VARIABLE SOURCE AREAS AND STORM-FLOW GENERATION: A UPDATE OF THE CONCEPT AND A SIMULATION EFFORT

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ABSTRACT


The variable source area concept of streamflow and storm-flow generation proposes that the area of land yielding surface water to streamflow varies with time and that vegetated basins, subsurface flow not only sustains baseflow but is also a major component of storm flow. Results from a growing number of trench and tracer studies strongly support the concept. VSAS2 is a deterministic storm-flow simulator for small forested basins based on the variable source area concept. Basins are divided into a number of sub-basins or segments. In each segment, subsurface flow is reduced to a two-dimensional flow problem while the third dimension is represented by segment convergence divergence to the stream. The irregular, time-varying grid attempts to represent the variable channel system, while keeping grid size within computational feasibility. Mathematical stability of the explicit solution is secured by proper combinations of space and space increments. The performance of VSAS2 on a 24 ha Georgia Piedmont basins poor for large winter storms and small summer storms. Discrepancies are traced to an inadequate representation of both micro-relief and soil-water properties.

INTRODUCTION

Classical hydrology crystallized at a time when upstream basins were viewed mainly as contributors to streamflow and storm flows in downstream developed areas. Horton's overland-flow theory (Horton, 1945) allows engineers and land managers to estimate discharge from these basins with sufficient accuracy to satisfy downstream engineering demands. However, classical hydrology could not explain the sources, pathways and turnover rates of water on forested or other vegetated areas where high infiltration rates precluded widespread overland flow.

Hewlett (1961), following the work of Dr. Charles Hursh and his workers at the Coweeta Hydrologic Laboratory, proposed an alternative...
concept of streamflow and storm-flow generation for natural basins. Developed in subsequent papers (e.g., Hewlett and Hibbert, 1963, 1967) these ideas became known as the variable source area concept of streamflow generation. The contrast with previous theories can be summarized in the following points:

1. The area of a drainage basin contributing directly to streamflow varies with time.
2. In vegetated basins, subsurface flow supplies all baseflow and is also a major contributor to storm flow.

The objective of this paper is first to analyze known field evidence on storm-flow origin in the light of this concept, and, secondly, to present a simulator where terrain representation attempts to emulate variable source area processes. Hydrological terms in this paper are used as defined in Hewlett (1982).

UPDATING THE VARIABLE SOURCE AREA CONCEPT

Investigations in storm-flow origins

Two types of techniques have been used in a direct attempt to distinguish between surface and subsurface storm flow. Both are imperfect. The first type is based on the interception of surface and subsurface flow from hillside plots, the trench approach. The second type involves the use of tracers.

The trench approach is based on the premise that the rise and fall of the storm flow in the stream channel is related to the variations of surface and subsurface flow through a cross-section of the slope uphill from the channel. Typically, a hillside plot is subtended by a trench equipped with several troughs intercepting flow from particular soil layers. To relate flow measurements to the runoff process, the analyst must overcome three major hurdles.

First, the digging of the trench and the placement of water collection devices below the ground profoundly affect the potential field in the surrounding soil mantle as pointed out by Atkinson (1978). The second problem is that the experiment yields no direct information on the origin and pathways of either the collected water or the water in the stream. The final problem is that hillside plots cannot reproduce the basin-wide dynamics of source areas, thus defeating any direct interpretation in terms of natural streamflow. These three problems are inherent to the trench technique, regardless of the care with which the flow-collection systems are designed.

Nevertheless trench studies generally show large amounts of subsurface flow on hillslopes (Whipkey, 1965; Weyman, 1970; Troendle and Homeyer, 1971). How much of this becomes storm flow requires additional evidence.

