The Soil Infiltration Model in KINEROS2

Introduction

KINEROS2 contains a new soil infiltration model which allows more detailed specification of the soil profile for each hydrologic element, including specification of the characteristics of the bed for an infiltrating channel. The additional detail is not required, however, and a simple single soil layer can be specified as for KINEROS. The new formulation allows either a one or two layer soil profile, a new physically-based approximation for the redistribution of soil water, including recovery of infiltration capacity during a hiatus, and a method that more accurately determines infiltration rates following a hiatus. Only one other parameter is required for the redistribution algorithm. Further, at the user s option, the ensemble effect of normally occurring spatial variation in soil hydraulic conductivity, K_s, can be simulated by providing a value for the coefficient of variation for this parameter, CvK.

The soil infiltration model is an extension of the model used in KINEROS, which describes infiltration capacity f_c as a function of infiltrated depth I. Independent (practically) of the pattern of rainfall rate (r) prior to a given time, the infiltration capacity during rainfall is a function of the total depth which has been infiltrated at values of $r > K_s$ prior to that time. Periods when $r < K_s$ are now dealt with realistically in KINEROS2.

General Infiltration Properties

<u>Describing the soil profile</u>. KINEROS required 3 basic parameters to describe the infiltration properties of a soil: the field effective saturated hydraulic conductivity, K_s, the integral capillary drive G, and the porosity ϕ. To allow estimation of the soil redistribution behavior, KINEROS2 asks for one additional parameter, λ, which is called the *pore size distribution index*, so named by Brooks and Corey (1964), whose simple description of the soil hydraulic characteristics is adopted for use in this model. As indicated above, there is an optional parameter CvK, which describes the random variation in space of the hydraulic properties of the soil. For a two layer profile, the above parameters will be indexed with a 1 or 2 to indicate that they apply to the upper or lower soil layers, respectively. KINEROS2 requires 4 parameters to describe the second soil layer: the upper soil layer depth, z_1; values for K_{s_2}, G_2, and ϕ_2; plus a value for λ_2 to be used in redistribution calculations. As in KINEROS there is another optional parameter that allows even more explicit characterization of a soil profile, the content of large rocks, which represent solid volume of larger than capillary size which restricts storage. A value for ROCK may be entered for each layer of a 2 layer profile.

As for KINEROS, there is an event-dependent variable: the initial relative saturation of the upper soil layer, SI. Relative saturation is a scaled value of water content, where a value of 1 is equal to a water content equal to the porosity, ϕ. Water content by volume is θ, where $\theta = \phi S$, and there is a natural upper limit to S which is less than 1 (parameter S_{max}). θ_s is used here for ϕS_{max}.

General model relationships. We define *infiltrability*, following Hillel (1971), as the limiting rate at which water can enter the soil surface. This is more often called infiltration capacity, but capacity is not a dynamic term. The general one-layer model for infiltrability, f_c, as a function of infiltrated depth, I, is (Parlange et al., 1978)

$$f_c = K_s\left[1 + \frac{\alpha}{exp(\alpha I / B) - 1}\right] \tag{1}$$

where B is $(G + h_w)(\theta_s - \theta_i)$, combining the effects of net capillary drive, G, surface water depth, h_w, and unit storage capacity, $\Delta\theta_i = (\theta_s - \theta_i)$. The parameter α represents the soil type: α is near 0 for a sand, in which case Eq. (1) approaches the Green-Ampt relation; and α is near 1 for a well-mixed loam, in which case Eq. (1) represents the Smith-Parlange infiltration equation. Most soils are best described by a value of α near 0.85, and this value is assumed in KINEROS2. While α is fixed in K2, the equation used is in principle quite general and physically-based.

