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A GIS-based variable source area hydrology model

Jane R. Frankenberger[1], Erin S. Brooks[2], M. Todd Walter[2]*, Michael F. Walter[3] and Tammo S. Steenhuis[3]

[1]*Assistant Professor, Department of Agricultural and Biological Engineering, Purdue University, West Lafayette, IN 47907-1146*
[2]*Research Associate, Department of Agricultural and Biological Engineering, Cornell University, Ithaca, NY 14853-5701*
[3]*Professor, Department of Agricultural and Biological Engineering, Cornell University, Ithaca, NY 14853-5701*

Abstract:

Effective control of nonpoint source pollution from contaminants transported by runoff requires information about the source areas of surface runoff. Variable source hydrology is widely recognized by hydrologists, yet few methods exist for identifying the saturated areas that generate most runoff in humid regions. The Soil Moisture Routing model is a daily water balance model that simulates the hydrology for watersheds with shallow sloping soils. The model combines elevation, soil, and land use data within the geographic information system GRASS, and predicts the spatial distribution of soil moisture, evapotranspiration, saturation-excess overland flow (i.e., surface runoff), and interflow throughout a watershed. The model was applied to a 170 hectare watershed in the Catskills region of New York State and observed stream flow hydrographs and soil moisture measurements were compared to model predictions. Stream flow prediction during non-winter periods generally agreed with measured flow resulting in an average r^2 of 0·73, a standard error of 0·01 m^3/s, and an average Nash-Sutcliffe efficiency R^2 of 0·62. Soil moisture predictions showed trends similar to observations with errors on the order of the standard error of measurements. The model results were most accurate for non-winter conditions. The model is currently used for making management decisions for reducing non-point source pollution from manure spread fields in the Catskill watersheds which supply New York City's drinking water.

KEY WORDS variable source hydrology; saturation-excess; runoff; subsurface lateral flow; distributed model; GIS; New York; GRASS; soil moisture routing model (SMR)

INTRODUCTION

An important watershed management strategy for controlling nonpoint source pollution is to minimise pollutant loading on runoff source areas; therefore, there is considerable need to accurately identify variable source areas. Saturated areas, which expand and contract seasonally, as well as during individual storms, have been shown to be the most important source areas of surface runoff in humid, well-vegetated areas (Dunne and Black, 1970; Hewlett and Nutter, 1970; Dunne, 1978). The term *variable source areas* has been adopted for referring to these saturated areas which often form where subsurface lateral flow converges, where the slope changes, or where depth to the restricting layer decreases.

Simple lumped models or hydrograph analysis can be used to estimate the percent of the total watershed area that produces surface runoff. For example, Steenhuis *et al.* (1995) presented a method consistent with the SCS curve number approach to predict the portion of a watershed contributing to runoff in shallow sloping soils.

* Correspondence to: M. Todd Walter, Department of Agricultural and Biological Engineering, Cornell University, Ithaca, NY 14853-5701, USA.

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In order to predict the spatial pattern of variable source areas, however, a distributed approach is needed. Many distributed, physically based models require large amounts of data and calibration, and have given mixed results (Bernier, 1985; Beven, 1989; Grayson *et al.*, 1992; Wigmosta *et al.*, 1994). Grayson *et al.* (1992) concluded that although complex process-based models are useful in research, models used for management decisions should be simple and unpretentious, with few data requirements and clearly stated assumptions.

One example of a simpler model which accounts for spatial variability is TOPMODEL (Beven and Kirkby, 1979; Beven, 1986), a semi-distributed model which falls between fully lumped and fully distributed models. TOPMODEL lumps hydrologically similar portions of a watershed based on a topographic index, ln(a/tan β), where 'a' is the upslope contributing area per unit length of contour and tan β is the local slope. Assumptions of steady state soil moisture distribution and an exponential decline in saturated conductivity with depth are necessary for relating moisture content (or deficit) to the topographic-soil index. The first assumption of steady state conditions, meaning that the entire potential subsurface contributing area contributes to flow, has been shown by Barling *et al.* (1994) to be a limitation. The second assumption, that soil conductivities decline exponentially with depth, may not be appropriate in shallow soils where saturation arises from perched water tables above a restricting layer, rather than a water table underlying the entire watershed. TOPMODEL has no mechanism for predicting these 'subsurface infiltration excess' processes (Beven *et al.*, 1994).

Although TOPMODEL provides some insight into spatial variability of runoff source areas, the inherent problems discussed above limit its applicability to watersheds which are hydrologically characterised by steep, shallow soils above a restricting layer. However, as an additional note, TOPMODEL is computationally very efficient due to its assumption of hydrologic similarity among areas with the same topographic index. This results in short model runtimes which can be an attractive feature. Nevertheless, faster computers and the widespread availability of digital geographic data, including elevation and soils, make the computational simplicity of lumped and semi-distributed models less advantageous.

The Soil Moisture Routing model (SMR), described here, is a simple distributed water balance model run on a daily time step, which predicts daily saturation-excess overland flow occurring at any point in a watershed. SMR's simplicity allows a physical modelling basis without over parameterisation. SMR's modest input requirements include: digital elevation data, soil parameters, and land use data for a distributed estimation of evapotranspiration and Hortonian flow. SMR was developed with the objective of aiding management decisions in potential runoff source areas of watersheds with steep hills and shallow soils. It was also equipped with routines for simulating the hydrologic impacts of diversions and subsurface drainage, two common water management practices. For applicability purposes, the model was designed to utilise readily available data and require essentially no calibration.

SMR development was initiated by a lack of reasonably parameterised, distributed models which could adequately describe the hydrology of the Catskills region of New York State. Interest in the area's hydrology is acute due to the potential of contamination of New York City's water supply reservoirs from agricultural non-point source pollution. SMR was developed specifically for topographically steep areas hydrologically characterised by relatively thin, permeable soil layers over a much less permeable fragipan, bedrock, or other restricting layer. This description is typical of upland soils in the glaciated region of New York State as well as many other places in the United States where a fragipan or bedrock limits root growth and water movement. The model is most effective where slopes are steep enough to be the main cause of lateral flow. SMR application is limited to regions fitting the description discussed above and should not be viewed as a general or universal hydrology model.