The closer the trench is constructed to a live stream, the more its hydrograph will relate to the basin hydrograph (Atkinson, 1978). However, initial rapid contributions from shallow layers or narrow stream-side bands are sometimes perceived as proportional to storm flow from the basin. Figure 1 shows a hydrograph produced on a 0.12 ha hillside plot at Sleepers River, Vermont, after application of 29 mm of water (fig. 8 of Dunne and Black, 1970). The speed of the initial response, originally perceived as the entire storm hydrograph (solid line), implied that most storm-flow contribution came from surface storm flow. However, the application of a storm-flow classification technique (Hewlett and Hibbert, 1967) suggests that the total storm hydrograph must have included a second pulse of greater duration and volume of possible subsurface origin. The total storm-flow response (classified storm flow/rainfall) of about 13% obtained after extrapolation compares well with responses computed by Engman (1981) under similar rainfalls for the 44 km² Sleepers River basin. This particular case illustrates well how plot studies can lead to misinterpretations of storm-flow sources and pathways on whole basins.

Studies using the natural tracers oxygen-18 (Sklash et al., 1976) and tritium (Atkinson, 1978) offer in theory a better approximation to actual delineation between surface and subsurface storm flow. These two tracers, present at different concentrations in rain water and in soil water, appear to behave like the surrounding water molecules. Analysis of tracer concentration in streamflow, rainfall and soil water should reveal the proportion of rainfall (new water) and soil water (pre-event water) in storm flow. However, pre-event water contribution is not necessarily equal to subsurface storm flow since new water can also join storm flow through short subsurface pathways.

Table 1 summarizes most of the recent experiments on storm-flow origin done with tritium and oxygen-18. Included are a few studies on snow-melt runoff where pre-event water is from earlier seasons. Results from experiments using dissolved ions as tracers (Pinder and Jones, 1969; Newbury et al., 1969; Cleaves et al., 1970; Ayers, 1976; Kennedy and Malcolm, 1977), not included in the table because of weaker inference, agree with those from elemental isotope studies. Contributions of pre-event water to storm-flow
volumes or peaks lie between 50 and 80 percent. The only exception presents preliminary conclusions based on a "... frequency of sampling (that) appeared to be inadequate" (Mook et al., 1974). The diversity of the basins involved in these studies suggests that similar results would be obtained in most vegetated areas.

These findings strongly support the view held in the variable source area concept that subsurface flow plays a major role in storm-flow generation on vegetated basins. Surface storm flow is dominant only in small storm flows. For example, it has long been known that storm flows from small storms can often be accounted for by channel precipitation alone (Hurst and Brater, 1941). But it is now apparent that the importance of surface storm flow in flooding has until recently been over-emphasized.

Streamflow on the drainage basin

Drainage basins are three-dimensional land forms that force rain or melt water to converge toward the stream. Some segments of the basin, i.e. draws and coves, favor subsurface flow convergence. Others, such as the spurs separating coves, favor divergence (Tischendorf, 1969). Basins give birth to streams at points where, due to subsurface convergence, the mantle can no longer transmit by subsurface flow all the water that the upslope areas are feeding into it. The location and extent of source areas is thus determined by a dynamic input-output equilibrium between subsurface flow and infiltration.

Rain falling on moist ground influences the energy status of the soil water to some depth with a rapidity that has always surprised observers (e.g., O’Brien, 1982). Flow increases through the already yielding seepage areas due to increased hydraulic gradients. Mounds of groundwater may form along streams, a phenomenon described as “groundwater ridging” by Sklash and Farvolden (1979). In all but the most deep and porous soils the exfiltrating areas eventually expand under increasing rainfall inputs into the coves and draws, appearing anywhere the flow transmission capacity of the mantle is exceeded. Later, as inputs to the saturated zones decreases, source areas slowly contract. Anderson and Burt (1981) discuss the source of recession outflow from hillslopes as being restricted to the drainage of the “saturated wedge” alone. However, Hewlett and Hibbert (1963) presented clear evidence from a hillslope soil model that the long-term outflow cannot be accounted for without considering the constant replenishment of the “saturated wedge” by water from the unsaturated mantle.

Figure 2 illustrates channel expansion and contraction on a 24 ha Georgia basin for a 10.5 cm summer storm (Tischendorf, 1969). The dot pattern shows non-channeled seepage waters as they are forced out onto the surface by clay layers or rock outcrops. Tischendorf also noticed that for most winter storms (wet initial conditions) the channel network kept on expanding after the rain ceased. This intricate pattern of source area expansion and
Hewlett and Troendle (1975) summed up the response of a basin to a rainstorm as follows:

"... the hydrograph rises because of channel precipitation and grows because of the expansion of the flow system into seeps and drains that are tapping shallow and deep subsurface flow paths. A dynamic storage zone expands under the steady downslope movement of soil water. This pattern expands rapidly under intense rainfall (and) shrinks slowly afterwards..."

