Experimental Study on the Mechanisms of Soil Water-Solute-Heat Transport and Nutrient Loss Control

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Abstract

The release and migration of nutrients, pesticides, and other chemicals in the runoff from agricultural lands is not only an economic loss but a threat to the quality of our surface and groundwater. In contrast to pollution from point sources, pollution from non-point sources is often low in intensity but high in volume. The development of a physically based model to simulate the transport of soil solutes would provide a better understanding of transport mechanisms and assist in the development of effective methods to control the loss of nutrients from soils and the pollution of waterways. As a result, numerous studies have been conducted in this area. But due to the soil genesis and human activity, the process is very complex, which can have a great impact on soil water movement, solute transport, as well as nutrient loss. In this study, we determined water movement and solute and heat transport through columns of disturbed soil samples. We also carried out simulated rainfall experiments on an artificial slope to study the nutrient loss.

Keywords: water movement, soil heat transport, nutrient loss

1. Introduction

Due to the fast development of agriculture and industry, water resource scarcity and nutrient loss became more and more serious in China. In order to solve those problems, it is necessary to present some high-efficiency technical methods and theories to understand the whole process of water-solute-heat transport and nutrient loss.

Water infiltration process is a complex process, and it is necessary to know the rules of water movement well and establish formulas and models to describe the whole process. Darcy’s law
was presented in 1856 [1]. In 1907, Edgar Buckingham applied “capillary potential” to soil water for the first time, which showed the energy state of soil water. Green and Ampt proposed an infiltration model based on capillary theory [2]. Richards introduced Darcy’s law to describe soil water flow [3]. Philip presented a basic equation to describe the water movement in a one-dimensional vertical soil column [4]. In addition, the formula for infiltration of the Kostiakov infiltration model, Horton infiltration model, and Holtan infiltration model is also used [5–7]. Shu puts forward the model of capillary bundle infiltration [8]. Ghosh combines the one-dimensional vertical infiltration formula and Kostiakov empirical formula to obtain the new infiltration formula [9]. Many experts and scholars have proposed some new methods because it is difficult to find out the results in accordance with the actual results. Parlange presented an approximate solution for Richards equation [10], and Hogarth and Parlange improved the solution [11]. The finite difference method and finite element method were also used to solve the solution of the water equation [12]. Yang and Lei established a numerical model for the one-dimensional saturated water flow in the FORTRAN language and verified it in laboratory [13]. There are a large number of basic formulas and empirical formulas to describe the process of one-dimensional soil water movement. However, both the classical infiltration model and empirical model have different parameters which make them difficult and time-consuming. In this study, we want to find a simple and feasible method to determine soil hydraulic parameters.

Soil thermal conductivity is not only one of the important indexes of soil thermal properties but also an important parameter for simulating the soil hydro-thermal-solute-coupled model. How to estimate soil thermal conductivity quickly and accurately is one of main contents of studying soil thermal properties [14]. At present, a number of indirect estimation models to describe the relationship between thermal conductivity and soil texture, bulk density, water content, and organic matter were proposed by domestic and foreign scholars [15–23]. There are two types of indirect estimation models: empirical models [15, 16] and semi-theoretical models [17–20]. The empirical model mainly established the relationship between thermal conductivity and soil moisture content, such as the Chung-Horton model and Campbell model. These models are simple to calculate, but the model parameters are uncertain which will lead to large errors between the calculated data and measured values due to the difference of soil qualities in different regions [21]. The semi-theoretical model showed the relationship between thermal conductivity and soil saturation, such as the Johansen model, Côté-Konrad model, and Lu-Ren model [22]. These models have a theoretical basis and have given the model parameters for different soil textures. However, the model parameter values varied greatly with different soil particles and organic matter content, which limited the application of this model. In general, different models have their own advantages and disadvantages, but the effect of particle composition and organic matter content on the parameters of different types of soils needs further study. In this chapter, the thermal conductivity of undisturbed soil was measured by heat pulse methods. So, in this study, a new method based on analyzing the influence of soil particle composition on thermal conductivity, the relationship between thermal conductivity, saturation, bulk density, soil particle composition, and organic matter, was established. The improved Côté-Konrad and Lu-Ren models were also proposed to provide a reference method for obtaining soil thermal conductivity in a simple and rapid manner.
The process of soil-dissolved chemical transfer to the runoff and transport to the field outlet was complex. Modeling the large number of processes involved and their interactions requires the solution of relatively complicated, coupled linear and nonlinear partial differential equations subject to time-dependent boundary conditions [24]. To reduce mathematical complexity, we applied the refined model [25] to data from our experiments in this study, in which the presumed exchange layer is replaced by a mixing zone, which can be regarded as an extension of the deposited layer or “shield” [26]. Assuming that the exchange rate was controlled by raindrop splash and that the effects of diffusion could be neglected, we replaced the exchange rate, $k_m$, with the variable $e_r$, the raindrop-induced water transfer rate, developed by Gao et al. [27]. This modification obviated the need to calibrate $k_m$. Laboratory experiments were performed to assess the accuracy of the new model’s predictions. So, in this study, we carried out simulated rainfall experiments on an artificial slope to study the nutrient loss and test our new theory.

2. Materials and methods

2.1. Experimental soils

In this study, four soils were collected, and the soil’s physical characteristics were listed in Table 1.

2.2. Experimental measurement

2.2.1. Horizontal infiltration experiment

Four soils in Table 1 were collected for the infiltration experiments, and the negative hydraulic heads were designed as $-2.5$, $-6$, $-9$, $-12$, $-15$, and $-18$ cm. Soil samples were filled in the column, every 10 cm, and the bulk density of soil was designed as 1.4 g/cm$^3$. The length of $h_2$ and $h_3$ were measured, and the values of $h_1$ were calculated by the formula $p = h_3 - h_1 + h_2$. The

| Soil textural | <0.002 mm | 0.002 mm < $d$ < 0.02 mm | >0.02 mm | Saturated hydraulic conductivity (cm/min) | Initial water content (cm$^3$/cm$^3$) | Saturated water content (cm$^3$/cm$^3$) |
|---------------|-----------|--------------------------|----------|------------------------------------------|----------------------------------------|----------------------------------------|
| Loessal soil  | 2.7       | 12.96                    | 84.34    | 0.0416                                   | 0.01                                   | 0.47                                   |
| Red glue soil | 14.89     | 38.88                    | 47.66    | 0.0155                                   | 0.04                                   | 0.63                                   |
| Dark loessial soil | 20.82     | 41.49                    | 35.6     | 0.0047                                   | 0.04                                   | 0.54                                   |
| Lou soil      | 16.65     | 44.76                    | 38.22    | 0.0058                                   | 0.04                                   | 0.53                                   |

Table 1. The soil’s physical characteristics.
water head in pressure regulator pipe values was adjusted by the values of \( h_1 \) (Figure 1). The standpipe was filled with distilled water before the experiment. Opening the right valve, the experiment continued until the bubble was emptied. For each infiltration measurement, cumulative infiltration was recorded every minute until it reached a steady state.

2.2.2. Soil thermal experiment

The test equipment uses a three-probe heat pulse probe (Figure 2) which was connected to the data collector, and sensor probes were used on two sides to observe and monitor the changing

![Figure 1. The sketch map of experimental equipment for horizontal soil column.](image1)

![Figure 2. Schematic diagram of the heat pulse probe.](image2)
temperature in the process with time after the middle probe sent the heat pulse (Figure 3). The diameter, length, and space distance of the three probes were 1.3, 40, and 6 mm, respectively (as shown in Figure 1). The 5–6 gL\(^{-1}\) agar solution was used to demarcate in advance in actual, which was to prevent natural convection of water when heated. The Data Collector (US CR1000 Data Collector) controls the heated input via a relay, and the electric current was determined by a precise resistance (10 Ω) of the assigned voltage. The data collector also recorded the temperature change of the sensing probe at intervals of 1 s. The volumetric heat capacity of the agar solution is 4.18MJm\(^{-3}\)C\(^{-1}\). The distance \(r\) was obtained by a non-linear fitting temperature–time curve and averaged by repeating the calibration process 10 times.

