PREDICTING STORMFLOW AND PEAKFLOW FROM SMALL BASINS IN HUMID AREAS BY THE R-INDEX METHOD

J. D. Hewlett, G. B. Cunningham, and C. A. Troendle

ABSTRACT: A relatively simple nonlinear equation was fitted to 468 stormflows larger than 0.05 area inches on 11 forested basins from New Hampshire to South Carolina, providing a predictive method for use on forest and wildlands in humid regions. Stormflow in area inches \( Q \) was:

\[
Q = 0.4 \times R \times P^{1.5} (1 + I^{0.25})
\]

where \( R \) is the mean value of \( Q/P \) for all \( P \) larger than one inch, \( P \) is storm rainfall in inches, and \( I \) is the initial flow rate in \( \text{ft}^3/\text{sec} \times \text{mi}^2 \). S.E. was 0.3 inch of stormflow. Peakflow was similarly estimated. S.E. 26 \( \text{ft}^3/\text{sec} \times \text{mi}^2 \). The R-index method is proposed as a practical tool in forest and wildland management. Similar to the SCS runoff curve number method, the R-index method requires no prior assumptions about infiltration capacities of forest lands, but calls for the mapping of all first-order streams for the average storage capacity index \( R \), i.e., the mean hydrologic response of the source areas. Tested against the runoff curve method on four independent basins, predictions by the R-index method were considerably more accurate when field information normally available to planners and managers was used in both methods.

(KEY TERMS: stormflow, peakflow, direct runoff, R-index method.)

INTRODUCTION

Existing procedures for predicting stormflow volumes and peak discharges from small drainage basins are relatively complex. The complexity tempts the forest and wildland manager to avoid use of hydrologic procedures when planning small structures or estimating the hydrologic impacts of land management practices. This paper proposes a parametric, deterministic, non-linear model (as classified by Jackson and Aron, 1971) for estimating stormflow volumes and peak discharges based on R-index or hydrologic response maps (Hewlett and Hibbert, 1967). The R-index method will be compared to a standard method for predicting stormflow and peak discharge from forest and wildlands.
MODEL DEVELOPMENT

Three factor complexes control the volumes of water that a small basin (those areas contributing to the first, second or third order stream, using Horton's classification) will deliver as stormflow (quickflow, direct runoff) during and following a rainstorm. These are storm precipitation, antecedent storage, and the dynamic or hydrologic storage capacity of the basin. In order to give these causative variables meaning over a wide range of basin types and sizes, it is necessary to define exactly what the dependent variable, stormflow, includes. The classification of streamflow into stormflow and baseflow is an arbitrary decision of the analyst, whose main objective is, or should be, to maintain a consistent criterion for separation over all basins and all hydrographs. Such a criterion has been proposed (Hewlett and Hibbert, 1967) and has been applied to many of the small basins gaged by the U.S. Forest Service in the East (Hibbert and Cunningham, 1967) and to basins up to 200 square miles gaged by the U.S. Geological Survey (Woodruff and Hewlett, 1970). The result is a body of rainfall-discharge data that has been reduced to a common format (Figure 1) for comparison and mapping (Figure 2).

Figure 1. Definition Diagram Showing the Relation Between Input and Output Variables.
Figure 2. Preliminary Response Map of Eastern United States (from Woodruff and Hewlett, 1970). The mean annual index is based on all stormflows and total precipitation. The values shown above may be multiplied by about 2 in order to approximate the R-index developed in this paper.
Stormflow is computationally defined as all water delivered from a basin at a rising rate of discharge in excess of 0.05 cubic feet per second per square mile per hour (cfsm/hr) from the beginning of the hydrograph rise. The mean annual hydrologic response of a basin was defined by Hewlett and Hibbert (1967) as the average value of Q/P, where Q is the sum of all stormflows over the year and P is annual precipitation, both in depth per unit area. The method has been found to be a satisfactory index of regional response (Colonell and Higgins, 1973; Sopper and Lull, 1970) and one of the most sensitive indices to urbanization of watersheds (Lull and Sopper, 1969). A few years of rainfall stormflow records, or even a dozen recorded storms, can provide an estimate of the mean annual response of a basin.

A numerical index of response could be mapped to provide a permanent planning factor for future designs or impact studies. The coarse-scaled map in Figure 2 shows that the mean annual hydrologic response to rainfall is clearly related to geologic provinces, but the accompanying analysis of basin factors (Woodruff and Hewlett, 1970) showed that response was quite insensitive to large variations in basin size, shape, and steepness, as well as in vegetal cover and land use (excluding urbanization and reservoirs). Basin size (A) and steepness (G) were again evaluated in this data set and found to be statistically unrelated to stormflow (Q).

The following general model was investigated:

\[
Q = f(P, P_{60}, D, I, S, A, G, R)
\] (1)

where Q and P in inches are as before; P_{60} is the inches of rain falling during the most intense hour of the storm; D is duration of the rainstorm in hours; I is the initial flow rate in cubic feet per second per square mile; S is season expressed as the sine of the day number from various points on the calendar (see below); A is the area of the basin in square miles; G is the gradient in feet per mile from the measuring station to a point on the water divide directly above the origin of the main channel; and R is the mean hydrologic response of the basin, redefined here as the average value of Q/P for all rainstorms greater than one inch. The restriction to storms greater than one inch gave R-values that are approximately 2 times larger than the average annual responses shown in Figure 2.