SOIL MOISTURE ROUTING MODEL OVERVIEW

The Soil Moisture Routing model is based on a hydrologic model for shallow soils of Steenhuis *et al.* (1986), adapted into the geographic information system (GIS), GRASS (U.S. Army CERL, 1991), by Caraco (1992) and Zollweg (1994) who used a similar model to predict stream flow event hydrographs.

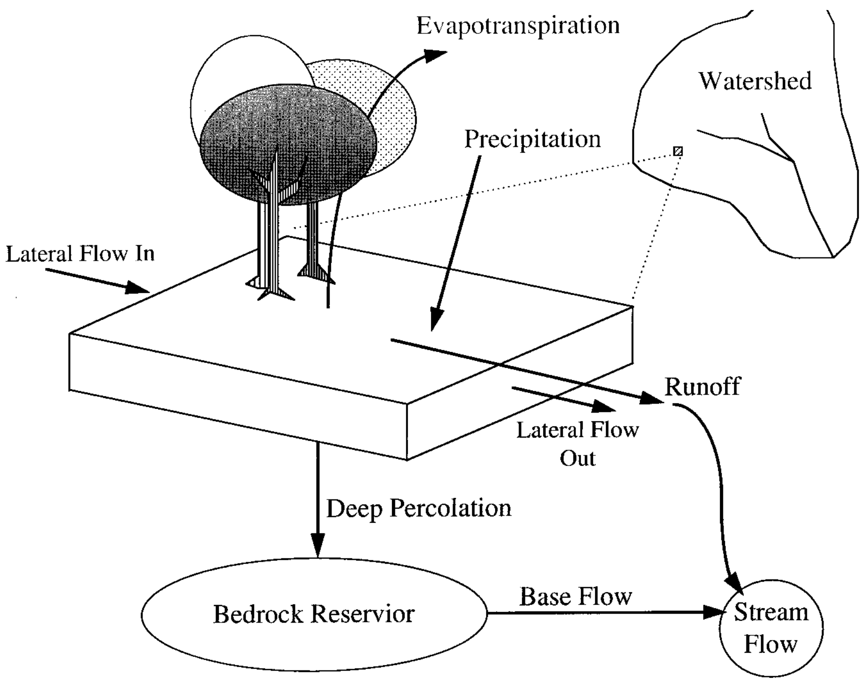


Figure 1. Conceptual hydrologic model

The model is 'tightly coupled' with the GIS (Stuart and Stocks, 1993), which means that it is written as a sequence of commands within GRASS. This coupling with a GIS simplifies data input and model calculations, as well as providing an efficient way to display and manipulate results. It also makes the processes clear for the user to understand and modify for different conditions. GRASS was chosen because it is public domain, widely available, and runs within the UNIX operating system on a variety of platforms. We have used the model on a SUN SPARC station as well as Linux running on a 386DX and Pentium PC.

SMR is based on a water balance at each time step for each cell of the watershed. Cells are typically of dimension 10 m to 30 m, so each square kilometer of watershed area is divided into 1100 to 10 000 cells, easily manipulated by the GIS. Soil moisture content for each cell is predicted, and any moisture above saturation results in surface runoff. Water inputs to each cell are daily precipitation and lateral flow from uphill cells, and outputs are lateral flow to downhill cells, percolation into the bedrock, evapotranspiration, and surface runoff (Figure 1). The calculations of the water balance and its individual elements are described later.

Below the soil layer is bedrock, which is often highly fractured. Since there is little hydrologic information concerning the bedrock, distributed modeling of percolated water is not feasible. In the model, water draining vertically out of the soil layer simply enters an uncharacterised 'bedrock reservoir', which eventually resurfaces as springs. Because the natural evolution of the landscape typically concentrates these springs near streams, the flow contribution from the bedrock reservoir is base flow for the stream. This base flow is assumed not to affect variable source areas. Though this assumption may be erroneous for some areas, especially near perennial streams, SMR is generally applicable to upland areas fitting the introductory hydrologic description. A lumped linear reservoir model, rather than a distributed model, is used to characterise the bedrock reservoir.

To predict variable runoff source areas, the input requirements for SMR are modest: elevation, soil, and land use maps (digital), six soil parameters, and three daily weather parameters: precipitation, potential evapotranspiration, and average air temperature. The six soil parameters are: depth to restrictive layer, type of restrictive layer (fragipan, bedrock, or impervious), saturated hydraulic conductivity, saturated moisture content, field capacity, and percent rock fragments. The land use map divides land uses among five categories: forest, grass, cropped, water, and farmstead. SMR uses the elevation map to estimate all topographic parameters, i.e., slope and flow direction. If stream flow estimates are desired, the only additional parameter needed is a linear bedrock reservoir coefficient to predict base flow. SMR automatically uses a coefficient based on the Catskills hydrology if a better estimate is not available. The effects of tiles and diversions may be incorporated into SMR simply by providing a digital map indicating the locations of these practices in a watershed.

Water balance

The water balance is calculated for each grid cell. GIS helps facilitate the water balance by providing a computational platform which simplifies the code for large array calculations. GIS stores input parameters such as soil data, slope, and flow direction for each cell. Soil moisture, for each cell, is updated for each time step using only simple GIS commands; for each cell:

$$D_i \frac{d\theta_i}{dt} = P - ET_i + \frac{\Sigma Q_{in.i} - \Sigma Q_{out.i}}{A} - L_i - R_i \tag{1}$$

where: i = cell address; D_i = depth to restrictive layer of the cell, (m); θ_i = average moisture content of the cell, (m^3/m^3); P = precipitation (rain + snowmelt), (m); ET_i = actual evapotranspiration, (m); $Q_{in.i}$ = lateral inflow from surrounding upslope cells, (m^3); $Q_{out.i}$ = lateral outflow to surrounding downslope cells, (m^3); L_i = leakage out of the surface soil layer to bedrock, (m); R_i = surface runoff, (m); A = area of a cell, (m^2).