This statement describing the storm-flow process in forests and wildlands may be extended to describe flows from fields, pastures and arid lands when it is realized that the difference is one of degree, not of kind. Shallow or non-wettable surface soils will cause very rapid channel or source area expansion under intense rain. In extreme cases, the variable source area concept converges on Hortonian overland flow theory. Studies in Arizona (Arteaga and Rantz, 1973; Lane et al., 1978), showed that the variable source area concept is not incompatible with traditional methods of analyzing runoff from semi-arid basins.

**SIMULATING BASIN STREAMFLOW**

**Background work**

Variable source area dynamics can be reproduced only in distributed-type simulators that allow non-vertical movement of water through a three-dimensional saturated—unsaturated soil mass. Freeze (1971) developed such a simulator where, in sharp contrast with classical overland flow simulators, infiltration of rain water was treated as a component of subsurface flow. In the simulations (Freeze, 1972a, b; Stephenson and Freeze, 1974) the simulator was collapsed back to two-dimensional format and no apparent allowance was made for lateral subsurface flow convergence and source area expansion.

Hewlett and Nutter (1970) and Hewlett and Troendle (1975) presented an elementary model based on variable source area dynamics. Troendle (1979) produced a first-generation variable source area simulator (VSAS1) of stormflow generation. The concept served as a means to simplify terrain and process description by identifying the dynamic zones of the basin and the important pathways and processes. VSAS1 was tested with moderate success on a West Virginia watershed. Following an unsuccessful field trial on another basin (Lefkoff, 1981), the simulator was extensively revised. This section reports the structure of the revised version (VSAS2) and results of a new field trial.

**VSAS2: A variable source area simulator**

VSAS2 is a physically based storm-flow simulator for small (first- to second-order) vegetated basins. Input requirements are basin topography,
soil mantle geometry, soil hydrological characteristics and rainfall. The key elements of VSAS2 are the geometric representation of the basin, the variable slope incrementation and the maintenance of greater sensitivity in hydrologically sensitive source areas. A detailed description of the simulator is provided in Bernier (1982). The procedure for segmentation is described in Bernier and Hewlett (1982).

The basin is first described to the simulator as a series of independent sub-basins or segments (Fig. 3). Segmentation is performed on a topographic map by the user. For each segment, mantle geometry is provided to the simulator through pairs of polynomial equations describing elevation and mantle depth as a function of distance from the stream (Fig. 4). The simulator divides each segment into increments (Fig. 5), or bands paralleling the stream, using the incrementation rule adopted by Troendle (1979):

\[ d_n = D(n/N) \]

where: \( d_n \) = horizontal distance from the stream to the upslope boundary of increment \( n \), \( D \) = total horizontal distance from stream to ridge, \( n \) = increment number, starting with increment number 1 at the stream-side, and \( N \) = total number of increments.

The total number of increments is computed internally, partially as a function of segment slope length. For example, a 150 m slope will be automatically divided into 16 increments ranging from 0.6 m of slope length at the stream-side to 18 m at the ridge. The smaller increments at the stream-side permit greater sensitivity in the normally more hydrologically active zones of the basin. Each increment is further divided into 3–5 layers of pre-determined depth to form the final volumetric unit, the soil element. Each soil element occupies the whole width of the segment. Within a segment, centers of mass of the elements form a non-orthogonal, irregular, two-dimensional grid used in computing and routing subsurface flow. The solution of the saturated–unsaturated flow problem is reached using a block-centered, finite difference, explicit scheme. Segment divergence or convergence, the third dimension, is expressed by the unequal widths of the increments.

Source areas that expanded laterally from the stream-sides have an upper edge located in increments larger than stream-side increments. But sensitivity has to be maintained at that upper edge in order to reproduce the dynamics of source area expansion and contraction. To do so, VSAS2 redivides the increments at the limit of expanded zones of surface saturation into
Fig. 6. Dynamics of grid restructuring in VSAS2 following source area expansion and contraction.

increments comparable in slope length with the original stream-side increments. The workings of this procedure are illustrated in Fig. 6.