The data is also arrayed to investigate the following general model for peakflow (Q_p, peak discharge rate in cfsm):

\[
Q_p = f(Q, P, P_{60}, D, I, S, A, G, R)
\] (2)

In most hydrologic studies storm rainfall is identified in terms of the length of break periods between rainfall pulses; six hours is most often used. In this study the duration of the computed hydrograph (that is, using the 0.05 cfsm/hr separation slope) was used to identify the contributing storm rainfall. Any clearly separated hydrograph produced by one or more inches of storm precipitation was accepted for analysis. Finally the criterion was imposed that storm rainfall includes all rain falling between a point in time two hours before hydrograph initiation and the termination of stormflow by the 0.05 cfsm/hr separation slope (Figure 1); the 2-hour advance allows for some clock error between water level and rainfall recorders and also for small aberrations in the computer-determined hydrograph rise.
The recording raingage data at various stations indicated that the characteristic rainfall duration that produces a single hydrograph separation ranges from one to about 100 hours, average 24 hours, and that 72 hours includes virtually all such rainstorms. The average rainstorm in this data set (assumed to be “characteristic” of the East) is 2 inches lasting 24 hours, with a 30-minute maximum intensity of 0.8 inch/hour, and a 60-minute maximum intensity of 0.4 inch/hour. Twelve hours without rain caused stormflow termination most of the time.

The R-index method for forest and wildland use will not require estimates of soil type, cover condition, land slope, cultural practice, infiltration, overland flow, time of concentration, or detailed time distribution of storm rainfall.

DATA SOURCES

Eleven small forested basins, carefully gaged for many years by the U.S. Forest Service, Penn State University and the University of Georgia, provided the 468 individual stormflow events used in the following model development. The basins, ranging in size from 0.09 to 18 square miles, and from very flat to very steep, are described in Table 1. The basins were selected to represent as wide a range in the variables of interest as possible. Sopper and Lull (1970) concluded after extensive comparison that these experimental basins were satisfactorily representative for most hydrological interpretations within their respective regions. Each basin is equipped with a sharp-crested weir, a recording raingage and a network of standard raingages, so that weighted basin precipitation by half-hour time units is available. Nearly all the basins have been reported in studies elsewhere (Douglass, 1972); for the most part they are control basins at experimental watershed stations. All are under relatively stable forest cover.

There is negligible correlation in this data set among rainstorm size, basin area, basin relief, elevation, initial flow and mean hydrologic response. There is some confounding of location (latitude, general climate) and rainstorm size and intensity. The characteristic rainstorm sampled at Coweeta was 1.8 times that at Hubbard Brook, and average rainfall intensity decreases from 0.8 in/hr in Georgia to 0.3 in/hr at Hubbard Brook.

In addition to the basic separation slope of 0.05 cfs/mi², "stormflow" was computed by three other rules of separation: 0.03, 0.08 and 0.11 cfs/mi². This yielded a vector of four successively smaller values for stormflow delivered by each event. Table 2 shows that these four variables correlated highly, which perhaps explains why an exhaustive analysis of stormflow as a vector of 4 values proved relatively fruitless in providing any response indices superior for predictive purposes to the simple one already proposed; that is, arithmetic mean Q/P where Q is computed using the 0.05 cfs/mi² separation slope and only storms equal to or greater than 1 inch are accepted. On basins smaller than 20 square miles, it seems to matter very little what fixed rule for hydrograph separation is used; all give a dependent variable that shows similar correlation to all other variables of interest, including peak discharge during the stormflow event. From this point forward, stormflow is classified by the 0.05 cfs/mi² rule only. The volume and duration of stormflows so classified do not differ appreciably from those determined by rules commonly used by other hydrologists.