These model components will each be examined in detail in the following sections. The model is run on a daily time step, a compromise between precision and speed for the model objectives of predicting variable source areas.

Precipitation

Precipitation consists of rainfall and snowmelt. Saturated hydraulic conductivities of soils in humid areas are generally higher than rainfall intensities, so all rainfall is assumed to infiltrate unless the soil is saturated or the area is disturbed or compacted.

When complete snow data is unavailable (depth and density), precipitation that occurs when the mean daily temperature is below 0°C is assumed to be snow, and remains in the snowpack until the mean daily temperature is above 0°C. A simple temperature index method is used to estimate snowmelt:

$$M = m_i T + k_i, \quad \text{if } T > 0°\text{C} \tag{2}$$

where: M = snowmelt (cm/day); T = average daily temperature (°C); m_i, k_i = melt factor (cm °C^{-1} day^{-1}) and constant (cm/day) respectively. The snowmelt factor, m_i, is 0·23 cm °C^{-1} day^{-1} in forested areas and 0·27 cm °C^{-1} day^{-1} in non-forested areas, while the constant, k_l, is 0 cm/day in forested areas and 1·22 cm/day in non-forested areas (U.S. Army Corps of Engineers, 1960).

Evapotranspiration

Evapotranspiration is calculated for each cell using the relationship developed by Thornthwaite and Mather (1955) as a function of daily potential evapotranspiration, vegetation and stage of growth, and moisture content in the cell:

$$\begin{aligned} ET_i &= c_i PET\left(\frac{\theta_i - \theta_{wp}}{\theta_{fc.i} - \theta_{wp}}\right) \quad \text{for: } \theta_i < \theta_{fc.i} \\ ET_i &= c_i PET \quad \text{for: } \theta_i \geqslant \theta_{fc.i} \end{aligned} \tag{3}$$

where: PET = potential evapotranspiration, (m); c_i = a vegetation coefficient from Jensen (1973), which varies throughout the year depending on vegetation type which is determined by land use class; land use in SMR; $\theta_{fc.i}$ = moisture content at field capacity as defined in the next section, (m^3/m^3); θ_{up} = moisture content at wilting point (m^3/m^3); θ_i = average soil moisture content of cell i, (m^3/m^3).

The actual evapotranspiration varies linearly between PET, when soil moisture content is at or above field capacity, and zero when soil moisture is below the wilting point. Determination of field capacity is discussed in the next section.

Subsurface lateral flow

Shallow subsurface lateral flow, or interflow, is a key component in the water balance. Lateral flow causes some areas, such as regions with convergent topographies, to be wetter than others, which often leads to the formation of saturated runoff source areas. The GIS, GRASS, is a convenient computational platform for routing subsurface lateral flow. Though subsurface lateral flow out of one cell and into an adjacent cell are physically coupled processes, SMR approximates each independently. The quantity of lateral flow leaving each cell is calculated and this flow is divided among all downhill neighbours as described below.

The quantity of lateral flow out of each cell is calculated from Darcy's Law and the kinematic approximation; i.e. the hydraulic gradient is equal to the land slope at each cell:

$$Q_{out.i} = wK_iD_i\left(\frac{dh}{dL}\right)_i \tag{4}$$

where: $Q_{out.i}$ =lateral flow out of cell l, (m^3); $(dh/dL)_i$ = land slope of cell i, (m/m); w = width of each cell, (m); K_i = hydraulic conductivity of the soil profile, (m/day); D_i = depth to restrictive layer, (m). The hydraulic conductivity, K_i, is dependent on soil moisture content, θ_i.

Because of the restricting layer at shallow depth, significant vertical water movement stops when the soil at the deepest part of the profile (i.e. directly above the restricting layer) becomes unsaturated. Field capacity of the soil layer above the restrictive layer is therefore defined as the average profile moisture content when the soil is just saturated at the fragipan interface (Steenhuis *et al.*, 1988). When soil moisture is above field capacity, a saturated layer of thickness $D_i(\theta_i - \theta_{fc})/(\theta_s - \theta_{fc})$ forms and the effective conductivity of the soil profile is:

$$K_i = (K_s - K(\theta_{fc.i}))\left(\frac{\theta_i - \theta_{fc.i}}{\theta_s - \theta_{fc.i}}\right) + K(\theta_{fc.i}) \quad \text{for: } \theta_s \geqslant \theta_i > \theta_{fc} \tag{5}$$

where: $\theta_{fc.i}$ = moisture content at field capacity as defined above, (m^3/m^3); θ_i = soil moisture content for cell i, (m^3/m^3); K_s = saturated hydraulic conductivity, (m/day); $K(\theta_{fc.i})$ = soil hydraulic conductivity at a field capacity, (m/day); (calculated with Equation (6) for $\theta = \theta_{fc}$).

If the average moisture content of the soil profile is less than field capacity, $\theta_{fc.i}$, the effective hydraulic conductivity is the unsaturated conductivity, $K(\theta_i)$. Based on the average moisture content of the profile, θ, unsaturated hydraulic conductivity is calculated using an exponential relationship:

$$K(\theta) = K_s \exp\left(\alpha\frac{\theta_s - \theta}{\theta_s - \theta_r}\right) \quad \text{for: } \theta \leqslant \theta_{fc} \tag{6}$$

where: θ_r = residual moisture content, (m^3/m^3); θ_s = saturated moisture content (assumed equal to porosity), (m^3/m^3); θ = soil moisture content, (m^3/m^3); α = a universal constant equal to 13 (Bresler *et al.*, 1978; Steenhuis and Van der Molen, 1986).