There are some fundamental properties of the infiltrability relation. It is useful to observe the relation in scaled terms. For this purpose one may define (taking $h_w = 0$)

$$f_* = \frac{f - \overline{K}_s}{\overline{K}_s} \tag{2}$$

$$I_* = \frac{I}{G\Delta\theta} \tag{3}$$

$$t_* = t\,\frac{\overline{K}_s}{G\Delta\theta} \tag{4}$$

It follows that rainrate r is scaled exactly as f, that is, $r_* = \left(r - K_s\right) / K_s$. Bars are used with K_s to indicate areal mean values. Below, we will deal with the spatial variation of this parameter. Using this scaling, Eq.(1) becomes

$$f_{c*} = \frac{\alpha}{exp\left(\alpha I_*\right) - 1} \tag{5}$$

Figure 1 illustrates Eq. (5) in scaled coordinates. For values of I_* smaller than about 0.1, infiltration is dominated by capillary gradient in the soil, and unaffected by gravity or α, so that the indicated asymptote is the gravity-free relation $f_* = 1 / I_*$. The large I or long-time asymptote ($I_* > 10$) is also independent of the value of α.

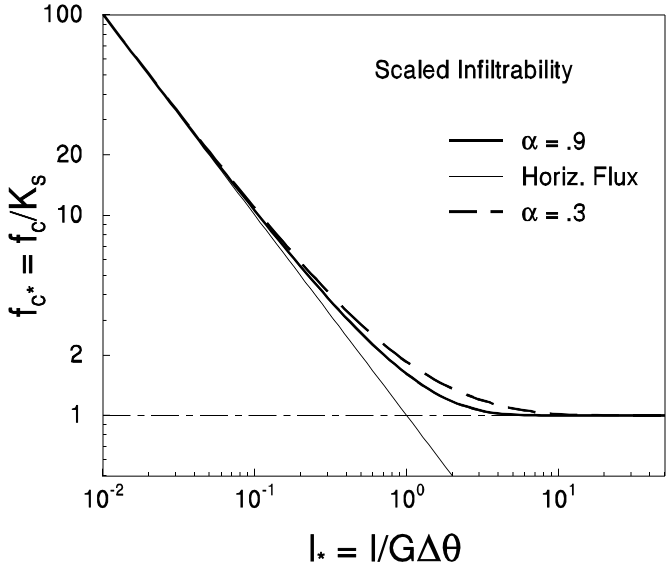

Figure 1. Scaled basic infiltration relation, Eq. (5).

The net capillary drive parameter G is defined as

$$G = G(0) = \int_{-\infty}^{0} \frac{K(h)}{K_s} dh \qquad (6)$$

When G is used without a subscript, it is assumed that the upper limit in Eq. (6) is 0, as shown here. For redistribution, as presented below, the upper limit is changed for a different use of this integral parameter. In KINEROS2, for simplicity, the relation $K(h)$ is taken from the soil characteristic relation of Brooks and Corey, which is

$$\frac{K(\Psi)}{K_s} = \left(\frac{\Psi_B}{\psi} \right)^{2+3\lambda} \qquad (7)$$

By integrating Eq. (6) using this relation, one can determine that

$$G = \Psi_B \frac{2 + 3\lambda}{1 + 3\lambda} \qquad (8)$$

From this, the parameter ψ_B may be obtained from an estimate of G and λ. Values of λ and G for a range of soil textures, which may provide guidelines in cases without direct measurements, are given in Table 1.

Table 1. Estimation guide for soil hydraulic properties based on sample data, showing ranges for ±one standard deviation[1]