A number of ways of computing response were tried in the non-linear least squares fitting of the basic model. Among these were (Q/P)X log (Q/P), (log Q/log P), QX/P, Q/PX, (P-Q)/P, (P-Q)/Q, and the S-value from the SCS (1972) runoff curve model.
<table>
<thead>
<tr>
<th>Basin</th>
<th>Physiographic Province</th>
<th>Area (Acres)</th>
<th>Cover Type</th>
<th>Soil Type</th>
<th>Basin Relief (Ft./Mi.)</th>
<th>Length of Record (Years)</th>
<th>Mean Annual Precip. (Inches)</th>
<th>Mean Annual Discharge (Inch)</th>
<th>Depth to Bedrock (Ft.)</th>
<th>Hydrologic Soil Group</th>
</tr>
</thead>
<tbody>
<tr>
<td>Berea, Ky.</td>
<td>Interior low plateau</td>
<td>200</td>
<td>Oak-Hickory</td>
<td>Sandy to gravelly loam</td>
<td>815</td>
<td>2</td>
<td>41.9</td>
<td>14.9</td>
<td>2 – 5</td>
<td>B/C</td>
</tr>
<tr>
<td>Leading Ridge 1, Pa.</td>
<td>Ridge &amp; valley</td>
<td>303</td>
<td>Oak-Hickory</td>
<td>Silty to stoney loam</td>
<td>740</td>
<td>12</td>
<td>39.6</td>
<td>14.9</td>
<td>2 – 10</td>
<td>C</td>
</tr>
<tr>
<td>Hubbard Brk 3 N. H.</td>
<td>White Mtns.</td>
<td>105</td>
<td>Northern Hardwoods</td>
<td>Stoney</td>
<td>1240</td>
<td>12</td>
<td>48.0</td>
<td>26.0</td>
<td>0 – 8</td>
<td>B</td>
</tr>
<tr>
<td>Charleston 77 S. C.</td>
<td>Coastal Plain (Flatwoods)</td>
<td>379</td>
<td>Pine and pine-hardwoods</td>
<td>Sandy to clayey loams</td>
<td>21</td>
<td>8</td>
<td>50.0</td>
<td>16.0</td>
<td>1 – 2</td>
<td>B/D</td>
</tr>
<tr>
<td>Charleston 78 S. C.</td>
<td>Coastal Plain (Flatwoods)</td>
<td>11,300</td>
<td>Pine and pine-hardwoods</td>
<td>Sandy to clayey loams</td>
<td>16</td>
<td>8</td>
<td>50.0</td>
<td>16.0</td>
<td>1 – 2</td>
<td>B/D</td>
</tr>
<tr>
<td>Fernow 4 W. Va.</td>
<td>Allegheny Mtns.</td>
<td>96</td>
<td>Northern Cove hardwoods</td>
<td>Silt loam</td>
<td>1011</td>
<td>23</td>
<td>56.9</td>
<td>23.6</td>
<td>3 – 8</td>
<td>C</td>
</tr>
<tr>
<td>Fernow 6 W. Va.</td>
<td>Allegheny Mtns.</td>
<td>54</td>
<td>Oak-Hickory</td>
<td>Silt loam</td>
<td>932</td>
<td>7</td>
<td>56.7</td>
<td>19.4</td>
<td>3 – 8</td>
<td>C</td>
</tr>
<tr>
<td>Coweeta 8 N. C.</td>
<td>Southern Appalachians</td>
<td>1,877</td>
<td>Oak-Hickory</td>
<td>Sandy clay loam</td>
<td>1162</td>
<td>40</td>
<td>78.2</td>
<td>45.2</td>
<td>3 – 70</td>
<td>B</td>
</tr>
<tr>
<td>Coweeta 34 N. C.</td>
<td>Southern Appalachians</td>
<td>81</td>
<td>Oak-Hickory</td>
<td>Sandy clay loam</td>
<td>2270</td>
<td>19</td>
<td>77.5</td>
<td>45.1</td>
<td>20 – 70</td>
<td>B</td>
</tr>
<tr>
<td>Coweeta 36 N. C.</td>
<td>Southern Appalachians</td>
<td>120</td>
<td>Oak-Hickory</td>
<td>Sandy clay loam</td>
<td>2323</td>
<td>27</td>
<td>87.0</td>
<td>66.0</td>
<td>3 – 50</td>
<td>B</td>
</tr>
<tr>
<td>Whitehall 1 Ga.</td>
<td>Southern Piedmont</td>
<td>60</td>
<td>Old Field</td>
<td>Sandy clay loam</td>
<td>333</td>
<td>8</td>
<td>49.0</td>
<td>12.0</td>
<td>1 – 100</td>
<td>B</td>
</tr>
</tbody>
</table>

1 Figure represents depth to continuous (basin-wide) water table in the Coastal Plain.

2 An average weighted by area of soil type present. From SCS Hydrology, Section 4, 1972.
TABLE 2. Simple Correlation Coefficients Between Stormflows Determined by Four Separation Slopes (.03, .05, .08, .11 cfs/hr) on Data from 11 Basins.

<table>
<thead>
<tr>
<th>Separation Slope (cfs/hr)</th>
<th>0.05</th>
<th>0.08</th>
<th>0.11</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.03</td>
<td>0.996</td>
<td>0.998</td>
<td>0.965</td>
</tr>
<tr>
<td>0.05</td>
<td></td>
<td>0.997</td>
<td>0.976</td>
</tr>
<tr>
<td>0.08</td>
<td></td>
<td></td>
<td>0.983</td>
</tr>
</tbody>
</table>

Whether expressed as arithmetic or geometric means, none of these showed any consistent tendency to improve predictability of stormflow across all basins. The criterion for comparison was the statistical significance of an improvement in the multiple coefficient of determination when Equation 1 was fitted in various ways. $Q^2/P$ proved to be almost normal in distribution but arithmetic mean $Q/P$ seems the simplest index of dynamic storage capacity to use in the current model.

Factor analysis of these variables and their transforms served only to verify prior expectations that four factor complexes existed in the data set: input factors ($P, P_6Q, D$), existing storage condition ($I, S$), dynamic storage capacity and output ($R, Q$), and superficial geography ($A, G$).

THE R-INDEX MODEL

After many trials, during which numerous transforms and interactions were examined, the following logical and simple form of the model was fitted by a Marquardt non-linear least squares method (Marquardt, 1963).

$$Q = \beta_1 R \frac{P^{\beta_2}}{(1 + P^{\beta_2})} (1 + 1^{\beta_3})$$  \hspace{1cm} (3)

The 1 is included to prevent indeterminate $Q$ when $I$ approaches zero. The other variables in Equation 1 contributed virtually nothing to reduce the standard error. The coefficients, suitably rounded, were fitted to the 468 cases on 11 basins:

$$Q = 0.4 R p^{1.5} (1 + 1^{2.5})$$  \hspace{1cm} (4)

The standard error of estimate of the model was 0.30 inch of stormflow at the mean value ($Q = 0.52$ inch). Figure 3 shows how the data fit the model; below the mean value the error is small and slightly positive (over-estimate), while in the high ranges of stormflow, errors are larger and tend toward the negative (under-estimate). Basins with high R indices gave larger absolute errors than the others (Figure 4); for this reason the model was fitted for each basin with $\beta_1$ constrained to its value in the general model (0.4), and individual basin estimates of $\beta_2$ and $\beta_3$ were made (Table 3). These reveal appreciable deviation of $\beta_2$ in the Whitehall Basin (Piedmont) but a fair agreement in the other ten. The parameter $\beta_3$ was relatively unstable, deviating widely on the Whitehall Basin.
Figure 3. A Plotting of Actual Versus Predicted Stormflow by Equation 4, Showing the Fit to the Original Data from all 11 Basins.
Figure 4. Composite Graph Showing the Fit of Equation 4 to Stormflow on Four Individual Basins in the Basic Data Set.
TABLE 3. Non-Linear Least Squares Fitting of $\beta_2$ and $\beta_3$ by Basin, Under the Constraint that $\beta_1 = 0.4$ and $R =$ the Individual Basin Response Index.