To study the variation characteristics of the undisturbed soil thermal conductivity, the ring knife was used to take samples in the experimental ground. Each measurement point was arranged as follows: 10 measuring points per column, step length of 3 m between 2 points, and setting in 2 columns; 4 kinds of water contents were given to measure soil thermal conductivity in each measurement point, and the actual water contents were determined by the measured value at the end of the measurement (that means that the actual moisture content of the soil sample is supposed to be equal to that in the ring knife after finishing the measurement).

2.2.3. Nutrient runoff experiment

The basic component of the experiments was a rain simulator, which could generate a variable intensity of rainfall. The nozzles used to simulate rainfall were 15 m from the soil surface. We used six steel soil flumes with the following dimensions: 1 m in length × 0.40 m in width × 0.50 m in height. The flumes were filled with soil to a depth of 0.35 m; this depth allowed

![Figure 3. Schematic diagram of experimental apparatus.](http://dx.doi.org/10.5772/intechopen.76280)
infiltration without causing the bottoms of the flumes to become dank and left a 0.15 cm “lip” above the soil level to prevent water losses from splashing. The flumes’ angle of inclination could be varied between $0^\circ$ and $30^\circ$ (Figure 4). The experiments were performed from April to September 2010 in a laboratory for simulating artificial rainfall at the Institute for Soil and Water Conservation, Chinese Academy of Sciences, Shaanxi Province, China.

Three treatments were established to test our model. In treatment 1, three initial levels of soil moisture (5, 10, and 20%, measured gravimetrically) were used to study the influence of the soil’s initial water content on our model. The rainfall rate was 90 mm/h and the slope gradient was $5^\circ$. Treatment 2 was designed to investigate the influence of variation in the rainfall intensity on our model. Three different rainfall intensities (60, 96, and 129 mm/h) were examined, with an initial soil moisture content of 10% and a slope gradient of $5^\circ$. Treatment 3 was designed to assess the influence of the slope gradient. Slopes of 5, 15, and $25^\circ$ were investigated, with a rainfall intensity of 90 mm/h and an initial (gravimetric) soil moisture content of 10%. All treatments were run three times.

The soil samples were sieved (0.004 m in aperture) to remove coarse rock and debris and then air dried (to about 2%, gravimetrically). Potassium, used as a tracer, was dissolved in water and added to the test soils based on their designed soil water contents and potassium concentrations; the soil was then thoroughly mixed. The soil flume was filled with the prepared soil sample layer by layer to achieve a dry bulk density of 1.35 g/cm$^3$. To obtain a flat surface, a sharp-edged straight blade was used to remove excess soil. The soil surface was covered with plastic for approximately 24 h before the beginning of the experiments. During the experiments, the outflow from one of the holes in the flume was collected into plastic containers every minute to measure the amount of runoff and its sediment and soil concentrations. We directly measured the depth of the exchange layers along a vertical section. The potassium content in the runoff was measured with an atomic absorption spectrophotometer (Perkin-Elmer 5100ZL). The soil water content was measured by drying, and the sediment was isolated by filtration on filter paper and weighed after drying.

Figure 4. Experimental setup of artificial rainfall.
2.3. Theory

2.3.1. Models of infiltration process

2.3.1.1. Kostiakov model

The Kostiakov model [5] was presented by large amount of experiments and can be expressed as:

\[ I = at^b \]  

(1)

where \( I \) is cumulative infiltration (cm), \( t \) is the infiltration time (min), \( a \) and \( b \) are empirical constants. The empirical constants \( a \) and \( b \) have no physical meaning and are obtained by experimental data fitting.

2.3.1.2. Philip model

Philip [4] proposed an infiltration equation based on Boltzmann transformation as power series:

\[ I = St^{0.5} + At + Bt^{1.5} + \ldots \]  

(2)

where \( S \) is the soil sorptivity (cm/min^{1/2}) and \( A \) is stable infiltration rate (cm/min). The result is exactly enough with two terms.

A horizontal one-dimensional infiltration equation, neglecting gravity action, can be expressed as:

\[ I = St^{0.5} \]  

(3)

A vertical one-dimensional infiltration equation, neglecting gravity action, can be expressed as:

\[ I = St^{0.5} + At \]  

(4)

2.3.1.3. Wang’s model

Wang et al. [28] use \( kh(hd/h)^n \) and \((\theta_0 - \theta_r)/(\theta_s - \theta_r) = (hd/h)^n \) to calculate unsaturated hydraulic conductivity and soil moisture characteristic curves. The relationship between wetting front and cumulative infiltration time can be expressed as:

\[ I = x_f(\theta_s - \theta_i) \left(1 - \frac{n}{m + n - 1}\right) \]  

(5)

\[ q = \frac{1}{x_f m - 1} \]  

(6)

\[ x_f = \sqrt{\frac{2hdk_s(m + n - 1)}{(m^2 - 1)(\theta_s - \theta_i)^2}} \]  

(7)
Eq. (5), (6), and (7) also can be expressed as:

\[ I = A_1 x_f, \quad i = A_2 / x_f, \quad x_f = A_3 t^{1/2} \]  \hspace{1cm} (8)

\[ A_1 = \frac{k_s h_d^n h_s^{1-m}}{m - 1} \]
\[ A_2 = (\theta_s - \theta_i) \left( 1 - \frac{nh_d^{-n-1}h_s^{m-1}}{m + n - 1} \right) \]  \hspace{1cm} (9)
\[ A_3 = \frac{2k_s h_d^n h_s^{1-m} (m + n - 1)}{(m - 1)(\theta_s - \theta)(m + n - 1 - nh_d^{-n-1}h_s^{m-1})} \]

Therefore, parameters \( n, m, \) and \( h_d \) can be expressed as:

\[ n = \sqrt[\frac{\theta_s - \theta_r}{A_1 + \theta_i - \theta_r} - 1} \]
\[ h_d = \frac{A_2}{anKs} \]
\[ m = \frac{A_3(\theta_s - \theta_r)}{aKs h_d} + 1 + \frac{n}{n} \]  \hspace{1cm} (10)

where \( \theta_i \) is the soil water content (cm\(^3\)/cm\(^3\)), \( \theta_s \) is the saturated soil water content (cm\(^3\)/cm\(^3\)), \( \theta_r \) is the residual water content (cm\(^3\)/cm\(^3\)), \( h_d \) is air entry suction (cm), \( K_s \) is saturated hydraulic conductivity (cm/min), and \( n, m \) is the parameter.

### 2.3.1.4. Green-Ampt model

Green and Ampt [2] proposed the model. The equation is expressed as:

\[ i = K_s \left( 1 + \frac{h_0 + h_f}{z_f} \right) \]  \hspace{1cm} (11)
\[ I = (\theta_s - \theta_i)z_f \]  \hspace{1cm} (12)

where \( i \) is infiltration rate (cm/min), \( K_s \) is saturate hydraulic conductivity (cm/min), \( h_0 \) is ponder depth (cm), \( h_f \) is wetting front suction (cm), \( I \) is cumulative infiltration (cm), and \( \theta_s, \theta_i \) are saturated water content (cm\(^3\)/cm\(^3\)) and initial water content (cm\(^3\)/cm\(^3\)), respectively.

### 2.3.2. Models of soil heat conductivity

1. Thermal conductivity empirical model by Campbell [16].

Campbell proposed an empirical formula for calculating soil thermal conductivity based on soil texture, bulk density, and volume moisture content, which can be specifically expressed as

\[ \lambda = A + B\theta - (A - D)\exp \left[ -(C\theta)^E \right] \]  \hspace{1cm} (13)
where $\theta$ is the volume of water content (cm$^3$/cm$^3$), and the parameters $A$, $B$, $C$, $D$, and $E$ can be calculated according to soil bulk density, clay content, quartz, and other mineral volume ratios as follows:

$$
\begin{align*}
A &= 0.65 - 0.78\rho_b + 0.60\rho_b^2 \\
B &= 1.06\rho_b \\
C &= 1 + \frac{2.6}{m_c} \\
D &= 0.03 + 0.1\rho_b^2 \\
E &= 4
\end{align*}
$$

(14)

where $m_c$ is the clay content and $\rho_b$ is the soil bulk density.

