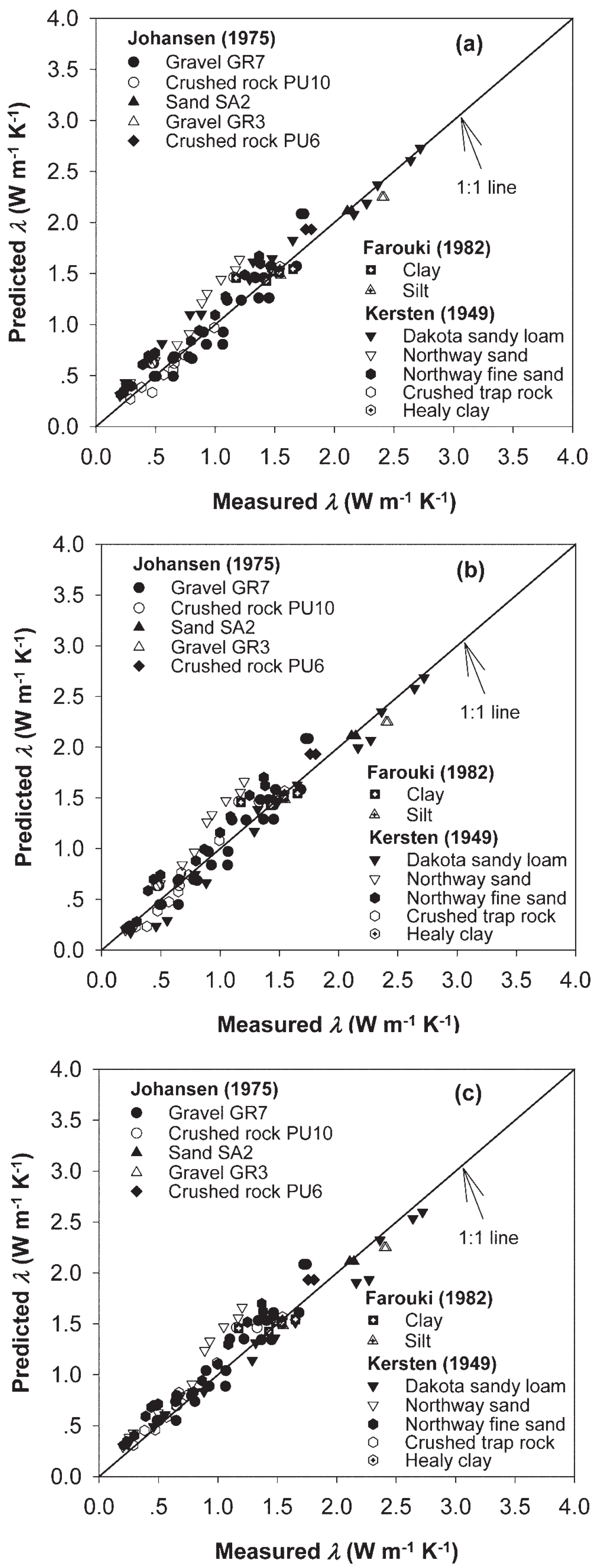

Fig. 6. Comparison of predicted soil thermal conductivity (λ) values from (a) the new model, (b) the Johansen (1975) model, and (c) the Côté and Konrad (2005) model vs. measured data from Kersten (1949), Johansen (1975), and Farouki (1982).

from soil porosity. For mineral soils, only three parameters, ρ_b, the sand (or quartz) fraction, and θ were required for the new model.

The new model was able to closely describe λ across the entire water content range for soils of various textures, and the RMSEs of λ estimates were lower than those of the other models tested. For sand, however, the new model showed a relatively high sensitivity to quartz fraction, especially at water contents >0.2 m^3 m^{-3}.

Evaluations using literature data showed that the predicted values from the new model agreed well with measured data on soils covering a wide range of textures.

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