<table>
<thead>
<tr>
<th>Basin</th>
<th>$\overline{Q}$ (Inch)</th>
<th>$\overline{P}$ (Inch)</th>
<th>$\overline{T}$ (cfs/m)</th>
<th>$R$</th>
<th>$\beta_1$</th>
<th>$\beta_2$</th>
<th>$\beta_3$ (Inch)</th>
<th>S.E.</th>
</tr>
</thead>
<tbody>
<tr>
<td>All</td>
<td>468</td>
<td>0.52</td>
<td>2.12</td>
<td>1.87</td>
<td>0.22</td>
<td>0.4</td>
<td>1.50</td>
<td>0.25</td>
</tr>
<tr>
<td>Leading Ridge</td>
<td>52</td>
<td>0.29</td>
<td>1.22</td>
<td>1.24</td>
<td>0.22</td>
<td>0.4</td>
<td>1.70</td>
<td>0.60</td>
</tr>
<tr>
<td>Berea</td>
<td>18</td>
<td>0.47</td>
<td>1.56</td>
<td>0.84</td>
<td>0.29</td>
<td>0.4</td>
<td>1.77</td>
<td>1.20</td>
</tr>
<tr>
<td>Hubbard Brook</td>
<td>77</td>
<td>0.48</td>
<td>1.53</td>
<td>1.28</td>
<td>0.27</td>
<td>0.4</td>
<td>1.70</td>
<td>0.63</td>
</tr>
<tr>
<td>Fernow 4</td>
<td>50</td>
<td>0.73</td>
<td>1.95</td>
<td>1.54</td>
<td>0.35</td>
<td>0.4</td>
<td>1.35</td>
<td>0.27</td>
</tr>
<tr>
<td>Fernow 6</td>
<td>18</td>
<td>0.79</td>
<td>2.21</td>
<td>1.13</td>
<td>0.33</td>
<td>0.4</td>
<td>1.41</td>
<td>0.34</td>
</tr>
<tr>
<td>Coweeta 8</td>
<td>60</td>
<td>0.38</td>
<td>2.78</td>
<td>3.28</td>
<td>0.11</td>
<td>0.4</td>
<td>1.34</td>
<td>0.39</td>
</tr>
<tr>
<td>Whitehall</td>
<td>48</td>
<td>0.31</td>
<td>2.14</td>
<td>0.79</td>
<td>0.11</td>
<td>0.4</td>
<td>1.91</td>
<td>1.87</td>
</tr>
<tr>
<td>Coweeta 34</td>
<td>26</td>
<td>0.11</td>
<td>2.66</td>
<td>3.26</td>
<td>0.034</td>
<td>0.4</td>
<td>1.44</td>
<td>0.00*</td>
</tr>
<tr>
<td>Coweeta 36</td>
<td>51</td>
<td>0.79</td>
<td>2.76</td>
<td>4.26</td>
<td>0.23</td>
<td>0.4</td>
<td>1.32</td>
<td>0.27</td>
</tr>
<tr>
<td>Charleston 77</td>
<td>38</td>
<td>0.93</td>
<td>2.45</td>
<td>0.78</td>
<td>0.34</td>
<td>0.4</td>
<td>1.52</td>
<td>0.47</td>
</tr>
<tr>
<td>Charleston 78</td>
<td>30</td>
<td>0.48</td>
<td>2.41</td>
<td>1.10</td>
<td>0.19</td>
<td>0.4</td>
<td>1.47</td>
<td>0.93</td>
</tr>
</tbody>
</table>

*Parameters in the equation: $Q = \beta_1 R \overline{P}^2 (1 + 10^{-2\beta_3})$.

**The zero estimate of $\beta_3$ on Coweeta 34 signifies that antecedent flow contributed nothing to the variation in $Q$ on that basin. Coweeta 34 has exceptionally deep soil mantles and a deeply-entrenched stream channel.

A further constraint must be placed on all the equations in Table 3 to prevent increments in predicted stormflow from exceeding increments in the causative rainfall. That is, for large values of $P$ the first derivative of $Q$ with respect to $P$ must not exceed 1.0.

$$\frac{\delta Q}{\delta P} \leq 1.0 = 0.6 R P^{-5} (1 + 1^{-25})$$

Solving the derivative under this constraint, the value of $P$ above which any further storm rainfall produces an equal amount of stormflow becomes:

$$P = \frac{1}{[0.6R(1 + 1^{-25})]^2}$$

For example, if $R = 0.30$ and $I = 1$, then all storm rainfall above 7.72 inches will be directly added to stormflow. If $R = 0.10$ and $I = 1$, then all storm rainfall above 69 inches would be added directly to stormflow. If $I$ is increased to 10 in the latter case, all rain above 36 inches would be added to stormflow. Figure 5 shows the solution of the general model under this constraint.