Each cell, potentially, has eight neighbouring cells. Lateral flow is divided among all neighbouring cells that are downhill from a particular cell. Allowing flow to divide into multiple flowpaths is particularly important in areas where topography may diverge. Some automatic drainage path algorithms route all downslope flow to one neighbouring cell, even though there may be two descending directions of equal slope (Tribe, 1992). This tends to lead to downslope accumulation of moisture into channels or gullies much more quickly than is actually seen in the field. Quinn *et al.* (1991) showed that TOPMODEL gave a more realistic prediction of wetness distribution when multiple flowpaths were allowed. The multiple flowpath division allocates to each neighbour a portion of the total flow, depending on the elevation difference between it and cell i, as well as on the distance between the cells. For any downslope neighbour j of cell i:

$$P_{ij} = \frac{(Z_i - Z_j)/L_j}{\sum_{j=1}^{n}[(Z_i - Z_j)/L_j]} \quad \text{for: } Z_i > Z_j \tag{7}$$

where: P_{ij} =the portion of the total flow out of cell i that is routed to neighbour, j; Z_i,Z_j = elevations of cell i and its downslope neighbour j, respectively; L_j = the distance from the center point of cell i to neighbour j; n =the number of downslope neighbours of cell i ($n \leqslant 8$).

Leakage out of the root zone layer

If a saturated layer is present, water can percolate into fractures or cracks in the bedrock or fragipan. An 'effective conductivity' of the bedrock and fragipan is specified which limits the rate at which water can leak out of the root zone. The bedrock in the Catskills is highly fractured. Flow takes place through the fractures in the bedrock and, to a lesser degree, in the dense fragipan (Soren, 1963). Though values vary widely, literature provides estimates of fragipan and bedrock conductivities (Theil and Bornstein, 1965; McCarty, 1980; Dabney and Selim, 1987; Smith and Wheatcraft, 1993). Typical reported fragipan conductivities range between 0·1 and 7 mm/day; SMR uses a value of 1 mm/day which is equal to the geometric mean of the extremes. Similarly, the SMR effective bedrock conductivity is 2 mm/day which is within the range of reported conductivities for bedrock similar in composition to the Catskills'.

If the SMR is used to predict stream flow, an estimate must be made of flow from the bedrock reservoir into the stream for each time step. Since little is known about the bedrock, the simple concept of a linear reservoir is used to estimate discharge. The reservoir coefficient is found from recession portions of stream flow hydrographs. Based on 168 recession events from three sub-basins in the Catskills' Cannonsville watershed over ten years, the SMR linear reservoir coefficient is 0·1 day^{-1} (mean $r^2 = 0{\cdot}96$) (Weiler, 1997).

Artificial drainage and compacted areas

The hydrologic impacts of diversions, tile drains, and ditches, common features on farms, can be simulated by SMR. These human-caused changes to the natural drainage may have a significant impact on watershed output and soil moisture distribution, particularly in locations adjacent to the drain. Although diversions are often constructed for the purpose of intercepting surface runoff in shallow soils, channels, as well as subsurface drains, intercept subsurface lateral flow from uphill. Lateral flow is assumed to be completely intercepted by the diversion or drain, and flow downslope starts as if there were a new watershed. Runoff from diversions and flow from tiles are summed over the watershed each day and added to daily stream flow.

Disturbed or compacted areas of the farm may generate Hortonian or infiltration-excess overland flow and are modeled by SMR as impermeable areas with surface detention storage as estimated by Miner (1967). These areas may include barnyards, other parts of farmsteads, or intensively cropped fields where infiltration has been reduced by growing continuous row crops or other poor field practices.

APPLICATION TO A SMALL WATERSHED

Location, climate, and land use

The model was tested on the Crowe Road Watershed, located in the northern Catskills region of the Appalachian plateau physiographic province. It has generally shallow soils and rolling topography, typical of upland watersheds in the region. Elevation in the 170 hectare watershed ranges from 580 to 732 m. The main stream draining the Crowe Road Watershed is an unnamed tributary of Wright Brook, which flows into the west branch of the Delaware River. This stream has been gauged since 1993 by the New York State Department of Environmental Conservation. A total of 23 months of 10 minute stream flow data were available. Average stream flow during the monitoring period was 0·030 m^3/s, with flows ranging from 0 to 0·5 m^3/s. Watershed location, elevation, and the main stream and pond can be seen in Figure 2.

The climate can be characterised as northern humid continental. The latitude is approximately 42°30′. Average yearly temperature is 8°C, with cold winters and warm summers. Average annual precipitation is 1123 mm and yearly stream flow for streams in the region averages 600 mm. Precipitation generally exceeds potential evapotranspiration except for four months in the growing season.

The entire watershed is owned by dairy farmers although, due to an abundance of steep rocky upper slopes with very shallow soil, more than 65% is forested. The lower slopes are mainly rotated corn and hay, with 2% in corn and 20% in hay during the 1994 growing season. The wetter, rockier areas are primarily permanent pasture (10% of the watershed). The remaining area is the farmstead (0·7%) and pond (0·8%). Land use for the watershed can be seen in Figure 3.

Geology and soils

The bedrock is of sedimentary origin, consisting of flat fractured layers of sandstone, siltstone, and shale (Soren, 1963). Glaciers have modified the landscape, rounding and smoothing the hills and depositing glacial till through much of the valley. Till, an unsorted mixture of particle sizes deposited by the glaciers, is the parent material for most soils in the watershed. Soil is generally very thin on hilltops and upper slopes, while in the lower slopes have deeper soils. A dense, brittle fragipan is found at a shallow depth in much of this area, roughly parallel to the land slope and relatively impervious to water. Soils have formed fairly recently in the glacial till, all soils being classified as inceptisols or entisols. The surface layer of most soils in the watershed is shallow and permeable, with a high percentage of rock fragments. Roughly half of the watershed consists of soils that overlie fractured bedrock and are well drained, while half overlie a dense fragipan layer and often have perched water tables during wet periods. A complete description of soils is given in Frankenberger (1996). Soils of the watershed can be found in Figure 4.

