Modeling Subsurface Stormflow on Steeply Sloping Forested Watersheds

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Five mathematical models for predicting subsurface flow were compared to discharge measurements made by Hewlett and Hibbert (1963) on a uniform sloping soil trough at the Coweeta Hydrologic Laboratory. The models included one- and two-dimensional finite element models based on the Richards equation, a kinematic wave model, and two simple storage-discharge models based on the kinematic wave and Boussinesq assumptions. The simple models simulated the subsurface response and water table positions as well as the more complex models based on the Richards equation and were much more economical to use from the point of view of computational costs. Such models have features that would allow them to be incorporated into more complex watershed models, thus placing hydrologic prediction on a more physically correct and less empirical footing.

INTRODUCTION

The Hortonian concept of runoff generation [Horton, 1933] due to surface saturation from above (i.e., overland flow generated when the rainfall intensity exceeds the soils infiltrability) has formed the basis of the runoff generation algorithms used in most watershed models developed to date. However, it is now recognized that Horton's infiltration theory represents only one extreme of the spectrum of mechanisms involved. Hursh [1936] identified the other extreme as being subsurface flow. Many field studies have been conducted where both of these extremes have been observed individually and in combination to varying degrees [e.g., Hewlett, 1961; Hewlett and Hibbert, 1963; Whipkey, 1965; Dunne and Black, 1970; Beasley, 1976; Pilgrim and Huff, 1978; Pilgrim et al., 1978; Corbett, 1979; Mosley, 1979].

Freeze [1972] and Smith and Hebbert [1983] identified the dominant mechanisms in the generation of streamflow as being overland flow, saturated near-surface flow (subsurface flow), and deep aquifer flow. Overland flow may occur because of saturation from above (Hortonian excess) or because of saturation from below and the resulting exfiltration (fed by infiltration and lateral subsurface flow). Hortonian overland flow occurs on soils exhibiting low infiltrabilities, such as on agricultural land, unvegetated surfaces, and in deserts and urban areas. Application of this concept to steeply sloping forested watersheds in humid climates is often inappropriate, since observance of this mechanism is extremely rare in such areas [Whipkey, 1965; Kirkby and Chorley, 1967; Mosley, 1979; Sloan et al., 1983]. Overland flow due to saturation from below in near-stream regions is the primary mechanism at work in watersheds whose hydrologic response fits Hewlett's [Hewlett and Nutter, 1970] variable (or partial) source area theory.

Subsurface flow is likely to be significant in watersheds with soils having high hydraulic conductivities and an impermeable or semipermeable layer at shallow depth that can support a perched water table. Such conditions often occur in humid forested watersheds where the organic litter protects the mineral soil and maintains high surface permeabilities, thus promoting high percolation rates to the A and B horizons. The upper soil profile of forested watersheds is often interlaced with roots, decayed root holes, animal burrows, worm holes, and other structural channels making a highly permeable medium for the rapid movement of water in all directions. When percolating water moving vertically in such a medium reaches an impermeable layer, lateral subsurface flow is generated [Whipkey 1965, 1967; Weyman, 1970; Pilgrim and Huff, 1978; Pilgrim et al., 1978; Corbett, 1979; Mosley, 1979]. Under such conditions water movement occurs in two domains: through the micropores (i.e., the soil matrix) and through the interconnected macropores of the soil system [Beven and Germann, 1982]. The terms micropore and macropore, as used here, refer more to the dynamics of channeling flow within the soil profile [Beven, 1981] than to strict pore sizes. Skopp [1981] proposed the following descriptors of these terms which are applicable in the present context: "macroporosity—designates the pore space which provides preferential paths of flow so that mixing and transfer between such pores and remaining pores is limited," and "matrix porosity (i.e., microporosity)—designates that pore space which transmits water and solute at a rate slow enough to result in extensive mixing and relatively rapid transfer of molecules between different pores." The response time of flow within these two domains to rainfall, as reflected in the runoff hydrograph, is quite different. Where a significant portion of the total flow takes place in the macropores, the response time of subsurface flow to rainfall approaches that of overland flow, giving rise to a high perceived hydraulic conductivity for the soil profile as a whole [Whipkey, 1965, 1967; Mosley, 1979]. The response time of flow within the soil matrix is much slower, and except under special conditions such flow may provide only a small contribution to the total stormflow response of a watershed [Freeze, 1972].