2. Semi-theoretical model of thermal conductivity by Johansen [18].

For the unsaturated soil, the relationship between $\lambda$ and $K_e$ (Kersten) is established based on thermal conductivity $\lambda_{dry}$ (W/(m·K)) of dry soil and thermal conductivity $\lambda_{sat}$ (W/(m·K)) of saturated soil.

$$
\lambda = (\lambda_{sat} - \lambda_{dry})K_e + \lambda_{dry}
$$

(15)

And the relationship between $K_e$ and conventional soil moisture content or saturation $S_r$ ($S_r = \theta/\theta_s$, where $\theta_s$ is saturated water content) is established:

$$
K_e = \begin{cases} 
0.7\log S_r + 1.0 & \text{if } 0.05 < S_r \leq 0.1 \\
\log S_r + 1.0 & \text{if } S_r > 0.1
\end{cases}
$$

(16)

$$
\lambda_{sat} = \lambda_s^{1-n}\lambda_w^n
$$

(17)

where $\lambda_w = 0.594$ W/(m·K) under the condition 20°C, $n$ is the soil porosity, and $\lambda_s$ is obtained by the quartz content ($q$) of the whole solid, its thermal conductivity being $\lambda_q = 7.7$ W/(m·K), and thermal conductivity ($\lambda_0$) of the other minerals is $\lambda_s = \lambda_q^{1-q}\lambda_0^{1-q}$, where, $\lambda_0 = 2.0$ W/(m·K) ($q > 0.2$), $\lambda_0 = 3.0$ W/(m·K) ($q \leq 0.2$).

$$
\lambda_{dry} = \frac{0.135\rho_b + 64.7}{2700 - 0.947\rho_b}
$$

(18)

3. Improved Johansen model by Côté and Konrad [19].

In order to simplify the calculation of the logarithmic function formula in the Johansen model, Côté and Konrad proposed a new relationship between $K_e$ and $S_r$ based on the parameter $k$:

$$
K_e = \frac{kS_r}{1 + (k - 1)S_r}
$$

(19)

where $k$ is an independent parameter related to the soil texture and its values for coarse sand, small sand, clay, and higher organic matter content are 4.60, 3.25, 1.40, and 1.20, respectively.

And a new formula to estimate $\lambda_{dry}$ is given as follows:

$$
\lambda_{dry} = \chi 10^{-\eta n}
$$

(20)
where $\chi$ (W/(m·K)) and $\eta$ are parameters that are affected by particle traits. The $\chi$ and $\eta$ values for crushed rock, mineral soil, and soil with high organic matter were 1.70 and 1.80, 0.75 and 1.2, and 0.30 and 0.87, respectively.

4. Improved Johansen model by Lu and Ren [20].

In order to make the Johansen model more suitable for calculating the thermal conductivity under the condition of low soil water content, Lu and Ren proposed a new exponential function expression of $K_e$ about $S_r$:

$$K_e = \exp\left\{ a \left[ 1 - S_r^{\alpha - 1.33} \right] \right\}$$

(21)

where $\alpha$ is the parameter determined by the soil texture, and for coarse soils with sand content larger than 40% and fine soils with sand content of less than 40%, $\alpha = 0.96$, 0.27. 1.33 refers to the shape parameters. A new formula is given for mineral soil as follows:

$$\lambda_{dry} = -an + b$$

(22)

where $a$ and $b$ are the empirical coefficients; when $0.2 < n < 0.6$, the value is: 0.56, 0.51.

2.3.3. Models of nutrient runoff on the slope

To better understand the factors affecting the loss of solutes to the runoff, we applied our experimental data to the model developed by Wang et al. [25]. This model is described and justified in full detail in the publication cited above and is only briefly outlined here. The model is based on a soil water system that is divided into three vertically distributed horizontal layers: runoff or water ponding on the surface; an exchange layer below that; and the underlying soil. The variation in solute mass in the exchange layer changes over time and can be modeled using a power function. The transport of solutes from the exchange layer to the surface runoff is assumed to be dependent on the mass exchange rate. The model can be expressed as:

$$c(t) = k_m \frac{C_0 \rho_b H_o}{r(t)(P t_p + \rho_b \theta_o H_o)^b} r^b$$

(23)

where $c(t)$ represents the solute concentration (mg/L) in the runoff, $k_m$ is the exchange rate, $C_0$ denotes the initial solute concentration in the surface soil (g/g), $\rho_b$ is the soil’s dry bulk density (g/cm), $H_o$ is the depth of the exchange layer (cm), $r(t)$ is the runoff volume (L), $p$ is the rainfall intensity (cm/min), $t_p$ is the time between the initiation of rainfall and the formation of the runoff (min), $\theta_o$ is the initial soil moisture content (%), $r$ is time (min), and $b$ is an empirical parameter.

We adopted a new model, in which the presumed exchange layer is replaced by a mixing zone, which can be regarded as an extension of the deposited layer or “shield” concept presented by Hairsine and Rose [26]. Assuming that the exchange rate is controlled by raindrop splash and that the effects of diffusion can be neglected, we replaced the exchange rate $k_m$ with the variable $e_r$ developed by Gao et al. [27]. This substitution obviates the need to calibrate $k_m$. This new variable is the rate at which soil water is ejected from the soil during rainfall:
where $e_r$ is the raindrop-induced water transfer rate, $a$ is the detachability of the bare soil (g/cm$^2$) [29], $e$ is the rainfall-induced soil detachment per unit soil area (g/cm), $\rho$ is a constant parameter, $\rho_b$ is the bulk density of the dry soil (g/cm), $p$ is the rainfall intensity (mm/min), and $\theta$ is the soil water content (%) [30, 31]. Eqs. (23) and (24) can be combined to give:

$$c(t) = \frac{ap\theta C_o H_o}{r(t)(Pt + \rho_b \theta_o H_o)^b}$$

(26)

3. Results and discussions

3.1. The infiltration in the horizontal soil column

3.1.1. Cumulative infiltration with times

In the horizontal one-dimensional suction process, the soil water content increases with times, along with cumulative infiltration. However, the cumulative infiltration amount is different at different negative hydraulic heads and at various soil textures in the same infiltration time. From Figure 5, it can be seen that under different negative hydraulic head conditions, the change of cumulative is the largest in Loessal soil, followed by red glue soil and black loessial soil, with Lou soil showing the smallest cumulative change in the same infiltration time.

It can be seen from Figure 6 that the cumulative infiltration capacity with infiltration time is almost the same for each soil. The regulation in infiltration gradually decreases with a negative hydraulic head increase. Among them, the most significant change was observed in Loessial soil, and there is a large difference between $-9$ and $-12$ cm; the reason is the soil porosity ratio. Lou soil also shows a large difference. From $-2.5$ to $-18$ cm, red glue soil, black Loessial soil, and Lou soil show no significant difference, because no difference in porosity was observed between these two hydraulic heads. As for various soil textures, there are great differences in soil moisture absorption characteristics at different negative hydraulic head conditions, and the lighter the soil texture, the more is the difference.

In order to obtain a negative pressure suction effect on soil infiltration characteristics of quantitative analysis, we use the Kostiakov infiltration equation to fit the measured data. The Kostiakov model could fit the cumulative infiltration and infiltration time very well, which were shown Table 2.

From Table 2, it could be found that the correlation coefficient $R^2$ are all larger than 0.99 which indicated that the relationship between cumulative infiltration and time all have followed a power function under different negative hydraulic heads. For different soil textures, the order of coefficient $a$ is Loessial soil $>$ red glue soil $>$ dark loessial soil $>$ Lou soil, and index $b$ had no
Figure 5. Relationship between cumulated infiltration and measured infiltration time at different hydraulic heads with various soils. (a) – 2.5 cm; (b) – 6 cm; (c) – 9 cm; (d) – 12 cm; (e) – 15 cm; (f) – 18 cm.
significant changing tendency. For each soil, the parameter increased with the increase of negative hydraulic heads, while for parameter b, the opposite is true.

3.1.2. Determining the soil sorptivity based on horizontal one-dimensional experiments

According to horizontal one-dimensional experiments, we can easily obtain the soil sorptivity with the analysis to cumulative infiltration change with $t^{1/2}$. In Figure 7, it shows the change processes of cumulative infiltration with $t^{1/2}$ under four negative hydraulic heads. Sub-graphs a–f are Loessal soils, sub-graphs g–l are red glue soils, sub-graphs m–r are dark Loessial soils, sub-graph s–x are Lou soils.