Soil data interpretation

Input data for the model were primarily obtained from the soil survey and other published literature and supplemented with field measurements. Although soil surveys are the best data source generally available, they were not developed with hydrologic modeling as an objective and, therefore, the parameters that determine soil moisture flow are not those that are emphasised in the soil survey (McKeague *et al*., 1984; Bouma, 1986; Soil Survey Staff, 1993). Depth and nature of the restricting layer were obtained from the typical pedon for that series in the soil survey and the volume occupied by rock fragments was obtained from the midpoint of the estimated range. Soil types and values used are shown in Figure 4 and Table I. Saturated conductivity, saturated moisture content (approximately soil porosity), and field capacity were obtained from field measurements. Using the auger hole method, the average measured saturated hydraulic conductivity was 2 m/day, somewhat higher than the permeability range of the soil survey (0·4 to 1·2 m/day for all soils). The differences are probably due to macropores which are not taken into account in the soil survey data. Porosity averaged 0·6 cm^3/cm^3 without rocks, and 0·45 cm^3/cm^3 when rock fragments were included. Field capacity, measured on soils draining to bedrock at a depth of 70 cm after a period of low rainfall

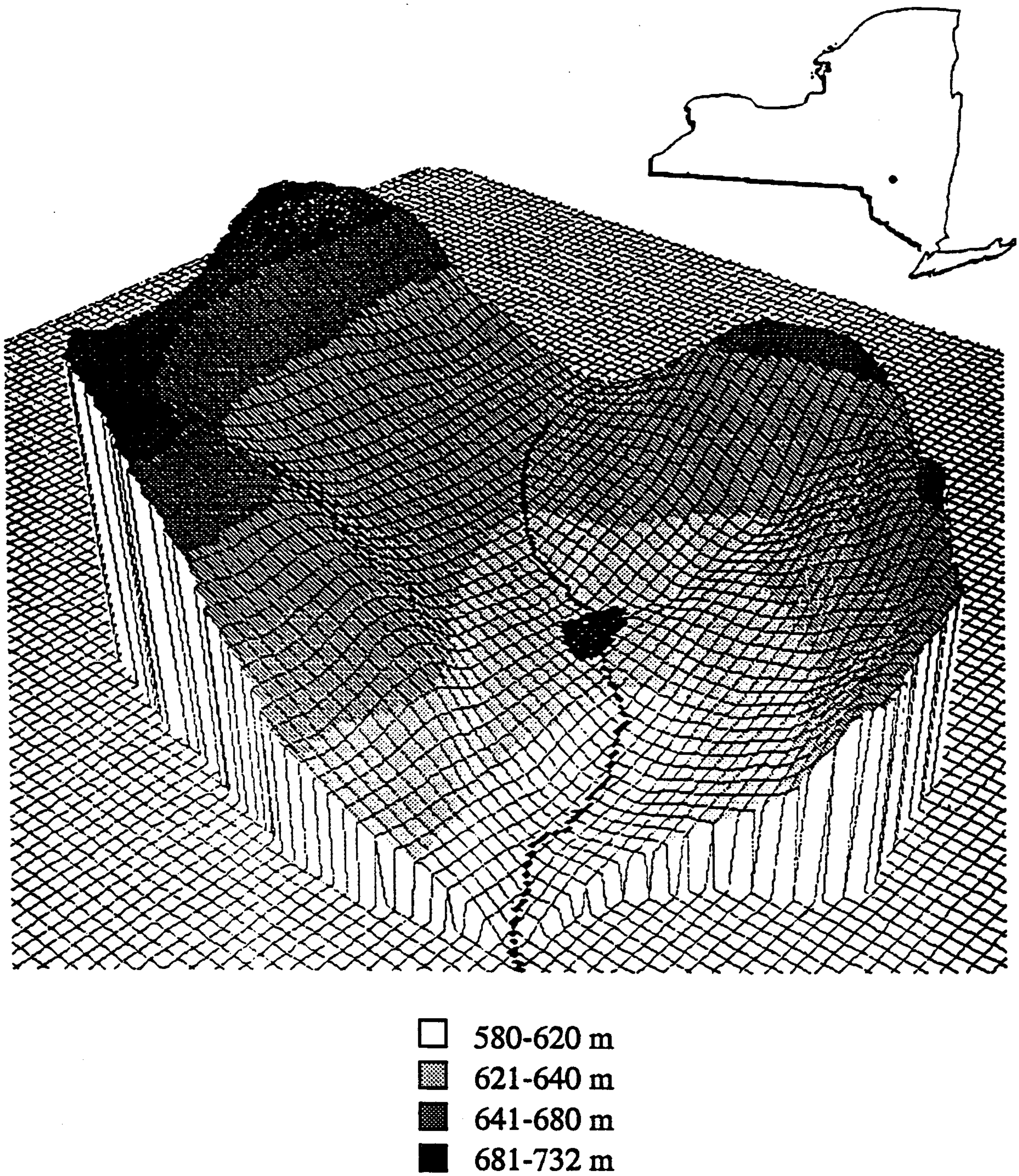


Figure 2. Watershed location and topography

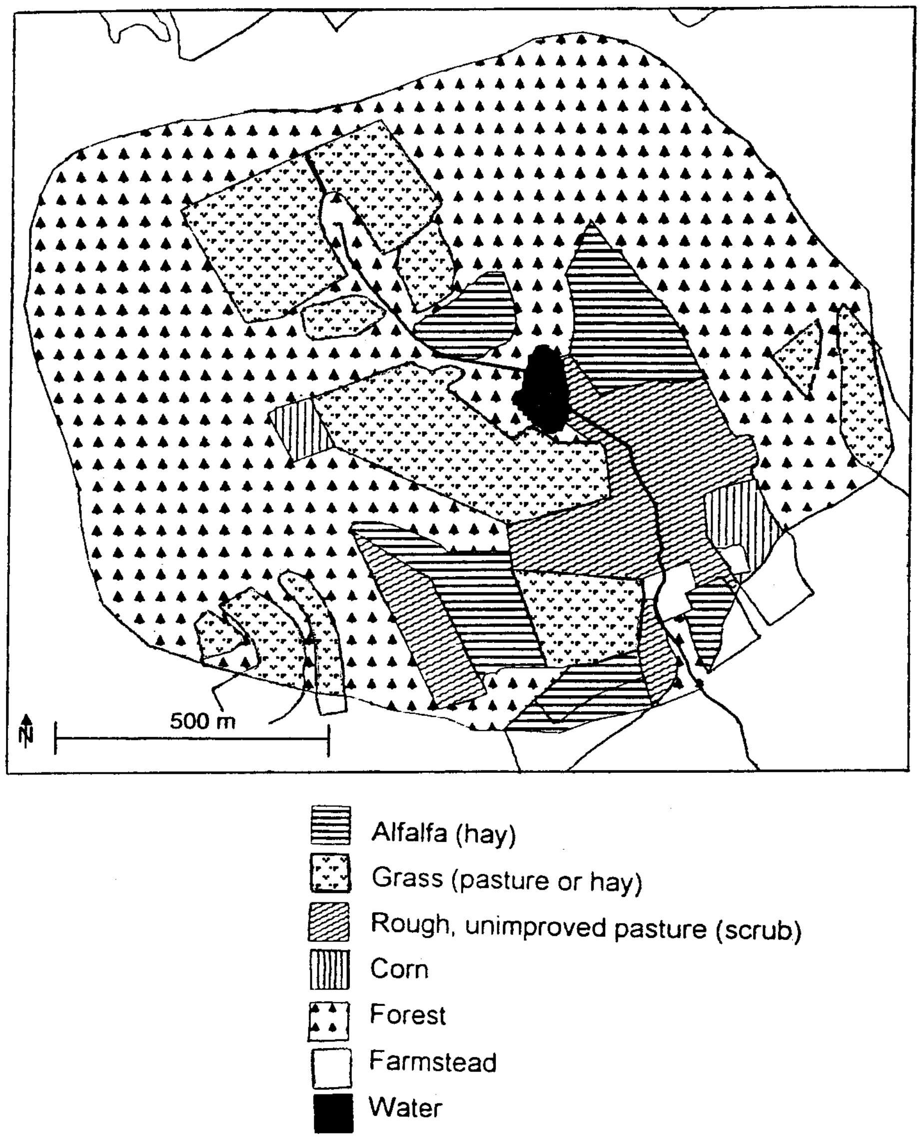
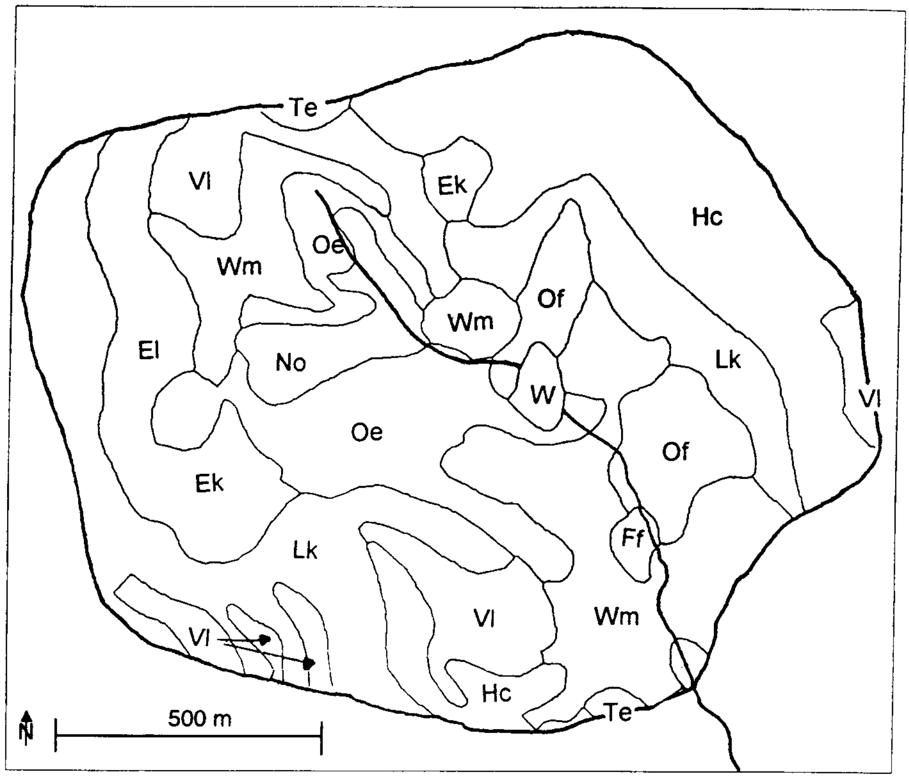


Figure 3. Land use in the Crowe Road watershed