The necessity for constraint, which in any case operates largely beyond the data range normally experienced, should be obvious; that is, seldom if ever will a free-draining basin without structures deliver more than an inch of stormflow for an inch of precipitation. To conclude otherwise would be to hypothesize a triggering device that would dump previously stored water in addition to all the rainfall, such as might happen if a dam burst during stormflow or if a warm rain fell on a wet snowpack overlying saturated soils. We
Figure 5. Solution of Equation 4, Constrained by Equation 6, for Four Assumed Values of R-Index and Three Values of Antecedent Flow (I).
find it reasonable to assume, therefore, that any curve calculated from the model must be tangent to a 1:1 line at the point where P equals Equation 6. The "true" function may be asymptotic to 1:1, but the tangent will make virtually no difference in practice and it has the advantage of simplifying the model. The method is obviously inapplicable to snowmelt hydrographs.

USE OF SINE OF DAY

The variable I (antecedent flow rate in cfsm), while theoretically ideal, is not the most practical storage variable because it will seldom be available to the user in advance. Furthermore, local measurements of I can be seriously in error because of underflow past the gaging station or leaks in and out of the basin. One remedy might be to predict the storage variable I from weather data (for example, by one of the evaporation or soil moisture methods). To test this idea, the present data set was fitted again to determine how much loss of accuracy would occur when a simple seasonal variable is substituted for (1 + I):

\[ Q = \beta_1 R S^{\beta_2} P^{\beta_3} \quad (7) \]

where S is the day number of the storm counted from November 21 = zero, converted to the "sine of the day" plus 2:

\[ S = \text{SIN} \left[ 360 \left( \text{day number}/365 \right) \right] + 2 \quad (8) \]

The numeral 2 is added to the sine merely to avoid negative numbers and zeros in the analysis (Table 4). November 21 was selected by stepwise linear regression to minimize standard error of Q on P and S, using 12 sine functions beginning, in sequence, on the 21st of each month. November 21 showed up consistently as the best beginning date for fitting a sine function to the residuals from regression of Q on P. The rationale for using the sine wave to represent seasonal effects stems from Helvey and Hewlett (1962), who showed that the annual march of both average soil moisture and monthly streamflow fit the sine quite well in the southern Appalachian mountains. In one sense, the variable S serves to correct R for season.

Equation 6 was fitted by non-linear least squares:

\[ Q = 0.56 R S^{0.38} P^{1.5} \quad (9) \]

The standard error (0.32 inch of stormflow) was slightly higher than in Equation 4. Time of the year appears to be a satisfactory substitute for the antecedent flow in our data set; on average, stormflow from a given rainstorm will be 50% greater in February than in August. Equation 9 is also subject to the tangency rule that \( \delta Q/\delta P < 1 \). Further refinement of the season variable for expected or known antecedent precipitation on the basin, or by predicted soil moisture, should allow reduction of the standard error at least to the mean of the individual errors in Table 3 (0.24 inch).
Predicting Stormflow and Peakflow

### TABLE 4. Sine of Day Values (S) for use With Flow Prediction Equations 9 and 14.

<table>
<thead>
<tr>
<th>Day</th>
<th>Nov</th>
<th>Dec</th>
<th>Jan</th>
<th>Feb</th>
<th>Mar</th>
<th>April</th>
<th>May</th>
<th>June</th>
<th>July</th>
<th>Aug</th>
<th>Sept</th>
<th>Oct</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1.66</td>
<td>2.17</td>
<td>2.65</td>
<td>2.94</td>
<td>2.99</td>
<td>2.77</td>
<td>2.36</td>
<td>1.84</td>
<td>1.37</td>
<td>1.06</td>
<td>1.01</td>
<td>1.23</td>
</tr>
<tr>
<td>7</td>
<td>1.76</td>
<td>2.27</td>
<td>2.72</td>
<td>2.98</td>
<td>2.97</td>
<td>2.70</td>
<td>2.26</td>
<td>1.74</td>
<td>1.29</td>
<td>1.03</td>
<td>1.04</td>
<td>1.30</td>
</tr>
<tr>
<td>14</td>
<td>1.88</td>
<td>2.39</td>
<td>2.80</td>
<td>2.99</td>
<td>2.93</td>
<td>2.61</td>
<td>2.15</td>
<td>1.62</td>
<td>1.21</td>
<td>1.01</td>
<td>1.08</td>
<td>1.39</td>
</tr>
<tr>
<td>21</td>
<td>2.00</td>
<td>2.49</td>
<td>2.87</td>
<td>3.00</td>
<td>2.88</td>
<td>2.52</td>
<td>2.03</td>
<td>1.51</td>
<td>1.14</td>
<td>1.00</td>
<td>1.13</td>
<td>1.49</td>
</tr>
<tr>
<td>28</td>
<td>2.12</td>
<td>2.59</td>
<td>2.92</td>
<td>2.99</td>
<td>2.82</td>
<td>2.41</td>
<td>1.90</td>
<td>1.43</td>
<td>1.09</td>
<td>1.01</td>
<td>1.20</td>
<td>1.60</td>
</tr>
</tbody>
</table>