Field studies of subsurface stormflow have shown that the direct application of the concepts of saturated and unsaturated Darcian flow to water flow in forested watersheds may not be realistic [Whipkey, 1965; Weyman, 1970; Pilgrim et al., 1978; Mosley, 1979; and Sloan et al., 1983]. Beven and Germann [1982] state that "there have been no significant theoretical advances in modeling hillslope flows that involve mac-
ropores,” and that “at the current time, attempts at modeling subsurface stormflows involving macropores must be highly speculative.” At present there are at least three deterministic methods of representing the turbulent nature of subsurface stormflow. Barcelo and Nieber [1982] used pipe-flow hydraulic equations coupled with the Richards equation to model the contribution from the soil pipe network (macropore system) and the soil matrix separately. The difficulty in this approach arises in defining the soil pipe network, which is very heterogeneous in the forest environment.

Another approach is to modify Darcy’s equation to account for turbulent flow. Whipkey [1967] cited several attempts to do this, such as adding a second-order term to Darcy’s equation. This approach of modifying Darcy’s equation has not been entirely successful. Most modifications have been developed using laboratory data and apply only to specific porous media conditions. Therefore general application to highly permeable shallow forest soils is not realistic [Whipkey, 1967].

A third approach of modeling subsurface stormflow uses Darcy’s equation and effective soil water parameters. These parameters, in effect, are averaged over the soil profile, removing the heterogeneous nature of forest soils and soil structure. Typically, the effective saturated hydraulic conductivity using such an approach may be an order of magnitude greater than that measured for the soil matrix using small undisturbed soil cores.

This paper compares five physically based subsurface flow models with a view toward evaluating their suitability for modeling the stormflow response of steeply sloping forested watersheds in the humid region. The models were compared using drainage discharge data from an inclined soil trough measured at the Coweeta Hydrologic Laboratory in the early 1960’s.

**DESCRIPTION OF THE SUBSURFACE FLOW MODELS**

Five subsurface flow models of varying complexity were considered in the study, all of which can be classed as deterministic conceptual type models [classification from Clarke, 1973]. These models are a two-dimensional finite element model [Nieber, 1979; Nieber and Walter, 1981] and a one-dimensional finite element model [Nieber, 1982] based on the Richards equation, a kinematic wave subsurface flow model [proposed by Beven, 1981a, 1982], and two simple storage-discharge models termed the kinematic storage model (uses the kinematic approximation in its derivation) and Boussinesq storage model (uses the Boussinesq assumptions in its derivation) [Sloan et al., 1983, also unpublished manuscript, 1983]. These models represent a range of mathematical sophistication, with the two-dimensional finite element model being the most sophisticated and the simple storage models being the least sophisticated. All five models represent subsurface flow in a two-dimensional cross section along a flow path down a steep hillslope.

**Nieber’s Two- and One-Dimensional Finite Element Models**

The two finite element models are based on the Richards equation (derived from Darcy’s equation and the mass continuity equation) for flow in saturated and unsaturated porous media, which in two-dimensional form is

\[
\frac{\partial \theta}{\partial t} = C(h) \frac{\partial H}{\partial t} = \frac{\partial}{\partial x} \left[ K(h) \frac{\partial H}{\partial x} \right] + \frac{\partial}{\partial z} \left[ K(h) \frac{\partial H}{\partial z} \right]
\]  

where \( \theta \) is the volumetric water content; \( C(h) \) is the specific water capacity (= \( \partial \theta / \partial h \)); \( K(h) \) is the unsaturated hydraulic conductivity; and \( x \) and \( z \) are distances in the horizontal and vertical directions, respectively. The parameters \( \Theta, C, \) and \( K \) are highly dependent on \( h \), and the models use the following expressions, proposed by Verma and Brutsaert [1971], to define these relationships:

\[
\theta(h) = m \left[ \frac{A}{A + h} \right] + \theta_s \tag{2}
\]

\[
C(h) = \frac{mBh^{n-1}}{(A + h)^2} \tag{3}
\]

\[
K(h) = K_0 \left[ \frac{\theta - \theta_s}{\theta_i - \theta_s} \right] \tag{4}
\]

where \( m \) is the effective porosity (= \( \theta_i - \theta_s \)); \( \theta_s \) is the saturated volumetric water content; \( \theta_i \) is the residual volumetric water content; \( K_0 \) is the saturated hydraulic conductivity; and \( A, B, \) and \( N \) are constants determined from the soil water characteristics of the soil.

Equation (1) reduces to a one-dimensional form by taking the one-dimension parallel to the hillslope gradient and assuming hydrostatic conditions normal to this gradient (i.e., assuming no flow normal to the hillslope gradient). The resulting one-dimensional Richards’ equation [Nieber, 1982] is

\[
DC(h) \frac{\partial H}{\partial t} = \frac{\partial}{\partial x} \left[ K(h) D \cos^2 \alpha \frac{\partial H}{\partial x} \right] + i \tag{5}
\]

where \( D \) is the vertical soil depth, \( \alpha \) is the angle of the impermeable bed to the horizontal, and \( i \) is the rate of water input to the saturated zone from the unsaturated zone normal to the surface (from a coupled model of flow from the unsaturated zone).

For both finite element models the finite element approach is applied to the space domain and a fully implicit backward finite difference scheme is used for the time domain. Discretization of the flow region is achieved using triangular and linear elements in the two- and one-dimensional models, respectively. More complete descriptions of the two- and one-dimensional models are presented by Nieber [1979], Nieber and Walter [1981], Nieber [1982], and Sloan et al. [1983].

**Kinematic Wave Model**

The kinematic wave approximation of saturated subsurface flow [Beven, 1981a] assumes that the flow lines in the saturated zone above the impermeable boundary or bed are parallel to the bed and that the hydraulic gradient equals the slope of the bed. Therefore

\[
q = K_s H_s \sin \alpha \tag{6}
\]

\[
C \frac{\partial H_s}{\partial t} = -K_s \sin \alpha \frac{\partial H_s}{\partial x} + i \tag{7}
\]

where \( q \) is the discharge per unit width, \( H_s \) is the thickness of the saturated zone above the impermeable boundary, and \( i \) is the rate of water input to the saturated zone from the unsaturated zone. The specific water capacity at saturation, \( C_s \), is equivalent to Beven’s [1981a] effective storage coefficient. Equation (7) is a linear kinematic wave equation. In a later paper Beven [1982] allowed \( \theta_s \) and \( K_s \) to vary with depth. Beven’s criterion for the applicability of the kinematic wave approximation to subsurface flow is that \( \lambda < 0.75 \), where
\[ \lambda = 4i \cos \alpha / (K_s \sin^2 \alpha) \]. As an example, this criterion indicates that for a slope of 50% and \( i = 10 \text{ mm/h} \), the approximation would be acceptable for \( K_s > 140 \text{ mm/h} \).

**Simple Storage-Discharge Models**

The two simple storage-discharge models developed by Sloan et al. [1983] are based on a water balance (i.e., uses the mass continuity equation) with the entire hillslope segment being the control volume. The idealized hillslope segment has an impermeable boundary or bed, of slope \( \alpha \) and length \( L \), and a soil profile of constant thickness, \( D \), as is shown in Figure 1. Neglecting evaporation, the mass continuity equation can be expressed in mixed finite difference form as

\[
\frac{S_2 - S_1}{t_2 - t_1} = iL - (q_1 + q_2)/2
\]  

(8)

where \( S \) is the drainable volume of water stored in the saturated zone per unit width, \( t \) is time, \( q \) is the discharge from the hillslope per unit width, \( i \) is the rate of water input to the saturated zone from the unsaturated zone per unit area, and subscripts 1 and 2 refer to the beginning and end of the time period, respectively.