Infiltration is treated mathematically as a component of subsurface flow. Soil hysteresis is not accounted for; only the desorption portion of the moisture-release curves has been determined on the soil samples. VSAS2 also assumes that all surface water generated within one iteration (15 min) reaches the stream during that iteration. Flows from the segments are lagged in accordance with average stream flow velocities (0.3–0.6 m s⁻¹) and cumulated into total basin outflow. These routing techniques are crude but were thought sufficient for the short distances involved.

The flow equation

In accordance with the variable source area concept, the physical domain of the simulator is the heterogeneous, saturated—unsaturated, upper 2–4 m of the soil mantle. Entrapment and compression of air bubbles, compression of water and deformation of the soil matrix are ignored. Isotropy is assumed within soil layers.

Saturated—unsaturated flow in the soil mantle is represented by Richard's modification of Darcy's law (Richards, 1931):

\[ q = K(\theta) \nabla H \]  

where: \( q \) = apparent water velocity (cm h⁻¹), \( K = \) hydraulic conductivity (cm h⁻¹), \( \theta = \) water content of the soil (cm³ cm⁻³), and \( \nabla H = \) hydraulic gradient (cm cm⁻¹), which, combined with the two-dimensional continuity equation (Hillel, 1971) becomes:

\[ \frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left[ K(\theta) \frac{\partial H}{\partial x} \right] + \frac{\partial}{\partial z} \left[ K(\theta) \frac{\partial H}{\partial z} + 1 \right] \]  

(2)

where: \( t = \) time, \( H = \Psi(\theta) + h, \Psi = \) matrix potential (cm), and \( h = \) hydraulic head (cm). The slopewise axis \( x^* \) is defined with respect to the \( x \)-axis and the slope angle \( \alpha \) (Fig. 5) as:

\[ x^* = x \cos \alpha \]  

(3)

which, when substituted in eqn. (2) yields:

\[ \frac{\partial \theta}{\partial t} = \cos^2 \alpha \frac{\partial}{\partial x} \left[ K(\theta) \left( \frac{\partial H}{\partial x} + \frac{\partial H}{\partial z} \right) \right] + \frac{\partial}{\partial z} \left[ K(\theta) \left( \frac{\partial H}{\partial z} + 1 \right) \right] \]  

(4)

Equation (4) is the final flow equation representing the non-orthogonal system. The variable of interest is the volumetric soil moisture content (\( \theta \)). When computing flow between soil elements, moisture content is accounted for in volumetric units. The cross-sectional area for flow is always computed normal to the axis of flow. The solution of eqn. (4) is obtained through a block-centered, finite difference, explicit scheme. For each non-boundary soil element, this can be expressed as:

\[ \theta_{n+\Delta t} = \theta_n + \frac{\Delta t}{V} (Q_1 + Q_2 - Q_3 - Q_4) \]  

(5)

where: \( V = \) volume of the soil element, \( \Delta t = \) time increment, and \( Q_1 - Q_4 = \) fluxes of water through the four interfaces of the soil element.

Stability of the solution was secured by using combinations of time and space increments that satisfied the stability criteria proposed in Remson et al. (1971, p. 75).

Let \( X_{n,j} \) and \( Z_{n,j} \) represent the center of mass coordinates of element \( n, j \) in the \( x^* \) and \( z \) dimensions and \( t \) be the time. Equation (5) can be formulated explicitly as:

\[ \theta_{n,j}^{t+\Delta t} = \theta_{n,j}^t + \frac{\Delta t}{V_{n,j}} \left[ \cos \alpha A_1 \left( \frac{H_{n,j}^{t+1,j} - H_{n,j}^{t,j} + Z_{n+1,j} - Z_{n,j}}{X_{n+1,j} - X_{n,j}} \right) \right. \]

\[ + A_2 \left( \frac{H_{n,j}^{t,j} - H_{n,j}^{t-1}}{Z_{n,j} - Z_{n,j}} + 1 \right) \]

\[ - \cos \alpha A_3 \left( \frac{H_{n,j}^{t,j} - H_{n,j}^{t-1,j} + Z_{n,j} - Z_{n,j}}{X_{n,j} - X_{n-1,j}} \right) \]

\[ - A_4 \left( \frac{H_{n,j}^{t,j} - H_{n,j}^{t,j+1}}{Z_{n,j} - Z_{n,j+1}} + 1 \right) \]  