As shown in Figure 7, the cumulative infiltrations of four kinds of soils had a linear relationship with $t^{1/2}$. We use a linear function to describe the curves, and the results were shown in Table 3.

**Figure 6.** Relationship between cumulated infiltration and measured infiltration time at different negative hydraulic heads. (a) Loessal soil; (b) red glue soil; (c) dark loessial soil; (d) Lou soil.
The regression coefficients $R^2$ in Table 3 were all above 0.95, which indicated that Philip equation can describe the infiltration rule very well under different negative hydraulic heads. Meanwhile, the soil sorptivity decreased with increasing soil viscosity (Loessal soil > red glue soil > dark Loessial soil > Lou soil). And beyond that, soil sorptivity decreases with the increasing negative hydraulic head.

### 3.1.3. Determining the parameters using Wang’s proposed equation

We used three kinds of textured soils (red glue soil, dark Loessial soil, and Lou soil) for horizontal one-dimensional infiltration experiments. The length of the soil column is 50 cm. The upper boundary was a constant hydraulic head (i.e., when $x = 0$, the hydraulic head was designed as a different negative hydraulic head). The hydraulic heads of red glue soil were $-21$ cm and $-30$ cm. The hydraulic heads of dark Loessial soil were $-18$ cm and $-24$ cm. The hydraulic heads of Lou soil were $-21$ cm and $-34$ cm. The lower boundary condition was free discharge. The duration of the experiment was 810 min. The saturated water content, the retention water content, and the initial water content were measured, noting down the changes of the cumulative infiltration with time. The parameters in the Brook-Corey model can be determined by MATLAB programming based on the experimental data. Figure 8 shows the relationship between cumulative infiltration and wetting front under different negative hydraulic head conditions.

As shown in Table 4, there is a good linear relationship between cumulative infiltration and wetting front which is in agreement with the theoretical derivation.
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Figure 7. Relationship between cumulated infiltration and square root of measured infiltration time at different negative heads. A. Loessal soil; B. Red glue soil; C. Dark loessial soil; D. Lou soil.
Substitution of $A_1$ and $A_2$ (listed in Table 5) into Eq. (10) yields the hydrodynamic parameters (in the Brooks-Core model). The results are listed in Table 5.

The soil water characteristic curves can be easily obtained by the values in Table 5. Comparing the calculated results and the experimental results (determined by centrifuge), the results were listed in Figure 9. As shown in Figure 9, the calculated data concur with experimental data. The results indicated that the parameters in the Brooks-Core model can be accurately and easily computed by the new method.

### 3.2. Analysis of soil thermal conductivity characteristics

Soil thermal conductivity reflects the size of soil thermal conductivity, and soil texture has a certain impact on thermal conductivity. According to the principle of thermal pulse probe, soil thermal parameters were measured and soil thermal conductivity was calculated. Figure 10 shows the curves of soil thermal conductivity with soil moisture content in four experimental sites of Shenmu (sand), Ansai, Yichuan, and Changwu. It can be seen from Figure 10 that soil thermal conductivity increases rapidly with the increase of water content when the soil water content is lower than 0.13 cm$^3$/cm$^3$. When the soil water content is higher than 0.13 cm$^3$/cm$^3$, the increasing trend of soil thermal conductivity is relatively reduced. Under the same moisture content, the trend of soil thermal conductivity is as follows: Shenmu sand soil > Ansai sandy loam soil > Yichuan clay loam soil > Changwu silty loam soil. So we can see that the higher the sand content, the lower the silt content, and the greater the soil thermal conductivity [22].

A study by Lu and Ren et al. showed that the soil can be divided into two categories according to sand content of the soil: It is coarse soil when the sand content is more than 40%, and $S_r = 0.3$, $K_e-S_r$ relationship curve of the coarse soil is divided into two linear ranges; when the sand content is less than 40%, it is fine soil, it is composed of $S_r = 0.13$ and $S_r = 0.30$, and the $K_e-S_r$ relationship curve of the fine soil is divided into three linear intervals; Figure 11 shows the $K_e-S_r$ curves of the normal form of soil thermal conductivity. The soil samples of Changwu and Ankang belong to fine soil, Shenmu, Mizhi, Ansai, Shangnan, Yichuan, Luochuan, and

| Hydraulic head (cm) | Loessal soil | Red glue soil | Dark loessial soil | Lou soil |
|--------------------|--------------|--------------|-------------------|---------|
|                    | $S$  | $R^2$ | $S$  | $R^2$ | $S$  | $R^2$ | $S$  | $R^2$ |
| −2.5               | 1.9904 | 0.994 | 0.6448 | 0.9903 | 0.4436 | 0.993 | 0.3466 | 0.9882 |
| −6                 | 1.8747 | 0.9903 | 0.6083 | 0.9878 | 0.4423 | 0.9893 | 0.344 | 0.9759 |
| −9                 | 1.7482 | 0.9836 | 0.5968 | 0.9856 | 0.3902 | 0.9814 | 0.3112 | 0.9843 |
| −12                | 1.4331 | 0.9706 | 0.6008 | 0.9959 | 0.4012 | 0.9838 | 0.2947 | 0.9777 |
| −15                | 1.3145 | 0.9686 | 0.5695 | 0.986 | 0.3776 | 0.9832 | 0.2875 | 0.9688 |
| −18                | 1.3654 | 0.9866 | 0.5597 | 0.9918 | 0.3552 | 0.9746 | 0.2918 | 0.9684 |

Table 3. The suction of four soil at different negative hydraulic heads.
Figure 8. Relation curve between cumulated infiltration and wetting front. A. Red glue soil; B. Dark loessial soil; C. Lou soil.
Table 4. Relation fitting values of cumulated infiltration and wetting front.

| Soil textural         | Hydraulic head (cm) | A1   | $R^2$  | Pressure head (cm) | A1   | $R^2$  |
|-----------------------|---------------------|------|--------|--------------------|------|--------|
| Red glue soil         | −21                 | 0.552| 0.9997 | −30                | 0.5192| 0.9995 |
| Dark loessial soil    | −18                 | 0.5539| 0.9997 | −24                | 0.5059| 0.9994 |
| Lou soil              | −21                 | 0.5356| 0.9995 | −24                | 0.5195| 0.9966 |

Table 5. Parameters calculation of three soils.

| Soil textural         | $n$   | $M$   | $h_a$ | $m/n$   |
|-----------------------|-------|-------|-------|---------|
| Red glue soil         | 0.17  | 2.51  | 26    | 14.65   |
| Dark loessial soil    | 0.32  | 2.94  | 40.11 | 9.18    |
| Lou soil              | 0.23  | 2.69  | 54.06 | 11.69   |

Figure 9. Compared between observed and calculated soil water characteristic curve. a. Red glue soil; b. Dark loessial soil; c. Lou soil.

Figure 10. Trend of soil thermal conductivity with water content.
Zhangye kind of matter is coarse soil. It can be seen from the figure that the $K_e$ value of fine soil is obviously smaller than the $K_e$ value of coarse soil when $0.2 < S_r < 0.6$.

3.2.1. Accuracy analysis of the soil thermal conductivity model

3.2.1.1. Campbell model

The sandy soil and sandy loam soil of coarse soil in Shenmu and Ansai, silty loam soil and silty clay loam soil of fine soil in ChangWu and Ankang, respectively, were selected. The thermal conductivities of these four soils were calculated by the Campbell model, and the results were shown in Figure 12. According to the statistical analysis, it can be seen that the difference between the calculated value and the measured value of the heat pulse is small when the water content of the soil is less than $0.20 \text{ cm}^3/\text{cm}^3$, and the relative error ($R_e$) of Shenmu sand soil and Ansai sandy loam soil are 13.51 and 9.56%, respectively; When the soil water content is higher than $0.20 \text{ cm}^3/\text{cm}^3$, the measured value of the heat pulse is larger than the calculated value, and the $R_e$ of Shenmu sand soil and Ansai sandy loam soil are 19.40 and 13.38%, respectively; the larger the volume of moisture content, the greater the difference; relative to the thermal pulse’s measured value, the calculation of the coarse soil model is too small for the coarse soil. For the fine soil, when the soil moisture content is less than $0.25 \text{ cm}^3/\text{cm}^3$, the $R_e$ are 26.29 and 21.19%, respectively, and the measured value of the heat pulse is larger than the calculated value of the model. When the soil moisture content is higher than $0.25 \text{ cm}^3/\text{cm}^3$, and the $R_e$ of Shenmu sand soil and Ansai sandy loam soil are 14.15 and 6.60%, respectively, the difference between the calculated value and the measured value of the heat pulse is small. Therefore, the model needs to be improved when calculating the thermal conductivity using the Campbell model [8, 9].