The kinematic storage model assumes that the water table has a constant slope between the upslope and downslope boundaries of the sloping soil mass, as is shown in Figure 1a, and that the hydraulic gradient equals the slope of the impermeable bed (the kinematic assumption). The drainable volume of water stored in the saturated zone of the hillslope is

\[
S = H_0 \theta_d L/2
\]  

(9)

where \( H_0 \) is the saturated thickness normal to the hillslope at the outlet, and \( \theta_d \) is the drainable porosity of the soil. At the outlet, \( q = H_0 v \), where \( v = K_s \sin \alpha \), which combined with (8) and (9) allows the hydraulic head at the outlet at the end of each time increment, \( \Delta t \), to be expressed explicitly as

\[
H_{02} = \left[ H_{01}(L \theta_d - v \Delta t) + 2Li\Delta t / (L \theta_d + v \Delta t) \right]
\]  

(10)

When the saturated zone rises so that the water table intersects the soil surface, (9) and (10) must be modified, so that

\[
S = D \theta_d (L + L_s)/2
\]  

(11)

\[ q = iL_s + Dv \]  

(12)

where \( L_s \) is the saturated slope length.

The Boussinesq storage model assumes that the water table has a constant slope (as shown in Figure 1b) and that the hydraulic gradient is equal to this slope (Boussinesq assumption), so that

\[
v = K_s \sin \beta
\]  

(13)

where \( \beta \) is the angle of the water table to the horizontal. At the outlet, \( q = Dv = DK_s \sin \beta \). The drainable volume of water stored in the saturated zone of the hillslope is

\[
S = \frac{D^2 \theta_d}{2 \tan (\alpha - \beta)} \quad \text{for} \quad \tan (\alpha - \beta) < D/L
\]  

(14)

or

\[
S = L \theta_d \left[ D - \frac{L}{2} \tan (\alpha - \beta) \right] \quad \text{for} \quad \tan (\alpha - \beta) > D/L
\]  

(15)

Equations (13) and either (14) or (15) can be substituted into the mass continuity equation (equation (8)) and solved iteratively for \( \beta \), thus yielding the hillslope discharge \( q = DK_s \sin \beta \) at successive times.

**Fig. 1.** Conceptual representation of the hillslope segment for (a) the kinematic storage model and (b) the Boussinesq storage model.
Coupled Vertical Water Input From the Unsaturated Zone

The one-dimensional finite element, kinematic wave, kinematic storage, and Boussinesq storage models require a coupled model to account for the vertical flow of water from the unsaturated zone to the saturated zone during wetting and drying events. For the kinematic wave model a piston flow approach, similar to that described by Beven [1982], was adopted for simulating the movement of wetting and drying fronts and hence input to the saturated zone.

In the forms used in the study reported herein the one-dimensional finite element and the kinematic and Boussinesq storage models all assume that the vertical input rate to the saturated zone is a function of the volume of water stored in the unsaturated zone, i.e.,

\[ i = K(\theta_u) \]  \hspace{1cm} (16)

where \( i \) is the vertical input rate to the saturated zone of the hillslope, \( \theta_u \) is the average volumetric water content in the unsaturated zone, assuming that the entire unsaturated region can be lumped together as a homogeneous unit. Equation (16) is a reasonable assumption if gravity dominates so that \( dH/dz = 1 \) [Davidson et al., 1969; Black et al., 1970]. The water content in the unsaturated zone is simulated using a water balance approach

\[ \theta_{z_2} = (u_a U_a + L \Delta \theta(r - \theta))/U_a \]  \hspace{1cm} (17)

where \( U_a \) is the total volume of the unsaturated zone, \( r \) is the precipitation rate, the subscripts 1 and 2 refer to the beginning and end of the time increment, respectively, and the other terms are as previously defined.

Model Boundary Conditions

The boundary conditions for each of the models were adopted to approximate the conditions that would occur on a hillslope with an impermeable bed and so that they would be consistent with the assumptions on which the models are based. They consisted of either specified flux type (Neumann) or specified head type (Dirichlet) boundaries.