(6)

where \( A_1 - A_4 \) are the cross-sectional area for flow of the four interfaces and \( K_1 - K_4 \) are the conductivities for flow through the composite layers (Hillel, 1971):
\[
K_1 = \frac{\Delta X_{n+1,j} + \Delta X_{n,j}}{K(\theta^l_{n+1,j})} \frac{\Delta X_{n+1,j} + \Delta X_{n,j}}{K(\theta^l_{n+1,j})}
\]  

Equation (6) is solved sequentially for all \( n \)'s and \( j \)'s. At time \( t + \Delta t \), all the terms on the right-hand side are known. The equation is solved for the new moisture content \( \theta^l_{n+1,j} \). Computation is initialized at \( t = 0 \) by specifying a first set of \( \theta \) and, where needed, positive pressure values. Positive pressures, initialized as equal to the difference in elevation between the channel level and the centers of mass of the elements, assume more realistic values after a few preliminary iterations. Initial boundary conditions allow flow only through the surface and stream-side face of the segments. Variations in the channel water level are not computed.

Previous reviewers have been critical of the large size of the elements and time steps used in this effort, and of the use of an explicit solution to the flow equation. Testing on individual segments indicated that halving the slope length and depth of the soil elements nearly quadrupled costs of computer runs but did not alter significantly the answers. Also, the choice to stay with the explicit approach already used in VSAS1 was, in a sense, mathematically naive because of the known superiority of implicit solutions. However, an implicit solution for the irregular, time-varying grid would have been complex to design and, within the time constraints of the study, would have detracted the author from the hydrological side of the problem. Both the large size of the soil elements (and associated long time steps) and the use of an explicit solution contributed unknown amounts of error to the simulations.

Data acquisition

The study area is a 24 ha forested second-order drainage basin, 5 km southeast of Athens, Georgia. The basin has features typical of the surrounding Piedmont terrain (Fig. 3). Slopes are gentle, ranging from 0 to 30%. Elevation above mean sea level varies from 170 to 222 m and depth to bedrock from zero along portions of the perennial channel to 30 m at some ridge top locations. Detailed description of the hydrology, geology, climate, soils and vegetation are given in Neal (1987), Tischendorf (1969) and Ayers (1976). All data used in this simulation effort, including the segmentation of the basin and characterization of the soil hydrological properties, were collected and prepared by Lefkoff (1981).

The central portion of the Whitehall basin (segments 11–16, 20, 21 and 23, see Fig. 3) is covered by soils of the Louiburg series. The remainder is occupied by soils of the Madison, Pacolet and Davidson (MPD) series. Louiburg soils are coarse and vertically undifferentiated with high measured saturated conductivities (40 cm h\(^{-1}\)). The MPD soils present a strongly layered profile in which a clayey B-horizon of low permeability separates highly permeable A- and C-horizons. Saturated conductivities of the top 0.5 m of MPD profiles equal or exceed 8 cm h\(^{-1}\).

A total of 65 undisturbed soil samples were obtained in three sets with a 5.4 x 3.0 cm brass core sampler. The first set of 26 samples was obtained at three slope locations (bottom, mid-slope, and upper slope) and four depths (15, 30, 45 and 60 cm) on one hillslope of the MPD group. The second set of 18 samples was obtained on Louiburg slope using a similar sampling scheme. The third set of 21 samples was obtained at four sites and two depths (120 and 210 cm) on MPD slopes and represents the deepest layers of that soil group. Two to three repetitive samples were taken at each site and depth.