in January, averaged 0·37 cm^3/cm^3 without rocks and 0·28 cm^3/cm^3 when rocks were included. Field capacity was assumed to vary linearly with depth to the restricting layer in other soils.

Weather data

The weather data needed by SMR are: daily precipitation, potential evapotranspiration, and average temperature. Precipitation was measured every 10 minutes with two recording gauges at the watershed outlet by the New York State Department of Environmental Conservation, and aggregated into daily values. During the growing season, pan evaporation was measured at a site approximately 30 km away and potential evapotranspiration was estimated from pan evaporation multiplied by a pan factor of 0·8. During the rest of the year, no local pan evaporation data was available so potential evapotranspiration was



Ek: Elka-Vly complex

El: Elka -Vly complex, very stony

Ff: Fluvaquents-Udifluvents complex, frequently flooded

Hc: Halcott, Mongaup, Vly complex, very rocky

Lk: Lewbeach and Lewbath, very stony

No: Norchip silt loam

Oe: Onteora channery silt loam

Of: Onteora and Ontusia, very stony

Te: Torull-Gretor complex

Vl: Vly channery silt loam

Wm: Willowėmoc channery silt loam

Figure 4. Soil series and complexes in the watershed

approximated from Hamon's (1961) method of calculating potential evapotranspiration which utilises potential hours of sunshine (from basic astronomical relationships) and saturated water vapor density at the daily mean temperature (from a psychrometric chart). Temperature was measured every 10 minutes near the watershed outlet after 8 December, 1993 and, before 8 December temperature was measured at a site 15 km from the watershed.