### BASIN INTERPRETATION

More is known about the subsurface morphology and general hydrology of the Whitehall and Fernow No. 4 basins than any others in the set. Fernow No. 4 (96 acres and never cultivated) is under typical Appalachian hardwood cover, with a permeable surface soil but a uniform soil depth of only five to seven feet over impermeable shale. The Whitehall basin (60 acres) is 90-percent covered with second growth pine and hardwoods typical of the Piedmont; the soil mantle above the undulating schistose bedrock varies from zero to over 100 feet (Neal, 1967) and averages about 40 feet overall; the lower 20 feet is mostly below a water table except along lower slopes and in the channel region. In addition, the upper reaches of the Whitehall Basin were severely cultivated and eroded until about 1930, leaving a 6- to 8-inch sandy plow layer over a clayey plow pan on about 10 percent of the basin area. Over much of the rest of the basin there exists a heavy B-horizon from about 3 to 5 feet below the surface. In summer and fall the B-horizon is relatively dry and serves as a sink for percolating waters, but when wet in winter and spring, this zone acts as a impedance layer, forcing lateral subsurface flow to emerge from the soil surface into an expanding channel (Tischendorf, 1969). Because of these two conditions (plow pan and B-horizon), this basin behaves as if it were shallow-soiled during winter and spring, or whenever stormflows exceed 2 to 3 inches. The Fernow basin, although relatively shallow, has long slopes and no time-variable impedance layer, and therefore responds more gradually to storm rainfall, as shown by the value of \( \beta_2 \) in Table 3. On the Whitehall basin, the variable source area (Hewlett and Hibbert, 1967) for stormflow expands slowly at first, then very rapidly as rainstorms exceed about 2 inches. Thus Fernow 4 fits the model well but Whitehall deviates from expected values above 2 inches of stormflow (Figure 4).

We conclude that the R-index satisfactorily adjusts the model to fit basins that do not exhibit distinct time-variable impedance layers; where these predominate, local adjustment of parameters may be desirable. But, as the following comparisons will show, this preliminary general model does a better job of predicting stormflows than the runoff curve method currently in use.
TESTING THE STORMFLOW MODEL

To test the model (Equation 4) four forested basins from four physiographic provinces were secured from records of the U.S. Forest Service (Table 5). R indices for each basin were estimated using all separable stormflows produced by storms larger than one inch. The National Engineering Handbook of Hydrology (SCS, 1972) provided rules for determining antecedent moisture conditions (AMC) for each event from 5-day antecedent rainfall and the season of the year (dormant and growing). An S-value (a compound index of storage capacity) was computed for each event in accordance with the SCS model for stormflow:

\[ Q = \frac{(P - 0.2S)^2}{P + 0.8S} \]  

from which we compute:

\[ S = 5P + 2Q - (4Q^2 + 5PQ)^5 \]  

S-values for each event on a given basin were computed and the basin mean S-value was used in the equation that defines the curve number (CN) for average basin response (Section 4, SCS, 1972):

\[ CN = \frac{1000}{10 + S} \]  

Basin curve numbers were adjusted for AMC to estimate stormflow Q by the curve number method and compared storm by storm with estimates by the R-index method. Therefore, in both cases, storage capacity indices (CN and R) are derived from actual data. There may be some doubt about our use of the S-value in this manner but it is the best that can be done. The S-value is defined in words by SCS (1972) simply as the sum of "potential maximum retention plus the initial abstraction of rainfall." This leaves only the solution of Equation 10 as a method of computing it.

In all, about 300 stormflow predictions on 4 basins were compared. Figure 6 shows better grouping of R-index estimates around the one-to-one line of perfect prediction in all four cases. The CN estimates below one inch of storm rainfall tend to show a large number of zeros because of the assumption about "initial abstractions" that is basic to Equation 10. The R-index estimates of Q are truncated at about 0.25 inch because only rainstorms larger than 1 inch were selected (except on Coweeta 14).

Comparison was also made between the two methods using CN derived from soil hydrologic groups instead of actual data on P and Q. As expected, this further reduced the accuracy of the CN method, because the value of S so determined has a further element of subjectivity. Our tests, of course, do not demonstrate conclusive superiority of the R-index method for all uses or all regions. Experienced personnel with sufficient regional knowledge of hydrologic responses can make almost any predictive method work to a specified level of accuracy, if local adjustments of the model are allowed. The case for the R-index method rests at this time not on its absolute superiority but on its relative simplicity.
TABLE 5. Summary Description of Four Basins Used to Test and Compare the Runoff Curve and R-Index Methods.

<table>
<thead>
<tr>
<th>Physiographic Province</th>
<th>Charleston S. C. 79</th>
<th>Coweeta N. C. 14</th>
<th>Union 3</th>
<th>Fernow 1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area (acres)</td>
<td>1,184</td>
<td>150</td>
<td>22</td>
<td>74</td>
</tr>
<tr>
<td>Cover type</td>
<td>Pine and pine-hardwoods</td>
<td>Oak-hickory cove hardwoods</td>
<td>Old-field pine</td>
<td>Northern cove hardwoods</td>
</tr>
<tr>
<td>Soil type</td>
<td>Sandy to clayey loams</td>
<td>Sandy clay loam</td>
<td>Sandy clay loam</td>
<td>Silt loam</td>
</tr>
<tr>
<td>Basin Relief (ft/mi)</td>
<td>20</td>
<td>1.145</td>
<td>254</td>
<td>950</td>
</tr>
<tr>
<td>Length of record (years)</td>
<td>7</td>
<td>16</td>
<td>9</td>
<td>7</td>
</tr>
<tr>
<td>Mean annual precip. (inches)</td>
<td>50.0</td>
<td>73.8</td>
<td>43.1</td>
<td>60.0</td>
</tr>
<tr>
<td>Mean annual discharge (inches)</td>
<td>15.0</td>
<td>39.7</td>
<td>15.8</td>
<td>23.0</td>
</tr>
<tr>
<td>Depth to bedrock (ft)</td>
<td>1-2^1</td>
<td>10-70</td>
<td>20-100</td>
<td>3-8</td>
</tr>
<tr>
<td>Hydrologic soil group^2</td>
<td>B/D</td>
<td>B</td>
<td>B</td>
<td>C</td>
</tr>
</tbody>
</table>

^1 Figure represents depth to continuous (basin-wide) water table in the Coastal Plain.
^2 An average weighted by area of soil type present, from SCS Hydrology, Section 4, 1972.