Measurement of porosity were made on each of the samples. Saturated conductivities of the samples were determined with a constant-head permeameter and matric potential \( \Psi(\theta) \) curves were obtained on a pressure-plate apparatus (Hillel, 1971). Unsaturated conductivities \( K(\theta) \) were determined by the method proposed by Green and Corey (1971). Log-log curves were fitted to both these functions for all samples. Geometrical averaging was performed between the repetitive samples. Because of their nearly identical properties, samples from the two depths in the third sampling set were combined as repetitive samples. Finally the saturated conductivity of the clayey B-horizon (15–45 cm) of MPD samples was lowered from a measured 4 to 1 mm h\(^{-1}\) to account for the normal bias of the “undisturbed-sample” laboratory procedure. These values were then used as the only descriptors of soil hydrological properties to the simulator.

Weighted storm precipitation was estimated using one recording and four standard rain gauges evenly distributed over the basin. Streamflow was monitored with a 120° V-notch weir and recorded on a Fisher analog-to-digital water-level recorder. Events covering a wide range of initial flow rates and rainfall amounts were chosen for the simulations from eight years of continuous records.

The initialization of moisture contents was difficult because of the nonunique relation between streamflow and soil moisture distribution on the slope. During a full year in the late sixties, over 12,000 field measurements of soil moisture were taken on the Whitehall basin down to 6 m of depth with a neutron device. Tischendorf (1969) developed a series of regression equations relating these measurements to depth, height above the perennial stream, antecedent rainfall and time of the year. The coefficient of determination \( (R) \) averaged 0.70 and the standard error of estimation was large (3–4% by volume). Unfortunately, these estimates gave highly unstable solutions and had to be discarded.

Initial moisture contents were finally allocated by using the simulator itself to converge on equilibrium conditions. Simulated segments were “soaked” and drained until their outflow decreased to a pre-specified base flow. Sets of \( \theta \) values were derived for wet, medium and dry initial conditions. Measured baseflow (never zero) prior to the actual storm hydrograph determined which set of initial soil water contents would be used for the simulation.
Simulation results and discussion

Simulations were performed on an IBM 370/158, requiring on average 30 s of CPU time per segment per simulated day. Results from a sample of these simulations are plotted in Fig. 7. A summary presented in Table 2 shows that both large peaks and large classified storm flows are generally under-estimated while small classified storm flows are over-estimated.

In the simulations, Louisburg segments produced hydrologic response ratios (classified storm flow/rainfall) only half as great as MPD segments, showed no surface saturation and in general yielded very small peaks. MPD segments produced greater responses, supported some large, short-lived, perched water tables, and generated most of the water in high peak flows. Field evidence (Tischendorf, 1969; Ayers, 1976) indicates that the simulated hydrological behavior of the two soil groups is realistic.

![Graphs](image)

**Fig. 7. Actual and simulated hydrographs.**

<table>
<thead>
<tr>
<th>Date (Y/M/D)</th>
<th>Precip. (mm)</th>
<th>Initial flow rate (m³ min⁻¹ km⁻²)</th>
<th>Peak (m³ min⁻¹ km⁻²)</th>
<th>Storm-flow volume</th>
</tr>
</thead>
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<tr>
<td>66/5/26</td>
<td>51</td>
<td>0.8</td>
<td>10.5</td>
<td>10.0</td>
</tr>
<tr>
<td>67/1/26</td>
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<td>0.4</td>
<td>3.9</td>
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</tr>
<tr>
<td>67/7/7</td>
<td>51</td>
<td>0.4</td>
<td>6.4</td>
<td>0.6</td>
</tr>
<tr>
<td>68/3/7</td>
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<td>14.1</td>
<td>1.1</td>
</tr>
<tr>
<td>69/1/19</td>
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<td>0.7</td>
<td>134.9</td>
<td>36.6</td>
</tr>
<tr>
<td>69/4/18</td>
<td>51</td>
<td>1.4</td>
<td>46.9</td>
<td>15.6</td>
</tr>
<tr>
<td>69/5/8</td>
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<td>0.6</td>
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</tr>
<tr>
<td>70/3/19 (h)</td>
<td>107</td>
<td>0.4</td>
<td>94.9</td>
<td>45.4</td>
</tr>
</tbody>
</table>

*Corresponds to Fig. 7.