Table I. Soil input parameters from Soil Survey and field measurements

Soil	Soil Survey data[1]			Data used to model Crowe Road Watershed[2]					
	K_s (m/d)	Depth (m)	Rock (%)	K_s (m/d)	Depth (m)	Rock[3] (%)	θ_s (cm³/cm³)	θ_{wp} (cm³/cm³)	K_{rl}^4 (mm/d)
Ek	0·36–1·2	10–150	15–35	2	150	25	0·6	0·1	2
El	0·36–1·2	10–150	35–60	2	150	40	0·6	0·1	2
Ff	0·04–1·2	10–180	5–15	2	25	10	0·6	0·1	1
Hc	0·36–1·2	8–81	35–60	2	50	40	0·6	0·1	2
Lk	0·36–1·2	10–180	35–60	2	82	40	0·6	0·1	1
No	0·36–1·2	28–180	5–15	2	28	10	0·6	0·1	1
Oe	0·36–1·2	15–180	15–35	2	35	25	0·6	0·1	1
Of	0·36–1·2	15–180	35–60	2	35	40	0·6	0·1	1
Te	0·36–1·2	10–58	15–35	2	40	25	0·6	0·1	2
Vl	0·36–1·2	N/A	15–35	2	70	25	0·6	0·1	2
Wm	0·36–1·2	N/A	15–35	2	150	25	0·6	0·1	1

[1] The Soil Survey for Delaware County is currently unpublished (publication date: 2007) so not all data have been compiled; moisture content information was readily unavailable.
[2] Field measurements were used to obtain unavailable information, i.e., moisture content, and to refine Soil Survey information on saturated conductivity and soil depth to restricting layer.
[3] % Rock fragments were taken as the mean of the Soil Survey range.
[4] Conductivity of restricting layer based on major soil pedon (available through Soil Survey) from which the type of restricting layer could be determined (fragipan or bedrock) and published conductivity ranges as discussed in the text.

RESULTS AND DISCUSSION

A comparison between observed and predicted runoff source areas throughout the watershed would be the ideal demonstration of its adequacy. Unfortunately, making this comparison is technologically very complicated due to difficulty in identifying and monitoring source areas. In lieu of methods for comparing predicted and observed runoff source areas, comparisons between measurements and predictions for integrated outflow and distributed soil moisture along a transect were used to confirm SMR. While not ideal, the combination of these comparisons gives good insight into SMR's adequacy as a variable source hydrology model.

Integrated results: stream flow

Although predicting hydrographs is not the primary objective of SMR, stream flow integrates hydrologic response from across the watershed and, can therefore, be used in assessing the overal predictions of the model. Hydrograph generation was based on daily stream flow, which is the sum of surface runoff generated anywhere in the watershed, subsurface lateral flow into the stream, diversions, or tile drains, and a portion of the water stored in the bedrock reservoir each day. Figure 5 shows predicted vs. measured stream flow for 1993 and 1994 and Table II summarises statistical comparisons. Winter and summer results are segregated to isolate snow and other winter-time factors.

From Figure 5, predictions show good visual agreement with measured flow trends during periods without snowmelt. The Nash-Sutcliffe efficiency R^2 (Nash and Sutcliffe, 1970) were 0·59 and 0·64 for daily stream flow during summer 1993 (23 June to 30 November, 1993) and summer 1994 (1 May to 15 November, 1994), respectively. In other words, SMR predicted measured values 59% and 64% better than simply using the average stream flow value during these periods. The average correlation coefficient, r^2, for the summer periods was 0·73. Considering the absence of any major calibration, this correlation is good. Comparing the mean, maximum, and minimum stream flows also shows good statistical agreement between predicted and measured stream flows (Table II). During the summer, the standard error was about one order of magnitude less than the range of measured stream flows, indicating good predicted stream flow. One way to investigate