Texture Class	Sample Size Used[1]	Total Porosity ***n***	Residual Water Content θ_r	Effective Saturation θ_s	Pore Size Distribution λ	Mean Capillary Drive[2] ***G*** (cm)
Sand	762	0.437 ± .063	0.020 ± .019	0.417 ± .063	0.69 ± .40	5.
Loamy Sand	338	0.437 ± .069	0.035 ± .032	0.401 ± .062	0.55 ± .32	7.
Sandy Loam	666	0.453 ± .102	0.041 ± .065	0.412 ± .129	0.38 ± .14	13.
Loam	383	0.463 ± .088	0.027 ± .047	0.434 ± .100	0.25 ± .17	11.
Silt Loam	1206	0.501 ± .081	0.015 ± .043	0.486 ± .092	0.23 ± .13	20.
Sandy Clay Loam	498	0.398 ± .066	0.068 ± .069	0.330 ± .095	0.32 ± .24	26.
Clay Loam	366	0.464 ± .055	0.075 ± .099	0.390 ± .111	0.24 ± .14	26.
Silty Clay Loam	689	0.471 ± .053	0.040 ± .078	0.432 ± .085	0.18 ± .14	35.
Sandy Clay	45	0.430 ± .060	0.109 ± .096	0.321 ± .114	0.22 ± .18	30.
Silty Clay	127	0.479 ± .054	0.056 ± .08	0.423 ± .089	0.15 ± 0.11	38.
Clay	291	0.475 ± .048	0.090 ± .105	0.385 ± .116	0.16 ± 0.13	41.

Notes:
1. Data from Rawls, et al., 1982
2. Obtained using mean class parameters and Eq.(7)

<u>Simulating Rainfall Infiltration (uniform soil).</u> Some initial portion of rainfall $r(t)$ increases the infiltrated depth, I, without causing runoff, since the value of f_c for small I is very large. When I increases until r begins to exceed f_c, ponding occurs and Eq. (1) is used to predict f and the rainfall excess, $r - f$. Any rainfall depth I is accompanied by a wetted soil depth with a downward-moving lower 'front' at depth $z = I / \Delta\theta_i$.

Two-Layered Soil Profiles

A two-layer profile, with upper soil depth z_1, can have either the upper layer or the lower layer acting as the most restrictive or limiting layer. This is a function in general of the relative values of

K_{s_1} and K_{s_2}, with the layer having the lowest value generally the most restrictive. A special case exists for short time infiltration (only under very high rainfalls and shallow surface layers) where the two soils may act identically if the products $(G_1)(K_{s_1})$ and $(G_2)(K_{s_2})$ are equal (Smith, 1990).

Restrictive Upper Layer. This is a common case for a *crusted* soils, where the upper layer is not only restrictive, but is relatively thin. However, a thin layer is not required by the model used here. In this case, while K_{s_1} is less than K_{s_2}, there always exist a rainfall intensity or pattern of intensities for which ponding and runoff will occur after the wetting front passes beyond the upper layer, as well as cases where ponding will occur prior to the wetting front filling the upper layer. We define a critical depth of rainfall just sufficient to fill the upper soil layer as I_c, where $I_c = \phi(S_{max} - SI)z_1 = (\theta_s - \theta_i)z_1$. There exists then one flux value for which ponding will occur just as $I = I_c$, and we refer to that value as r_c. For larger values of r, ponding will occur in the upper layer, and ponding can be simulated just as for a single soil profile. For smaller r, the front will move into the second layer prior to ponding.

When the wetting front enters the second layer, both the profile effective values for final or asymptotic infiltration rate, now referred to as $K\infty$, will change. For a homogeneous soil or infiltrated soil depths confined to an upper layer, this value is K_{s_1}. The value of $K\infty$ must be calculated based on both layer profiles when there is an upper restrictive layer. For this case the effective value of G, called G_e, will also change.

$K\infty$ is found as follows (Smith, 1990). The soil capillary head ψ is continuous across the layer interface before and after wetting. The value of $K\infty$ may be found by solving the steady Darcy flow equation across the layer interface, resulting in

$$z_1 = \int_{\Psi_c}^{\Psi_0} \frac{K_1(\Psi)\,d\Psi}{K_2\left(\Psi_c\right) - K_1(\Psi)} \tag{9}$$

This equation is solved iteratively based on the relations $K(\psi)$, Eq. (7), for each layer, to obtain the values of both ψ_c and $K_2(\psi_c)$. Since there is a unit gradient in the second layer at steady flow, $K_2(\psi_c)$ is equal to $K\infty$. When the first layer is passed, the value of $K\infty$ is assumed to change directly to the asymptotic value.