Boundaries CD and DA (Figure 1). All five models assume no-flow boundaries:

\[ \frac{\partial H}{\partial z} = 0 \quad z = 0 \quad 0 < x < L \]  \hspace{1cm} (18)

\[ \frac{\partial H}{\partial x} = 0 \quad x = L \quad 0 < z < D \]  \hspace{1cm} (19)

Boundary BC (Figure 1). The two-dimensional finite element model can handle either a no-flow boundary or a seepage boundary, i.e.,

\[ \frac{\partial H}{\partial x} = 0 \quad x = 0 \quad 0 < z < D \]  \hspace{1cm} (20)

or

\[ H = e \quad h = 0 \quad x = 0 \quad 0 < z < D \]  \hspace{1cm} (21)

with

\[ H = D \quad h = 0 \quad x = 0 \quad z = D \]  \hspace{1cm} (22)

The one-dimensional finite element and Boussinesq storage models assume a seepage boundary (equations (20) and (22)). The kinematic wave and kinematic storage models also assume a seepage boundary, but with

\[ \frac{\partial H}{\partial x} = 0 \quad x = 0 \quad H(t) < z < D \]  \hspace{1cm} (23)

\[ \frac{\partial H}{\partial x} = 0 \quad x = 0 \quad H(t) < z < D \]  \hspace{1cm} (24)

in which \( H(t) < D \).

Boundary AB (Figure 1). The two-dimensional finite element and kinematic storage models assume an infiltration and seepage boundary:

\[ -K(\theta) \frac{\partial H}{\partial z} = r \cos \alpha < K, \quad z = D \quad L_s < x < L \]  \hspace{1cm} (25)

\[ H = e \quad h = 0 \quad z = D \quad 0 < x < L_s \]  \hspace{1cm} (26)

where \( L_s \) is the length of the saturated seepage face along AB.

The two-dimensional finite element, kinematic wave, and Boussinesq storage models all assume an infiltration boundary:

\[ -K(\theta) \frac{\partial H}{\partial z} = r \cos \alpha < K, \quad z = D \quad 0 < x < L \]  \hspace{1cm} (27)

The two-dimensional finite element model can handle either a seepage or no-flow boundary along the boundary BC (Figure 1) because it formulates the relevant water flow equations in two dimensions. All the other models are one-dimensional approximations of the two-dimensional flow regime and therefore can only handle a seepage face along boundary BC.

SIMULATION OF DRAINAGE FROM A SLOPING SOIL BED: THE COWEETA EXPERIMENT

The five subsurface flow models were examined using published drainage discharge data from a soil trough measured by Hewlett and Hibbert [1963] at the Coweeta Hydrological Laboratory in the southern Appalachian mountains. The soil trough consisted of a 0.92 x 0.92 x 13.72-m concrete-lined trough constructed on a 40% slope and filled with recompacted C horizon forest soil (Halewood) that was excavated nearby (Figure 2). Instrumentation included tensiometers, piezometers, and access tubes for nuclear moisture meter
measurements. Discharge was measured using a water level recorder in a tank at the base of the trough. The soil was soaked using sprinklers, covered with plastic to prevent evaporation, and then allowed to drain.

The Coweeta study is of practical interest because it provides data that can be used to evaluate the ability of subsurface flow models to simulate porus media flow in shallow soil overlying a steeply sloping impermeable bed. Because the soil used in the soil trough was mixed and compacted in the bed of the trough, the effects of macropores, such as root and worm holes and animal burrows, on subsurface flow can not be evaluated using Hewlett and Hibbert's [1963] data. Therefore these data are not directly applicable to undisturbed forested watersheds. However, they do provide a basis for testing the ability of different models to simulate that portion of subsurface flow that occurs within and through the soil matrix (as opposed to the macropores) in steeply sloping watersheds. Water movement in homogeneous soils with no macropores is the simplest physical subsurface flow system to represent mathematically. Therefore because Hewlett's drainage discharge data represents the simplest "ideal" condition, use of this information is a logical place to begin testing, validating, and/or developing physically based models of subsurface flow.