3.2.1.2. Johansen model, Côté-Konrad model and Lu-Ren model

Côté-Konrad model and Lu-Ren model are all semi-theoretical models of thermal conductivity based on Johansen model. The three models are used to calculate soil thermal conductivity for
the following four soils: Shenmu sand soil, Ansai sandy loam soil, Changwu silty loam soil, and Ankang silty clay loam soil. The model’s calculated values and measured values are shown in Figure 13. It can be seen from the figure that the calculated values of Johansen model are significantly smaller than the measured values, the calculation error is larger, the coefficient of determination $R^2$ is in the range of 0.656–0.827, the root mean square error (RMSE) is in the range of 0.0848–0.2548, and the relative error $R_e$ is in the range of 10.32–20.41%. For fine soil, the Côté-Konrad model and Lu-Ren model have a good fitting effect on soil thermal conductivity and the precision is high. Where the variation coefficient of $R^2$ is in the range from 0.842 to 0.940, the variation range of RMSE is from 0.0810 to 0.1208, the relative error $R_e$ is in the range from 9.67 to 10.57%. The coefficient of determination $R^2$ of the Lu-Ren model is in the range from 0.874 to 0.937, RMSE varied from 0.0725 to 0.1238, and the relative error $R_e$ varied from 8.28% to 9.91%. For coarse soil (sand content greater than 40%), the Côté-Konrad model and Lu-Ren model can still well fit soil thermal conductivity when the saturation $S_r < 50\%$, but the prediction accuracy of the model is poor, and the calculated value is obviously smaller than the measured value when the saturation $S_r > 50\%$. This phenomenon may be due to the large soil voids and the weak water-holding capacity, resulting in the measured value of water content being lower.

3.2.2. The improved Côté-Konrad model and the improved Lu-Ren model

The comparison between the calculated values of Campbell model, Johansen model, Côté-Konrad model, and Lu-Ren model and the measured values of the thermal pulses show that the soil thermal conductivity is closely related to the soil particle composition, organic matter

![Figure 12. Comparison of soil thermal conductivity values calculated by Campbell model with measured. (a) Shenmu sand soil, (b) Ansai sandy loam soil, (c) Changwu silty loam soil and (d) Ankang silty clay loam soil.](image-url)
content, and bulk density. For different soils with different textures, model parameters are also different. In Johansen model, the parameter $\lambda_s$ is related to the quartz content of the whole solid, and the thermal conductivity $\lambda_{\text{dry}}$ of the dry soil is related to soil bulk density. In Côté-Konrad model, the parameter $k$ is related to the content of coarse sand, small sand, clay, and organic matter. In Ren model, the parameter $\alpha$ is related to the soil sand content, and the soil is divided into coarse soil and fine soil according to the sand content. Under certain conditions, these models can calculate the soil thermal conductivity more accurately but cannot reflect the effect of soil particle composition and organic matter content on soil thermal conductivity. The improved Côté-Konrad model and the improved Lu-Ren model established the relationship between the model parameters, the composition of the particles, and the content of organic matter, respectively, and can describe the relationship between soil texture and soil thermal conductivity in detail.

In this chapter, the data of the five sites (466 sample points) of Mizhi, Shenmu (sandy loam), Ansai, Yichuan, and Changwu combined with $R^2$, $R_e$, and Figure 13, the relationship between the soil texture and the thermal conductivity, is fitting for the improved model, and the results

![Comparison of soil thermal conductivity values calculated by different models (Johansen model, Côté-Konrad model and Lu-Ren model) with measured. (a) Shenmu sand soil, (b) Ansai sandy loam soil, (c) Changwu silty loam soil and (d) Ankang silty clay loam soil.](image-url)
of the parameter fitting are shown in Table 6. The comparison between soil thermal conductivity and the measured values is shown in Figure 14. From the fitting error, it can be seen that the accuracy of the two improved models is not very different, and they have high accuracy. However, it can be seen from Figure 14 that the fitting value is larger than the measured value, and the soil thermal conductivity is more than 1.1 W/(m·K), the RMSE, $R^2$, and $Re$ are 0.0964, 0.9274, and 9.62% for the improved Côté-Konrad model when soil thermal conductivity is less than 0.6 W/(m·K), respectively. For the improved Lu-Ren model, although the fitting value and the measured value are also different, the discrete points in the figure are evenly distributed near the 1:1 line; RMSE, $R^2$, and $Re$ are 0.0961, 0.9278, and 9.59%, respectively.

According to $R^2$, $Re$, soil thermal conductivity of four samples of Shenmu (sand), Shangluo, Luochuan, and Ankang combined with the model parameter fitting values in Table 6 was predicted. The comparison between the predicted values and the measured values about the different models is shown in Figure 15a–d and Table 7, where, the sand content in four experimental sites was as follows: Shenmu > Shangluo > Luochuan > Ankang; clay content: Shenmu < Shangluo < Luochuan < Ankang; silt content: Shenmu < Shangluo < Luochuan < Ankang. Analysis of the simulation error shows that two improved models can be used to simulate soil thermal conductivity of different soils. For Shenmu sand soil and Ankang silty clay loam soil, the RMSE of the improved Côté-Konrad model is less than 0.1183, the $R^2$ is greater than 0.9259, and the $Re$ is less than 9.47%, which is better than the Côté-Konrad model, Model | Model parameters | RMSE | $R^2$ | $Re/%$
|----------------|----------------|--------|--------|----------|
| Improved Côté-Konrad model | $a_1 / b_1$ | $a_2 / b_2$ | $a_3 / b_3$ | $a_4 / b_4$ | 4.1381 | −0.8413 | 4.1506 | −0.2200 | 0.0964 | 0.9274 | 9.62 |
| Improved Lu-Ren model | $a_1 / b_1$ | $a_2 / b_2$ | 0.1080 | 0.0567 | 0.0961 | 0.9278 | 9.59 |

Note: $a_i$ is the parameter in the improved Côté-Konrad model; $b_i$ is the parameter in the improved Lu-Ren model; $i = 1, 2, 3, 4$.

Table 6. Parameters fitted values and errors by improved Côté-Konrad model and improved Lu-Ren model.

Figure 14. Fitted values of soil thermal conductivity by improved Côté-Konrad model and improved Lu-Ren model. (a) Improved Côté-Konrad model and (b) Improved Lu-Ren model.
Lu-Ren model, and improved Lu-Ren model. In other words, the improved Côté-Konrad model can be used to simulate the soil thermal conductivity for the soil with high sand content or high silt content. For the Shangnan loam soil and Luochuan clay loam soil, the RMSE of improved Lu-Ren model is less than 0.0815, $R^2$ is greater than 0.9326, and $R_e$ is less than 8.11%, which are obviously better than the other three models. In other words, the improved Lu-Ren model can be used to simulate soil thermal conductivity.

Figure 15. Comparison of soil thermal conductivity values predicted by improved Côté-Konrad model and improved Lu-Ren model with measured. (a) Shenmu sand soil, (b) Shangluo loam soil, (c) Luochuan clay loam soil, (d) Ankang silty clay loam soil and (e) Zhangye sandy clay loam soil.
In order to further verify whether the improved model can be extended to other soils, soil thermal conductivity of Zhangye samples in Gansu Province is predicted by the improved model. As the soil samples are sandy clay loam soil, the sand content is 60.13%. From the above model comparison analysis, we can see that the improved Côté-Konrad model is better for soil thermal conductivity with higher sand content. Figure 15e and Table 7 show the prediction results of thermal conductivity and the measured values and the simulation error, respectively. Through the error analysis, we can see that the results show that the improved Côté-Konrad model is slightly higher than other three models where the RMSE and \( R^2 \) of the improved Côté-Konrad model are 0.1026 and 0.9069, respectively, which is slightly higher than other three models, \( R_e \) is 8.15%, slightly lower than the other three models. Therefore, by selecting the appropriate improved model, soil thermal conductivity for different soil textures can be calculated accurately.