The value of G_e is transitional, based on the depth that the wetting front has moved into the second layer. Just at the time the wetting front passes the interface, a "matching" value of G is found, called G_a, such that infiltration capacity predicted using G_a and $K\infty$ just matches the value r_c, defined above, which uses G_1 and K_{s_1}. Conversely, at very large times, $G_e = G_2(\psi_c)$, since the wetting front is entirely in the second soil and the effect of the upper layer on G is only in control of ψ at the interface. The value of G_e at any intermediate time is found as a weighted sum of the two values

G_a and G_2:

$$G_e = \omega G_a + (1 - \omega) G_2 \tag{10}$$

where is a weighting function which decreases exponentially as the wetting front moves into the second soil:

$$\omega = exp\left[-b\left(I / I_c - 1\right)\right] \qquad I > I_c \tag{11}$$

Coefficient b varies somewhat depending on the relative values of K_{s_1} and K_{s_2}. Equation (11) and the value range of coefficient b (included in KINEROS2 code) have been established by analysis of experiments using a numerical solution to Richards' equation.

Restrictive Lower Layer. ($K_{s_2} < K_{s_1}$) Similar to the case for a restrictive upper layer, and also depending on relative values of soil parameters and rainfall rate, ponding can occur with the wetting front in either layer. However, unlike the previous case, the long term final infiltration rate is always K_{s_2}, and unlike the previous case, parameters change rather suddenly upon the encounter of the wetting front with the second layer. G_a is G_1 when the wetting front is in the upper layer, and G_a is G_2 when in the lower layer. Further, there can be ponding caused by the lower layer for cases where the rainfall is less the K_{s_1}, when ponding can not occur in the upper layer. For this case, which is called saturation runoff, the upper layer must fill with water that cannot enter the lower soil before surface ponding and runoff occur.

Soil Water Redistribution.

Many rainfall events consist of more than one period of runoff producing rainfall, with an intervening period during which significant drying of the soil can occur. Until recently the recovery of infiltrability during this period was crudely approximated or else ignored. The redistribution/reinfiltration method used in KINEROS2 is described in Smith *et al.* (1993) and Corradini *et al.* (1994). By equating the downward flux movement at the wetting front with the reduction in water content in an assumed pseudo-rectangular wetted region containing water of depth I, an equation may be written for the change in water content at the surface, θ_0:

$$\frac{d\theta}{dt} = \frac{\Delta\theta_{i0}}{I}\left[r - K_i - \left(K(\theta_0) + \frac{\beta p K_s \Delta\theta_{i0} G(\theta_i, \theta_0)}{I}\right)\right] \tag{12}$$

in which

$\Delta\theta_{i0}$ is θ_0 - θ_i, the water content of the redistributing portion,

θ_i is the original water content, below the redistribution front,

r is the rate of input at the surface during redistribution, which may be small, negative (evaporation), positive, or 0),

is a shape factor,

p is an effective depth factor, and

$G(\theta_i,\theta_0)$ is the effective capillary drive of the shrinking wetting front, which is reduced relative to that of an infiltrating G due to the fact that θ_0 is less than θ_s.

This equation is easily solved during simulation by the Runge-Kutta method. For two-layer soils, where the redistributing block includes both soil layers, the reduction of θ in both layers is treated by linking them with an assumed common value of ψ across the interface, and using $K_s = K_{s_2}$ and $G(\theta_i,\theta_0) = G_2(\theta_i,\theta_0)$ in Eq. (12).