**Soil Water Characteristics**

All input parameters for the five models are physically based and measureable and were determined a priori from measurements reported by Hewlett [1961] and Hewlett and Hibbert [1963]. No fitting of the models or optimization of model parameters was carried out. The constants in (2), (3), and (4) for characterizing the soil water properties were fitted to the soil water characteristic data measured by Hewlett ([1961], A = 1.76, B = 0.36, and N = 14.63). A Ks = 168 mm/h, θe = 0%, and θr = 49% by volume were used in the models [Hewlett, 1961; Hewlett, personal communication, 1981; Hewlett and Hibbert, 1963]. Hysteresis was not considered in the simulations because the soil trough was wetted to saturation before being allowed to drain, thus allowing the drying curve of the soil water characteristic to be used without error.

**Initial Hillslope Conditions**

It was assumed that steady state discharge conditions (0.692 m²/d/m) existed before drainage of the soil profile began. This was approximated in each of the models by applying an input precipitation rate of 2.1 mm/h normal to the soil surface (≈ 0.692 m²/d/m) until steady state conditions were achieved and then allowing the system to drain with no precipitation input.

The boundary conditions assumed for each of the models were as described previously, with the two-dimensional finite element model assuming a no-flow boundary along boundary BC with a fixed head at B (Figure 1) corresponding to the outlet level shown in Figure 2. The lower end of Hewlett and Hibbert's [1963] soil trough, near the outlet and below the water table (Figure 2), consisted of sand graded into large gravel and was somewhat diffuse. Therefore the approximation of the actual flow region (Figure 2) and boundary conditions by a rectangular section (Figure 1) and the boundary assumptions described above should produce little error in the predicted discharges and only small errors in the position and shape of the predicted water tables in the immediate vicinity of the outflow from the soil trough.

**Results**

The drainage hydrographs and cumulative runoff curves predicted by the five subsurface flow models are presented in Figures 3 and 4, respectively. The drainage hydrograph measured by Hewlett and Hibbert [1963] is also presented in Figure 3 for comparison. It can be seen that all the models, with the exception of the kinematic wave model, predict the drainage hydrograph reasonably well. However, none of the models predict the extended high flows for 1,000 < t < 3,000 min. It is during this time that the predicted water tables fall rapidly. On the basis of Beven's [1981] criterion, the kinematic wave assumption appears to be applicable to the present case. Beven's criterion indicates that it is valid if i < 4.7 mm/h. The maximum rate of water input to the saturated zone from the unsaturated zone was i = 2.1 mm/h at t = 0 min. Reeves and Duguid [1975] also present a comparison of a two-dimensional element model similar to the one used here (but using quadrilateral, rather than triangular elements) and the Coweeta data and report similar results to those presented here.

The results presented by Hewlett and Hibbert [1963] were inadequate to reliably define the transient positions of the water tables, and so the two-dimensional finite element model predictions were used as the standard for comparison to the other models. The transient water table positions predicted by the five models at t = 0 and t = 1,000 min from the beginning of drainage are presented in Figure 5. The steady state water tables (t = 0 Figure 5a) for the two finite element models based on the Richards equation and the models using the kinematic wave assumption show very close agreement for distances greater than 2 m upslope. However, 2 m upslope the predictions of the other models predict the extended high flows for 1,000 < t < 3,000 min. The Boussinesq storage model predicts a water table that is at about the same slope as the one- and two-dimensional finite element models at the outlet. However, 2 m upslope from the outlet it deviates significantly from the predictions of the other models.

![Fig. 3. Comparison between observed [adapted from Hewlett and Hibbert, 1968] and predicted drainage hydrographs for Nieber's two- and one-dimensional finite element models, the kinematic wave model, and the kinematic and Boussinesq storage models for Ks = 168 mm/h.](image-url)
models. At $t = 1,000$ min (Figure 5b) the water tables generated by the finite element models and the kinematic storage models show good agreement with each other for $x > 3$ m. The kinematic wave model overestimated the saturated zone depth by comparison to these. This is because of the overestimation of the water input to the saturated zone from the unsaturated zone rather than the kinematic assumption being in error. The water table predicted by the Boussinesq storage model was almost horizontal at $t = 1,000$ min.