MAPING THE R-INDEX

Recognition that stormflow from forests and wildlands has its origin in a variable source area that operates primarily by subsurface routes throws emphasis on the first order basin as the basic hydrological unit, rather than on units of a given soil type, which do not necessarily conform to the dynamic source area pattern (Hewlett and Troendle, 1975). Because land use has different impacts on stormflow, depending on how far from the channel the activity occurs, the responsiveness of a basin cannot be simply an interpretation of the weighted infiltration capacities of the various soil types present. Rather it is an expression of the conditions of the channel network and the dynamic subsurface storage capacities of those parts of the basin having an immediate influence on stormflow. Most first order basins have R indices that respond conservatively to large changes in land use but are quite sensitive to the inherent geological differences between basins (Woodruff and Hewlett, 1970). The stormflow model proposed here requires that the index R,
Figure 6. Composite Graph Showing Comparison of Stormflow Predictions by the Runoff Curve and the R-Index Methods on Four Basins not in the Original Data Set.
or some appropriate modification of it, be mapped for all first order basins of interest to the manager.

The mapping of the forest and wildland properties of the East for the index R might not be as formidable as it appears. Gross-scale maps are already available for some regions (Woodruff and Hewlett, 1970; Colonell and Higgins, 1973; Sopper and Lull, 1970). Forest properties within a small county in Georgia have been mapped in 10-acre units as a pilot study (Hewlett and Moore, 1976). There is reason to believe that remote sensing techniques will reveal geologic features that dominate the dynamic response of source areas (Ishaq and Huff, 1974). Short-term observations (ground truth) on rainfall and streamflow, together with soil type maps, local examination of stream channels (high water marks, etc.), soil mantle surveys by field scouting (wells, mining operations, road cuts and the like), locally-tested unit hydrographs and runoff curve numbers, will provide a preliminary estimate of R for many first order streams. Adjacent areas could be interpolated as first approximations, which can be adjusted upward or downward as new local information becomes available. The land manager would quickly acquire a working knowledge of hydrologic responses in his district, and the R-index map could be passed along to his successor for gradual improvement.

An analysis of the distribution of the variate Q/P suggests possibilities for speeding up the mapping of hydrologic response. Q, P and Q/P all approximate Pearson Type III distributions on all basins with sufficient observations to make a reasonable test (8 out of the 11 basins). Consequently, the mean ratio Q/P based on a number of consecutive storms will be biased downward because of the low probability of securing a properly-weighted sample of the larger storms; in short, Q/P is not normally distributed around its long-term mean. On the other hand, Equations 4 and 9 suggest that P is a direct function of $Q^{2/3}$ on the average, and therefore the ratio $(Q^{2/3}/P)$ might tend toward a normal distribution.

Fortunately, there is an unusual 30-year period of record available for Coweeta Number 8, an 1800-acre forested basin in the Southern Appalachians (Table 1). All storm events that produced 0.05 inches of stormflow or more were extracted from 1943 through 1973, totaling 545 events, averaging 18 per year. All seasonal and antecedent conditions were represented. Table 6 shows that the coefficient of variation of $(Q^{2/3}/P)$ is half that of Q/P and it appears that the distribution of $(Q^{2/3}/P)$ does not depart greatly from the expected normal (skewness = 0; kurtosis = 3). Analysis of the first four moments on data from 8 widely scattered basins with more than 50 observations tended to verify the hypothesis that the 2/3rds power of Q normalizes the ratio and therefore yields an unbiased estimate of mean $(Q^{2/3}/P)$ regardless of the range over which storm precipitation is sampled. At first it appeared that about 15 to 20 sequential storm events of any size on Coweeta 8 were sufficient to estimate this mean index within 5 percent, whereas over 100 events were required to estimate the mean of Q/P within 5 percent. However, analysis of the data by season (June-November and December-May) gave a significantly higher mean for the winter-spring season (0.160, N = 382) than for the summer-fall season (0.130, N = 164), although the variance, skewness and kurtosis were not significantly changed from their yearly values. This difference in mean response between seasons suggests some difficulty in using short-term sequential sampling of Q and P.
to determine long-term average response indices. We feel this difficulty can be overcome by analysis of existing small watershed data sets to determine coefficients for adjusting a 15-to-20 sample mean toward the expected value of the long-term mean. The distribution, sampling and exact form of \( R \) deserve intensive study if the \( R \)-index method proves useful.

### TABLE 6. The First Four Moments of Variation of \( Q, P, Q/P \) and \( (Q^{2/3}/P) \) for 545 Consecutive Events from 1943 Through 1973 on Coweeta Number 8.

<table>
<thead>
<tr>
<th>Variate</th>
<th>Mean</th>
<th>Variance</th>
<th>(Coeff. of Variation)</th>
<th>Skewness</th>
<th>Kurtosis</th>
</tr>
</thead>
<tbody>
<tr>
<td>( Q^* )</td>
<td>.289</td>
<td>.142</td>
<td>130%</td>
<td>3.070</td>
<td>14.500</td>
</tr>
<tr>
<td>( P )</td>
<td>2.460</td>
<td>2.150</td>
<td>60%</td>
<td>1.950</td>
<td>7.960</td>
</tr>
<tr>
<td>( Q/P )</td>
<td>.094</td>
<td>.0033</td>
<td>60%</td>
<td>1.420</td>
<td>4.860</td>
</tr>
<tr>
<td>( Q^{2/3}/P )</td>
<td>.150</td>
<td>.0020</td>
<td>30%</td>
<td>.528</td>
<td>3.630</td>
</tr>
</tbody>
</table>

*For all \( Q \geq .05 \) inch.