Rainfall Prewetting. Another normal phenomenon in actual rainfall patterns that is not dealt with by other infiltration models is the effect of initial, slow rainfalls which may precede a runoff-producing period, in which the effective "initial" soil water content used in the infiltration calculations is changed. KINEROS2 uses a method based on soil dynamic studies to estimate the change in effective θ_i due to an initial slow rainfall ($r < K_s$). The method is similar to that for redistribution, and is based on the soil properties and the relative value of rainrate (Corradini et al., 1994). Further, there may be a short initial pulse of higher rainrates which would produce runoff, but does not last long enough, and a hiatus intervenes. For this, a similar method is used to estimate the resulting wetting depth and θ_0 that best describes the water then to be redistributed as described above. All of these methods insure that a more accurate estimate of runoff from any rainfall pattern is obtained.

Expanded Infiltration Expressions.

Very Wet Initial Conditions. After a relatively short period of redistribution, the soil profile will be relatively wet, which creates a condition that violates the assumptions of most infiltration models. Not only is the initial water content of the upper soil quite high, but the total infiltration flux must include the relatively steady flux of the water already in the profile. A modification of Eq. (1) (Smith *et al.*, 1993) deals relatively accurately with wet initial conditions. In the wet case, $I' = (I - K_i t')$ is used for I, where K_i is the hydraulic conductivity of the initial soil profile, $K(\theta_i)$, and t' is the time from start of wetting or rewetting.

Heterogeneous Areas The variation of K_s over an area in its interaction with the value of r determines the apparent overall areal value of K_s. When K_s has a random distribution, especially considering the tail of the distribution, there will for a finite r always be some points having $K_s >$

r. The portion of area thus not contributing to runoff will increase with decreasing r, and the result will be an areal effective K_s, which we will call K_e, that increases with r. This value can be expressed as follows:

$$K_e = \left[1 - P_k(r)\right] + \int_0^r k p_k(k)\,dk \tag{13}$$

in which P_k is the CDF of K_s, and p_k is the corresponding PDF. Figure 2 illustrates the relation between r_*, K_e and $Cv(K_s)$ for a lognormal PDF.

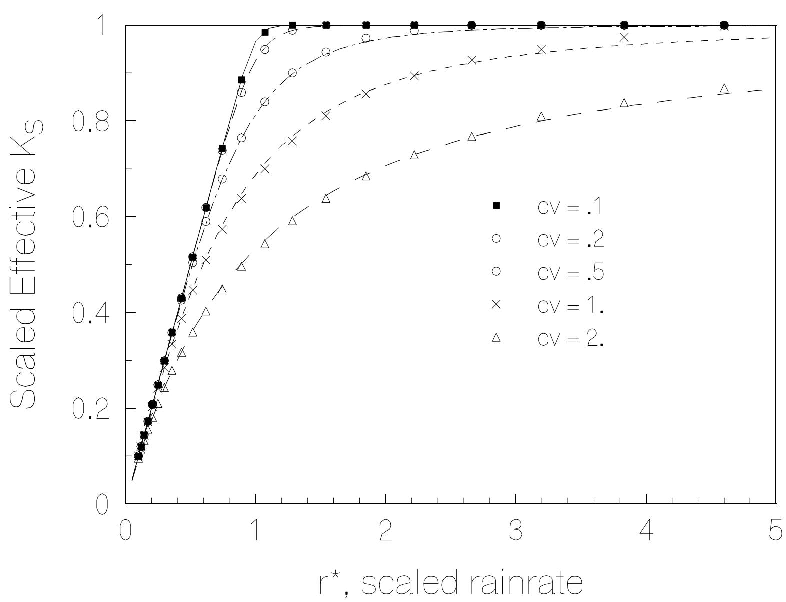

Figure 2. . Relation of effective ensemble K_s to coefficient of variation of K_s and scaled rainrate.