### Table 7. Soil thermal conductivity simulated values and errors by different soil thermal conductivity models in sampling area.

| Sampling area | Côté-Konrad model | Lu Ren model | Improved Côté-Konrad model | Improved Lu-Ren model |
|---------------|-------------------|-------------|---------------------------|-----------------------|
|               | RMSE  | \( R^2 \) | \( R_e \)% | RMSE  | \( R^2 \) | \( R_e \)% | RMSE  | \( R^2 \) | \( R_e \)% | RMSE  | \( R^2 \) | \( R_e \)% |
| Shenmu        | 0.1208 | 0.9401 | 9.67    | 0.1238 | 0.937  | 9.91    | 0.1183 | 0.9425 | 9.47    | 0.1366 | 0.9234 | 10.94   |
| Ankang        | 0.1088 | 0.9062 | 10.57   | 0.1014 | 0.9185 | 9.85    | 0.0951 | 0.9259 | 9.55    | 0.0986 | 0.8775 | 10.94   |
| Shngluo       | 0.081  | 0.8422 | 9.87    | 0.0725 | 0.8736 | 8.83    | 0.1243 | 0.8451 | 13.17   | 0.0766 | 0.9412 | 8.11    |
| Luochuan      | 0.0946 | 0.8872 | 10.5    | 0.0747 | 0.9298 | 8.28    | 0.1063 | 0.8514 | 10.97   | 0.0815 | 0.9326 | 8.21    |
| Zhangye       | 0.1216 | 0.8985 | 8.68    | 0.1349 | 0.8911 | 8.81    | 0.1026 | 0.9069 | 8.15    | 0.1034 | 0.9053 | 8.22    |

In order to further verify whether the improved model can be extended to other soils, soil thermal conductivity of Zhangye samples in Gansu Province is predicted by the improved model. As the soil samples are sandy clay loam soil, the sand content is 60.13%. From the above model comparison analysis, we can see that the improved Côté-Konrad model is better for soil thermal conductivity with higher sand content. Figure 15e and Table 7 show the prediction results of thermal conductivity and the measured values and the simulation error, respectively. Through the error analysis, we can see that the results show that the improved Côté-Konrad model is slightly higher than other three models where the RMSE and \( R^2 \) of the improved Côté-Konrad model are 0.1026 and 0.9069, respectively, which is slightly higher than other three models, \( R_e \) is 8.15%, slightly lower than the other three models. Therefore, by selecting the appropriate improved model, soil thermal conductivity for different soil textures can be calculated accurately.

#### 3.3. Solute transport in runoff by raindrops

Most of the parameters in our model were measured directly. The depths of the exchange layers, \( H_o \), were measured directly in the soil profiles and are shown in Table 8. Assuming that the exchange layer was saturated when the runoff was generated, the water content \( \theta \) was assumed to be equal to that in saturated soil, that is, 0.42 cm\(^3\)/cm\(^3\). The initial soil moisture content, \( \theta_o \), was 0.1 g/g except in the experiments where the initial soil moisture content was varied. The rainfall intensity, \( p \), was controlled by a computer. In all experiments except those investigating the influence of varying this parameter, \( p \) was fixed at 90 mm/h. The ponding times, \( t_P \), are shown in Table 8, \( \rho_o \) has a value of 1.35 g/cm\(^3\), and \( r(t) \) was measured directly.

To determine the relationship between the rainfall-induced soil detachment per unit area, \( e \), and the rainfall intensity, \( p \), we performed a series of rainfall experiments at rainfall intensities of 36 and 90 mm/h with a slope gradient of 5\(^\circ\) and a second series of experiments at rainfall intensities of 60 and 84 mm/h with a slope gradient of 15\(^\circ\).

The measured and simulated (assuming \( \rho = 2 \)) relationships between the rainfall-induced soil detachment per unit area, \( e \), and the rainfall intensity, \( p \), for a slope gradient of 5\(^\circ\) are shown in Figure 16. The model's predictions agree well with the experimental results (\( R^2 > 0.90 \)); this
result is consistent with the findings of Meyer [32], Foster [33], and Liebenow et al. [34], whose results were incorporated into the USDA WEPP model [35]. When the slope gradient is 15° or greater, \( r = 1 \), which is consistent with the results of Gao et al. [29] and corroborates the conclusions drawn by Sharma et al. [36, 37] and Jayawardena and Bhuiyan [38]. The relationships under these conditions between the rainfall-induced soil detachment per unit area, \( e \), and rainfall intensity, \( p \), at slope gradients of 15°, 25°, are shown in Figure 17. The values of the bare-soil detachability parameter, \( a \), were calculated using Eq. (25) and are shown in Table 9. The values of the bare-soil detachability parameter, \( a \), shown in Table 9, were quite similar, which may be attributed to the fact that the soil detachability is constant when the ponding depth is below a critical or breakpoint depth [30, 31, 38, 39]. Parameter \( b \) was estimated using

| Treatment                               | \( t_p \) (min) | \( H_o \) (cm) | \( \theta \) (%) |
|-----------------------------------------|-----------------|----------------|-----------------|
| Initial soil moisture content           |                 |                |                 |
| 5%                                      | 2.83            | 0.2            | 5               |
| 15%                                     | 1.83            | 0.23           | 15              |
| 20%                                     | 1.21            | 0.24           | 20              |
| Rainfall intensity                      |                 |                |                 |
| 60 mm h\(^{-1}\)                        | 4.00            | 0.18           | 10              |
| 96 mm h\(^{-1}\)                        | 2.40            | 0.23           | 10              |
| 129 mm h\(^{-1}\)                       | 1.70            | 0.27           | 10              |
| Slope gradient                          |                 |                |                 |
| 5°                                      | 3.9             | 0.16           | 10              |
| 15°                                     | 3.3             | 0.18           | 10              |
| 25°                                     | 2.3             | 0.3            | 10              |

*Values shown represent the average of three runs.

Table 8. Experimental parameters used in the numerical model.

![Figure 16](image.png)

Figure 16. The relationship between the rainfall-induced soil detachment per unit area, \( e \), and rainfall intensity, \( p \).
The best fit to the experimental data listed in Table 9. The predictions made using Eq. (26) are compared to our experimental results in Figure 18.

The simulated data agreed well with the experimental results for all three treatments except for the experiment where the rainfall intensity was 129 mm/h, suggesting that the use of the raindrop-induced water transfer rate, $e_r$, in place of the exchange rate, $k_m$, is reasonable for conditions involving relatively natural rainfall. The model, however, did not appear to accurately predict the solute concentrations in the runoff observed under conditions that give rise to severe soil erosion.

Figure 17. The relationship between the rainfall-induced soil detachment per unit area, $e$, and rainfall intensity, $p$, at slope gradients of 15° and 25°.

| Treatment                                      | $e$ ($\times 10^{-3}$ g cm$^{-2}$) | $p$ (cm min$^{-1}$) | $\rho$ | $a$ (g cm$^{-3}$) | $b^*$ |
|------------------------------------------------|-----------------------------------|---------------------|--------|-----------------|-------|
| Initial soil moisture content (%)              |                                   |                     |        |                 |       |
| 5                                              | 2.18                              | 0.15                | 2      | −0.4            | 0.094 |
| 15                                             | 1.9                               | 0.15                | 2      | −0.4            | 0.083 |
| 20                                             | 1.68                              | 0.15                | 2      | −0.4            | 0.075 |
| Rainfall intensity (mm h$^{-1}$)               |                                   |                     |        |                 |       |
| 60                                             | 1.16                              | 0.1                 | 2      | −0.2            | 0.116 |
| 96                                             | 1.86                              | 0.16                | 2      | −0.2            | 0.082 |
| 129                                            | 3.93                              | 0.215               | 2      | −0.2            | 0.085 |
| Slope gradient (°)                             |                                   |                     |        |                 |       |
| 5                                              | 1.85                              | 0.15                | 2      | −0.35           | 0.082 |
| 15                                             | 10.11                             | 0.15                | 1      | −0.35           | 0.072 |
| 25                                             | 11.93                             | 0.15                | 1      | −0.35           | 0.082 |

*Calibrated to best fit the runoff solute data.