Source area expansion on a drainage basin is influenced by broad terrain features that favor subsurface flow convergence or divergence. As the water content of any particular part of the basin increases in response to rainfall, the probability of occurrence of surface saturation and seepage is enhanced by a host of small irregularities, such as rock outcrops, soil discontinuities and micro-relief, which may cause temporary and localized exceedance of the flow transmission capacity of the mantle. The increased store of soil water is thus tapped by the expanded channel system long before possible occurrence of surface saturation over large portions of the hillside.

In the simulation, surface saturation is an abrupt and surficially uniform phenomenon. The spatial resolution (large size of segments and increments) adopted for this effort obliterates small terrain features. Increments are perceived by VSSAS2 as smooth and homogeneous over their width and length; saturation occurs over their whole surface or not at all. This homogeneity of response is somewhat alleviated by redividing the initial slope to increments of smaller slope length at the upper limit of contiguous surface saturation (Fig. 6). Nevertheless, low spatial resolution resulting from necessarily finite amount of field information limits the occurrence, extent and duration of simulated surface saturation from infiltrating or exfiltrating water. When surface saturation finally occurs, the convergence created by the broad features that can be "seen" by VSSAS2 cannot generate the amount of exfiltration these saturated zones would yield in the field. As a result, peakflows are under-estimated in nearly all large-event simulations (Fig. 7h). Most simulated peaks are dominated by pulses of rainfall intercepted by surface-saturated lower increments. Direct observation made on the basin by Tischendorf showed actual peaks delayed until the main rainfall
had ceased, indicating much subsurface contribution. Furthermore, simulated outflows plunge down immediately after the peak (Fig. 7e,f) and fail to generate the second peak of double-peaked hydrographs (Fig. 7b), a clear indication that exhaustion of subsurface storm flow is under-estimated. Finally, since more water is left on the slope at the end of the simulated flow hydrograph, the decline of the base flow component is too slow to properly fit the actual hydrograph.

The response of a real basin to small rainfall events is controlled in large part by channel and stream-side geometry, and by the moisture content of stream-side soil. All segments were initialized at identical baseflow rates, thus making the stream-side portions of segments bordering intermittent reaches (Fig. 3) wetter than their real-life counterparts. This simplifying assumption could only exaggerate simulated volumes of small-event storm flows. Better simulation of peaks and volumes for small events could be achieved by securing a more accurate description of stream-side geometry and moisture contents. However, the extra effort seems unwarranted in view of the limited practical importance of small storm flows.

In short, even with better mathematics and smaller soil elements and time steps, disagreement between actual and simulated hydrographs would still occur due to the compounding of three types of errors: Inadequate representation of small terrain features, inadequate quantification the hydraulic properties of soils and inaccurate initial moisture contents.

CONCLUSION

As with any simulator, fittings between actual and predicted could have been improved by a process of repeated calibration on some of the soil properties. However, this was not done since the objective of this work was not to produce the closest agreement, but to see how well hydrographs could be predicted from known physical and hydrological properties of a basin in light of the variable source area concept of streamflow generation.

Although technically improved from VSAS1 (Troendle, 1979), VSAS2 remains conceptually similar to the earlier version. The test basin for VSAS1 featured even-depth and undisturbed forest soils of silt-loam texture and no strong horizonation. Slopes were gentle and relatively constant. There were no major irregularities on the basis, such as gulles or rock outcrops. On such a regular basin, the agreement between actual and simulated hydrographs appeared satisfactory. On the other hand, the Whitehall basin has a naturally complex hydrology which has been further perturbed by abusive cotton farming. When storm-flow simulations were attempted with VSAS2, the heterogeneity of the basin revealed shortcomings of the approach that were not apparent in the earlier trials with VSAS1. VSAS2 approximates hydrologic processes closer than many previous simulators. But results from the simulation effort reinforce the conclusion drawn by Philip (1980) that the primary barrier to further progress is field heterogeneity.

A functional model of streamflow source areas, one that incorporates our best understanding of the process, is necessary to properly interpret cause and effect in problems with water pollution, water yield and erosion, as these are influenced by land use. This research effort was pursued with the conviction that all avenues leading to a useful, cost-efficient and sufficiently accurate simulator of the first-order basin had not yet been explored. The author believes that a better understanding of variable source area dynamics is the key to clearer identification of land and water management problems.

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