Simulation experiments with a runoff surface whose values of K_s are lognormally distributed have been performed to establish a relationship for infiltration of the heterogeneous surface as a whole. Define $\overline{K_s}$ in Eqs (2) and (4) as the areal expected value or mean of K_s, and define $Cv(K_s)$ as the coefficient of variation, or $(K_s)/K_e$, where () is the standard deviation. Numerical experiments have demonstrated that the surface ensemble behavior can be described by an equation based on Eq. (5) which can be written in scaled terms (based on K_e) as

$$\frac{f_{c*}-1}{r_*-1}=\left\{1+\left[\frac{r_*-1}{\alpha}\left(exp\left(\alpha I_*\right)-1\right)\right]^C\right\}^{-\frac{1}{C}} \tag{14}$$

in which C is a parameter greater than 1, whose values are directly related to the scaled rainrate r_* $[r / K_e]$ and the $Cv(K_s)$. Figure 3 illustrates the estimation of C as a function of $Cv(K_s)$ and r_*.

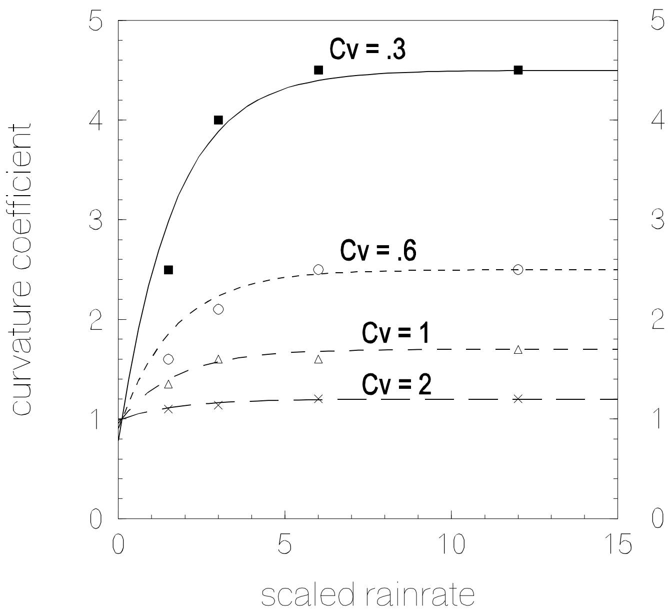

Figure 3. Variation of the coefficient C with scaled rainrate and $Cv(K_s)$

The estimating equation shown in Figure 3 for $C(r_*,Cv)$ is

$$C=1+\left[0.75(Cv)^{-1.3}\right]\left[1-exp\left(0.1-r_*\right)\right] \tag{15}$$

and the estimating equation for K_e shown in Figure 2 is

$$K_e=\left[1+\left(\frac{1}{r_*}\right)^p\right]^{-\frac{1}{p}}$$

$$p=0.7+\frac{1.2}{Cv^{1.2}} \tag{16}$$

Rewetting Conditions The conductivity of the water initially in the soil, K_i, is normally negligibly small and is then neglected. For rewetting conditions, however, K_i and G are estimated by the model based on the soil parameter λ and the value of θ_i, and infiltration is treated as described in 5.1 above. Rewetting parameters apply until the rewetting "front" moves to the depth of the original wetted front. After that, original values of θ_i and G apply, as the original wetting "front" is rejoined (for more details see Corradini *et al*., 1994).

Estimation of New Parameter (λ). The additional parameter for water redistribution in KINEROS2 that is not required in KINEROS is the pore size distribution index, λ. Table 1, below, is taken from the statistical study of soils by texture of Rawls *et al.* (1982). This provides a guideline for user estimation of the new parameters. Note that there is significant variation within textural classes, and that, while λ is consistently larger for coarser soils and smaller for clays, there is no consistency for soils with medium textures. It should be remembered that as a *distribution* index, this parameter should be expected to be smaller for soils with a large range of particle sizes, and larger for soils which are relatively uniform in particle size, whatever the mean size may be. This table also provides some guidance for estimating S_{max}, by using the ratio of θ_s to porosity, n.

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