### PEAK DISCHARGE ESTIMATION

While stormflow volume estimates are necessary to plan reservoir capacities and to predict flows farther downstream, peak discharge estimates are needed to map local flood hazard and design small structures. Numerous comparisons of peak discharge above initial flow rate with other variables in this and other data sets indicate that about 75 percent of the variation in peak discharge is accounted for by variation in stormflow volume \( (Q) \). Three logical variables to add are basin area \( (A) \), maximum rainfall intensity \( (P_{60}) \) and the main channel gradient \( (G) \). Unfortunately, the main channel gradient was unavailable in this study; however, the landslope gradient \( G \), as defined in Equation 1, should serve as a highly-correlated substitute until better data is at hand. Accordingly, the following model was fitted by non-linear least squares to \( R, P, S \) and the three additional variables \( A, P_{60} \) and \( G \):

\[
Q_p = \beta_1 R S^{2} P^{1.4} A^{1.3} P_{60}^{0.5} G^{0.6}
\]  

(13)

Fitted to 468 cases on 11 basins, the standard error of estimate was 26 cfs, about 100 percent of the mean value of \( Q_p \) in this data set. But the addition of \( A, P_{60} \) and \( G \) was intuitive only and it remained to be demonstrated that they were improving the predictability of \( Q_p \). Surprisingly the standard error of 26 cfs and the plotted pattern of observed-minus-predicted values were unaffected by dropping first the basin gradient, then the maximum one-hour intensity, and then area. The final fitted model contains the same independent variables as the model for stormflow, with altered parameters:

\[
Q_p = 23 R S^{3.3} P^{1.6}
\]  

(14)
A further fitting with I substituted for S gave no improvement. As with all stormflow models, prediction weakens in the higher ranges of peak discharges (150 cfs or up), and a few extreme peaks in the range 200 to 300 cfs appear to be seriously underestimated (Figure 7). Apparently the variation between basins is great enough to conceal the effect of rainfall intensity on \( Q_p \). However, the similarity of Equations 14 and 9 reinforces the concept, implicit also in the “triangular hydrograph” model of Ogrosky and Mockus (1964), that average peak discharge from small basins is determined primarily by stormflow volume, which in turn is determined primarily by dynamic storage capacity, antecedent storage and rainfall size.

**TESTING THE PEAKFLOW MODEL**

The same four basin records used to test Equation 9 were used to compare Equation 14 with predictions from the nomographs for Type II storms in the appendix of SCS-TP-149 (SCS, 1973). The nomographs are for use with 24-hour or shorter rainstorms, so only these were used in the comparison. The more elaborate SCS “triangular hydrograph method” (SCS, 1973), in which overland flow estimates from short periods of rainfall are piled to give the peak rate, was not used because of the impracticality of predicting detailed rainfall distribution within rainstorms. Thus we feel the methods are compared on an equal basis, given the objectives of land managers or planners, and the hydrological information apt to be available to them in practice.

The comparisons (Figure 8) again show fair accuracy for the R-index method but quite wild predictions using the nomographs in SCS-TP-149. Equation 14 underpredicts the larger peak discharges rather consistently, indicating some missing information about the larger storms — most probably rainfall intensity. However, the objectives of this study are met even by the preliminary models proposed: the R-index method gives quick, relatively accurate estimates of both stormflow and peak discharge from small forested watersheds of the East.

**CONCLUSIONS**

While similar in many respects to the runoff curve number system, which it seeks to supplement rather than to replace, the R-index method differs in a number of important ways.

1. There are no assumptions about initial abstractions from stormflow, about limited infiltration rates, or about the effects of land use and cover on storm discharge and the index \( R \). These effects are known to be minor in most forests and wildlands. On the other hand, the depth of the soil rock mantle is implicit in the \( R \) index.

2. Hydrologic response maps rather than soils maps are the primary data base. The first order drainage basin, not a particular soil grouping, is the basic land unit for mapping and prediction.

3. Inclusion of antecedent flow (I) or a seasonal variable such as the sine of the day (S) allows several approaches to a current storage index that will be more directly related to stormflow production than antecedent 5-day rainfall.
Figure 7. A Plotting of Actual Versus Predicted Peakflow by Equation 16, Showing the Fit to the Original Data from all 11 Basins.
Figure 8. A Composite Graph Showing Comparison of Predictions by the Runoff Curve (SCS-TP-149) and the R-Index Methods on Four Basins not in the Original Data Set.
4. The R-index method is built upon a particular definition of stormflow and storm rainfall on source areas. If an R-index map is available, the method is much simpler and on early evidence more accurate than the curve number method when used for quick estimates on forests and wildlands of the humid East.

The R-index method contains fewer steps, tables and graphs, and eliminates a number of questionable assumptions about runoff from forest land that are implicit in methods now in use. At the same time, the method makes maximum use of the simplest information available to the land use planner and forest manager.

In conclusion, the authors recognize that 11 basins, regardless of the promise of the R-index method, constitute too small a sample to provide final estimates of the parameters in the stormflow model, or perhaps the exact best form of the model. Improvement calls for two things: First, existing small watershed data must be reduced to a standard format if the model hypothesized here is to be refined and further tested. Second, a substantial effort must be made to map the R-index on forest and wildland properties if the model is to serve a useful purpose. Existing hydrologic record, field surveys, short-term rainfall-streamflow stations, soil maps, aerial photographs, ERTS photographs and possibly other remote sensing techniques can be employed to reduce this task to feasibility.

ACKNOWLEDGMENTS

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