Table 9. Data used in calculating the bare-soil detachability, $a$. 

the best fit to the experimental data listed in Table 9. The predictions made using Eq. (26) are compared to our experimental results in Figure 18.
Our results support the conclusion drawn by Walter et al. [40], who argued that the depth of the exchange layer decreases as the rate of infiltration increases. The initial soil moisture content, rainfall intensity, and slope gradient influence the solute concentration of the runoff solution by virtue of their effects on the depth of the exchange layer, the infiltration rate, and the length of time between the initiation of rainfall and the formation of the runoff.

Figure 18. Simulated and measured runoff concentrations of potassium under different initial soil moisture contents (A), rainfall intensities (B), and slope gradients (C).

Our results support the conclusion drawn by Walter et al. [40], who argued that the depth of the exchange layer decreases as the rate of infiltration increases. The initial soil moisture content, rainfall intensity, and slope gradient influence the solute concentration of the runoff solution by virtue of their effects on the depth of the exchange layer, the infiltration rate, and the length of time between the initiation of rainfall and the formation of the runoff.
The agreement of the simulated results with the measured data was quantified by calculating the root mean square error (RMSE) [31]. RMSE can be expressed as:

$$RMSE = \sqrt{\frac{1}{N} \sum_{i=1}^{N} (p_i - o_i)^2}$$

(27)

where $N$ is the total number of data points, $p_i$ is a given simulated data point, and $o_i$ is the corresponding experimental data point. The RMSEs are shown in Table 10. Table 10 and Figure 18 illustrate that the model has correctly captured the temporal behavior of the solute concentration in the runoff under all conditions investigated.

Figure 18A shows that the measured and simulated solute concentrations for different initial soil moisture contents changed with time. The results indicated that the refined model [25] could predict the movement of solutes in the overland flow under different initial soil moisture contents. Also, the higher initial soil moisture contents were associated with higher solute

| Treatment                        | RMSE (mg l$^{-1}$) | $R^2$ |
|----------------------------------|--------------------|-------|
| Initial soil moisture content (%)|                    |       |
| 5                                | 0.227              | 0.90  |
| 15                               | 0.245              | 0.93  |
| 20                               | 0.308              | 0.91  |
| Rainfall intensity (mm/h)        |                    |       |
| 60                               | 0.295              | 0.80  |
| 96                               | 0.229              | 0.88  |
| 129                              | 0.508              | 0.46  |
| Slope gradient (°)               |                    |       |
| 5                                | 0.081              | 0.94  |
| 15                               | 0.127              | 0.88  |
| 25                               | 0.336              | 0.86  |

Table 10. The root means square errors (RMSEs) and $R^2$ between the measured and simulated data.

Figure 19. Variation of the sediment concentrations over time at different rainfall intensities.
Figure 20. Graph of potassium concentration calculated by Eq. (26) against potassium concentration measured (under different initial soil moisture content, rainfall intensity conditions).
concentrations per unit time. **Figure 18A** and **Table 10** indicate that the differences between the measured and simulated solute concentrations under an initial soil moisture content of 20% were more distinct than those under the other two initial soil moisture contents, which implied

![Graphs showing potassium concentration](image_url)

**Figure 21.** Graph of potassium concentration calculated by Eq. (26) against potassium concentration measured (under different slope gradient conditions).
that the model did not accurately predict the solute concentrations of the runoff in conditions of severe soil erosion.

Comparisons between the simulated and the experimental solute concentrations for the different rainfall intensities over 60 min are shown in Figure 18B. At a rainfall intensity of 129 mm/h, the solute concentration of the runoff increased substantially between 37 and 49 min after the initial generation of the runoff (Figure 18B). The mass of sediment in the runoff between 37 and 43–49 min showed a corresponding spike (Figure 19), which indicated that solute loss is closely related to sediment loss [41–45]. These results indicated that significant erosion of the surface soil occurred at the bottom of the slope during the experiments. Deeper soil layers were exposed to water in which the solute concentrations were higher than in those washed away. Consequently, the solute concentration of the runoff increased as these solutes were transferred from the soil under the influence of the runoff and the splashing caused by raindrops. Soil erosion thus promoted increased solute concentrations in the runoff.

Figure 18C shows that the measured and simulated solute concentrations for different slope gradients also changed with time. The simulated data were highly correlated with the measured data for solute concentration in the runoff. This degree of correlation demonstrated that the model captured the temporal behavior of the solute transport in the runoff. Increasing the gradient of the slope increased the erosion capacity of rain drops and water flow. Increasing the slope gradient also led to increases in the RMSE (Table 10) and $R^2$. Figures 20 and 21 show the relationships between potassium concentrations observed in the runoff and predicted using Eq. (26). The graph indicates the model accurately predict the solute transport in the runoff with the solute concentration being at a much lower level.

4. Conclusions

In order to understand the whole process of water-solute-heat transport and nutrient loss, we determined water movement, solute, and heat transport through columns of disturbed soil samples. And we also carried out simulated rainfall experiments on an artificial slope to study nutrient loss.

The results were as follows:

1. Data obtained with experimental infiltration under negative hydraulic heads were employed to analyze the relationship between the Philip model and Kostiakov empirical model, showing as well that they were identical in terms of negative hydraulic heads; Wang’s equation could describe the infiltration process very well.

2. The Horton empirical model can be used to describe the variation of soil thermal conductivity; the calculated values of Campbell model and Johansen model have large differences with the measured values. However, the calculated results of Côté-Konrad model and Lu-Ren model are in good agreement with the measured values. The improved Côté-Konrad model and improved Lu-Ren model can use the soil texture to predict soil thermal conductivity. For two improved models, the coefficients of determination $R^2$ are above 0.92 and the relative errors Re are less than 9.6%. For the soils with high sand content or silt
content, the improved Côté-Konrad model is superior to Côté-Konrad model, Lu-Ren model, and the improved Lu-Ren model. For the soils with low sand content and silt content, the Lu-Ren model is obviously better than the other three models. The relationship between the parameters of the model, particle composition, and organic matter content can be predicted by two improved models. These models can describe the relationship between the soil’s basic physical parameters and thermal conductivity in detail. Thus, soil thermal conductivity can be predicted more accurately by choosing the appropriate improved model based on the different soil texture.

3. The refined power functions of a model of solute transport were illustrated and tested using simple experiments. The model fit the experimental data very well. Our results also indicated that the constant parameter, $\rho$, was equal to 1 when the slope gradient was 15° or larger and equal to 2 when the slope gradient was less than 15°. The soil detachability was confirmed to be independent of the rain intensity and was a constant in all treatments. The model, however, could not accurately predict the solute concentrations in the runoff under conditions of severe soil erosion. The initial soil moisture content, rainfall intensity, and slope gradient influenced the solute concentration in the runoff, depth of the exchange layer, infiltration rate, and length of time between the initiation of rainfall and the generation of the runoff.