Table 1 presents a summary of the coefficients of determination ($0 < t < 7,000$ min), timing errors, simulation costs, and computer core storage requirements for each of the models. These results show that the more complex finite element models require extensive computer resources, while those required by the simpler models are very small. The more sophisticated one- and two-dimensional finite element models based on the Richards equation were the most accurate at small times ($0 < t < 1,000$ min), while the simple storage models performed better at large times ($5,000 < t < 50,000$ min).

Discussion of Results

The infiltration model that was coupled with the one-dimensional finite element, the kinematic wave, and the two storage models had a large impact on the predicted drainage hydrographs, particularly for $t > 3,000$ min when the capillary forces began to dominate the hillslope response. If vertical drainage from the unsaturated zone is ignored in the models, then the predicted discharge drops off rapidly to close to zero for $t > 3,000$ min.

The piston drying front model used in conjunction with the kinematic wave model did not perform as well as the algorithm used in the other models, because input from the unsaturated zone continued at a constant rate until the drying front reached the saturated zone (hence the reason for the elevated discharges for $600 < t < 2,500$ min). Both coupled infiltration models overestimated vertical input from the unsaturated to the saturated zones early in the simulation, but the kinematic wave model produced the worst results, for the reasons given above.

The results of the comparison demonstrate that simple subsurface flow models that make assumptions consistent with the physical processes can be as effective as the more sophisticated models (i.e., the one- and two-dimensional finite element models) in predicting hillslope discharge (kinematic and Boussinesq storage models) and the extent and position of the saturated zone (kinematic wave and kinematic storage models). However, because of the assumptions made in these simple models (in their present forms) it should be realized that they cannot simulate the water content gradients in the unsaturated zone.

The kinematic storage model coupled with the simple infiltration model, assuming gravity drainage in the unsaturated zone, produced very satisfactory results for the Coweeta study. Such simple algorithms have potential for incorporation into more comprehensive distributed watershed models and may provide a means for removing much of the empiricism surrounding the development and application of these types of watershed models.

Based on preliminary field observations and a preliminary analysis of field data by the authors [Sloan et al., 1983], together with field observations reported by Whipkey [1965, 1967], Weyman [1970], and Mosley [1979], it is postulated that there are two subsurface flow components contributing to
the hydrologic response of steeply sloping forested watersheds: (1) macropore flow, which is responsible for the stormflow response; and (2) soil matrix flow, which is responsible for base flow or the delayed response. Where such flow conditions occur the finite element models based on the Richards equation would be expected to simulate storm flow poorly because the diffusion-type flow assumptions inherent in the models would be violated. Under such conditions the kinematic wave and the two storage-discharge models may predict the stormflow response adequately if average or effective hillslope hydraulic conductivities were used. However, further research is necessary to answer these questions and to evaluate subsurface flow models under actual field conditions where the two-domain system consisting of macropores and the soil matrix is likely to contribute to the overall hillslope hydrologic response.

CONCLUSIONS

Simple physically based models were shown to adequately simulate the stormflow response of a steep-sloped forested watershed in comparison to observed discharges and those predicted by more complex models based on the Richards equation. They are also the most economical to use from the point of view of computational costs. Such models have the features that would allow them to be incorporated into more complete watershed models. If this were done, then the prediction of the hydrologic response of undisturbed forested watersheds would be placed on a more rational, physically correct, and less empirical footing.

Results of the simulation comparisons raise questions about the process of validating subsurface flow models. The complex finite difference and finite element models are often used as a reference against which to evaluate simpler models and are often used to demonstrate how flow systems respond to rainfall. The results presented in this paper indicate that field verification is essential because the complex models may not be very good standards for testing and validating other models. The assumptions inherent in the development of such models can be as limiting as the assumptions involved in the simpler models. In addition, the predictive accuracy of the simpler models, such as the kinematic storage model, were shown to be as good or better than that of the complex models.

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