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References

[1] Darcy H. Less Fontaines Publiques de la Ville de Dijon. Paris, France: Dalmont; 1956

[2] Green WH, Ampt CA. Studies on soil physics. I. Flow of air and water through soils. Journal of Agricultural Science. 1911;4:1-24
[3] Richards LA. Capillary conduction of liquids in porous mediums. Physics. 1931;1:318-333

[4] Philip JR. Theory of infiltration. Soil Science. 1957;83(5):345-357

[5] Kostiakov AN. On the dynamics of the coefficient of water percolation in soils and on the necessity for studying it from a dynamic point of view for purpose of amelioration. In: Transactions of 6th Committee International Society of Soil Science; Russia. 1932. pp. 17-21

[6] Horton RE. An approach toward a physical interpretation of infiltration capacity. Soil Science Society of America Journal. 1940;5:399-417. DOI: 10.2136/sssaj1941.036159950005000C0075x

[7] Holton HN. A concept of infiltration estimates in watershed engineering. AICHE Journal. 1961;150(1):41-51

[8] Shu ST. Capillary-tube infiltration model. Journal of Irrigation and Drainage Engineering. 1993;119(3):514-521. DOI: 10.1061/(ASCE)0733-9437(1993)119:3(514)

[9] Ghosh Kumar R. A note on the infiltration equation. Soil Science. 1983;136(6):333-338. DOI: 10.1097/00010694-198312000-00001

[10] Parlange JY, Barry DA, Parlange MB, et al. New approximate analytical technique to solve Richards Equation for arbitrary surface boundary conditions. Water Resources Research. 1997;33(4):903-906. DOI: 10.1029/96WR03846

[11] Hogarth WL, Parlange JY. Application and improvement of a recent approximate analytical solution of Richards' equation. Water Resources Research. 2000;36(7):1965-1968. DOI: 10.1029/2000WR900042

[12] Lei ZD, Yang SX, Xie SC. Soil Water Dynamics. Beijing: Tsinghua University Press; 1988 (in Chinese)

[13] Yang SX, Lei ZD, Xie SC. General program of one-dimensional flow through unsaturated homogeneous soil. Acta Pedologica Sinica. 1985;1:24-34 (in Chinese)

[14] Yi L, Mingan S, Wenyan W, et al. Influence of soil textures on the thermal properties. Transactions of the Chinese Society of Agricultural Engineering. 2003;19(4):62-65 (in Chinese)

[15] Chung SO, Horton R. Soil heat and water flow with a partial surface mulch. Water Resources Research. 1987;12(11):2175-2186. DOI: 10.1029/WR023i012p02175

[16] Campbell GS. Soil Physics with BASIC. Amsterdam, the Netherlands: Elsevier; 1985. pp. 221-234

[17] De Vries DA. Thermal properties of soils. In: Physics of Plant Environment. Amsterdam: North-Holland; 1963. pp. 210-235

[18] Johansen O. Thermal Conductivity of Soils. Trondheim: Norwegian University of Science and Technology; 1977

[19] Côté J, Konrad JM. A generalized thermal conductivity model for soils and construction materials. Canadian Geotechnical Journal. 2005;42(3):443-458. DOI: 10.1139/t04-106
[20] Lu S, Ren T, Gong YS. An improved model for predicting soil thermal conductivity from water content at room temperature. Soil Science Society of America Journal. 2006;71(1):8-14. DOI: 10.2136/sssaj2006.0041

[21] Ting L, Wang Q, Jun F. Modification and comparison of methods for determining soil thermal parameters. Transactions of the Chinese Society of Agricultural Engineering. 2008;24(3):59-64. (in Chinese). DOI: 10.3969/j.issn.1002-6819.2008.3.012

[22] Wang S, Wang Q, Jun F, et al. Soil thermal properties determination and prediction model comparison. Transactions of the Chinese Society of Agricultural Engineering. 2012;28(5):78-84. (in Chinese). DOI: 10.3969/j.issn.1002-6819.2012.05.014

[23] Wang W, Jianbo L, Wang S, et al. Spatial variability of soil thermal parameters and its fitting method. Transactions of the Chinese Society for Agricultural Machinery. 2015;46(4):120-125. DOI: 10.6041/j.issn.1000-1298.2015.04.018

[24] Wallach R, Galina G, Rivlin J. A comprehensive mathematical model for transport of soil-dissolved chemicals by overland flow. Journal of Hydrology. 2001;247:85-89. DOI: 10.1016/S0022-1694(01)00365-1

[25] Wang QJ, Wang WY, Shen B, Shao MA. Interacting depth of rainfall–runoff–soil solute. Journal of Soil Erosion and Water Conservation. 1998;2(4):41-46 (in Chinese)

[26] Hairsine PB, Rose CW. Rainfall detachment and deposition: Sediment transport in the absence of flow-driven processes. Soil Science Society of America Journal. 1991;55(2):320-324. DOI: 10.1016/S0022-1694(04)00174-X

[27] Gao B, Walter MT, Steenhuis TS, et al. Rain induced chemical transport from soil to runoff: Theory and experiments. Journal of Hydrology. 2004;295:291-304. DOI: 10.1016/S0022-1694(04)00174-X

[28] Wang QJ, Robert H, Shao MA. Horizontal infiltration method for determining Brooks-Corey model parameters. Soil Science Society of America Journal. 2002;66:1733-1739. DOI: 10.2136/sssaj2002.1733

[29] Gao B, Walter MT, Steenhuis TS, et al. Investigating ponding depth and soil detachability for a mechanistic erosion model using a simple experiment. Journal of Hydrology. 2003;277(1–2):116-124. DOI: 10.1016/S0022-1694(03)00085-4

[30] Moss AJ, Green P. Movement of solids in air and water by raindrop impact—Effects of drop-size and water-depth variations. Australian Journal of Soil Research. 1983;21(3):257-269. DOI: 10.1071/SR9830257

[31] Willmott CJ. Some comments on the evaluation of model performance. Bulletin of the American Meteorological Society. 1982;63(11):1309-1313. DOI: 10.1175/1520-0477(1982)063<1309:SCOTEO>2.0.CO;2

[32] Meyer LD. Soil-erosion research leading to development of the universal soil erosion loss equation. Science. 1982;26:1-16. DOI: 10.2136/sssaj1993.03615995005700030007x
[33] Foster GR. Modeling the erosion process. In: Han CT, editor. Hydrological Modeling of Small Watersheds: Monograph No. 5. St Joseph, MI: ASAE; 1982. pp. 297-379

[34] Liebenow AM, Elliot WJ, Laflen JM, et al. Inter rill erodibility-collection and analysis of data from crop-land soils. Transactions of the Chinese Society of Agricultural Engineering. 1990;33(6):1882-1888 (in Chinese)

[35] Laflen JM, Elliot WJ, Simanton JR, et al. WEPP soil erodibility experiments for rangeland and cropland soils. Journal of Soil and Water Conservation. 1991;49(1):39-44

[36] Sharma PP, Gupta SC, Foster GR. Predicting soil detachment by raindrops. Soil Science Society of America Journal. 1993;57:674-680

[37] Sharma PP, Gupta SC, Foster GR. Raindrop-induced soil detachment and sediment transport from interrill areas. Soil Science Society of America Journal. 1995;59:727-734. DOI: 10.2136/sssaj1995.03615995005900030014x

[38] Jayawardena AW, Bhuiyan RR. Evaluation of an inter rill soil erosion model using laboratory catchment data. Hydrological Processes. 2015;13(1):89-100. DOI: 10.1002/(sici)1099-1085(199901)13:1<89::aid-hyp677>3.0.co;2-t

[39] Proffitt APB, Rose CW, Hairsine PB. Rainfall detachment and deposition: Experiments with low slopes and significant water depths. Soil Science Society of America Journal. 1991;55:325-332. DOI: 10.2136/sssaj1991.03615995005500020004x

[40] Walter MT, Gao B, Parlange JY. Modeling soil solute release into runoff with infiltration. Journal of Hydrology. 2007;347:430-437. DOI: 10.1016/j.jhydrol.2007.09.033

[41] Catt JA, Quinton JN, Rickson RJ, et al. Nutrient losses and crop yields in the Woburn erosion reference experiment. In: Rickson RJ, editor. Conserving Soil Resources: European Perspective. Oxford: CAB International; 1994. pp. 94-104

[42] Hansen AC, Nielsen JD. Runoff and loss of soil and nutrients. In: Correll A, editor. Surface Runoff Erosion and Loss of Phosphorus at two Agricultural Soils in Denmark. Tjele: Danish Institute of Plant and Soil Science; 1995. pp. 149-188

[43] Hargrave AP, Shaykewich CF. Rainfall induced nitrogen and phosphorus losses from Manitoba soils. European Journal of Soil Science. 1997;77:59-65. DOI: 10.4141/S95-034

[44] Teixeira PC, Misra RK. Measurement and prediction of nitrogen loss by simulated erosion events on cultivated forest soils of contrasting structure. Soil and Tillage Research. 2005;83:204-217. DOI: 10.1016/j.still.2004.07.014

[45] Guo TL, Wang QJ, Li DQ, et al. Sediment and solute transport on soil slope under simultaneous influence of rainfall impact and scouring flow. Hydrological Processes. 2010;24:1446-1454. DOI: 10.1002/hyp.7605