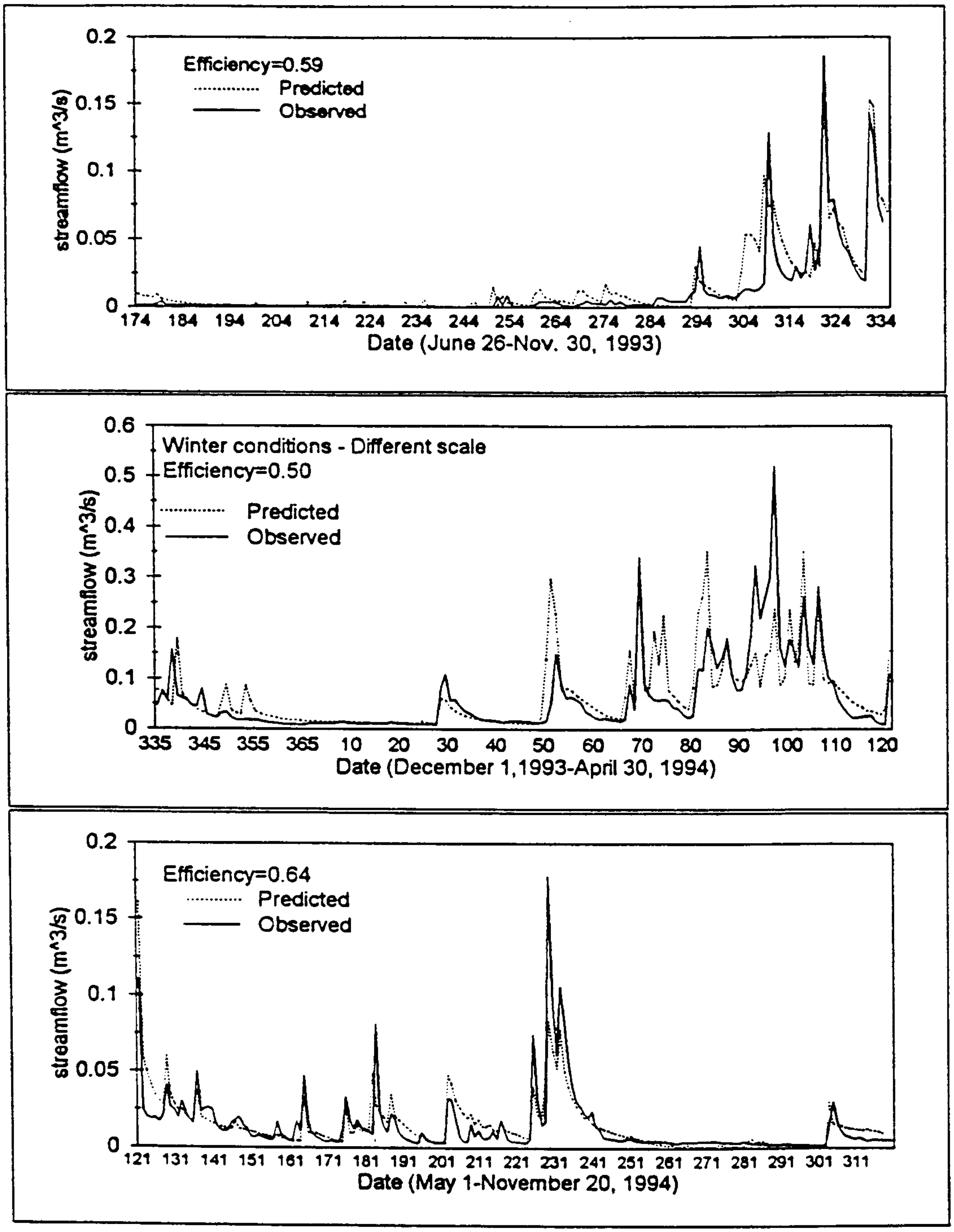


Figure 5. Predicted and measured stream flow with Nash-Sutcliffe efficiency for summer 1993, winter 1994, and summer 1994

Table II. Summary of stream flow statistics: predicted vs. measured

	Simulated			Observed			$r^{2(a)}$	ste[(b)]	$R^{2(c)}$
	Mean	Max	Min	Mean	Max	Min			
Summer 1993	0·015	0·18	0	0·012	0·19	0	0·82	0·01	0·59
Winter 1994	0·068	0·35	0·011	0·061	0·052	0·008	0·54	0·049	0·50
Summer 1994	0·015	0·16	$<10^{-3}$	0·014	0·18	0·002	0·66	0·012	0·64

Note: The units of Mean, Max, Min are m^3/s.
[a] Correlation coefficient.
[b] Standard error of the predicted to measured stream flow; m^3/s.
[c] Nash-Sutcliffe efficiency R^2 (Nash and Sutcliffe, 1970).

the accuracy of SMR is to see if its stream flow predictions are comparable to other accepted models. Even though, as discussed earlier, TOPMODEL is overly simplified in many respects, it was used, after calibration, to simulate the summer 1994 stream flow. TOPMODEL predicted flow with an efficiency of 0·59, similar to, but somewhat worse than, SMR for the same period. The similarity of results is not surprising because the lateral flow components of the two models are based on the same assumptions.

As expected, winter-time stream flows were not predicted well due to the absence of soil frost simulation and an overly-simplified snowmelt model. The Nash-Sutcliffe efficiency was 0·50 for winter 1994 (November 1993 through April 1994). During winter conditions, errors in precipitation measurements, and in snowmelt especially, accentuated errors in stream flow predictions. An early or late prediction in snowmelt results in poor stream flow correlation on both the day of the erroneously simulated melt and the day of the actual melt. This problem can be seen in Figure 5 by comparing the period around day 82, when SMR badly overestimated flow due to premature snowmelt in the simulation, to the period around day 98, when SMR underpredicted flow because of a melt event at the site. SMR might have predicted the snowmelt event around day 98 if it had not already predicted that all the snow melted around day 82.

Distributed results: soil moisture

Because the integrated output measured at the outlet of the watershed provides ambiguous information about the detailed hydrology within the watershed (Steenhuis *et al*., 1998), an effort was also made to assess distributed model predictions. Soil moisture, a state variable of the model which is updated at every time step, was measured in order to assess the distributed predictions of the model. During 1994 and 1995, soil moisture was sampled on sites representing a variety of soil types, land uses, and average moisture contents found in the watershed. Each cell's soil moisture was characterised by gravimetric sampling at three locations within the cell. For further details on sampling methods and results see Frankenberger (1996).

Due to variability within a 30 m^2 area, there is difficulty in characterising the soil moisture of a cell with a single value. In an attempt to account for this microvariability, two cells representing different moistures were sampled at thirteen locations and average soil moisture at each location was estimated by averaging gravimetric soil moistures from two depths. The average standard deviation of measured soil moisture for the two cells was 0·07 cm^3/cm^3, representing both spatial variability and sampling error. Assuming 0·07 cm^3/cm^3 is a good approximation of the spatial variability of soil moisture content in a cell, 0·04 cm^3/cm^3 is the standard error of the mean obtained with three samples per cell. This provides some guidance in assessing the performance of SMR in predicting soil moisture, i.e., predictions within 0·04 cm^3/cm^3 are within errors in measured mean cell soil moisture.

Figure 6 shows predicted and measured soil moisture for all the samples throughout the watershed. The standard error was 0·044 cm^3/cm^3, approximately the same as the errors in measuring mean soil moisture for a cell. Figure 7 shows SMR predicted and measured soil moisture for three sampling dates along a transect: 6 May, 1994, was moderately wet; 28 October, 1994, relatively dry; and 18 January, 1995, was very wet (although the soil was not frozen). The transect ran uphill with a 14% average slope, intersected three

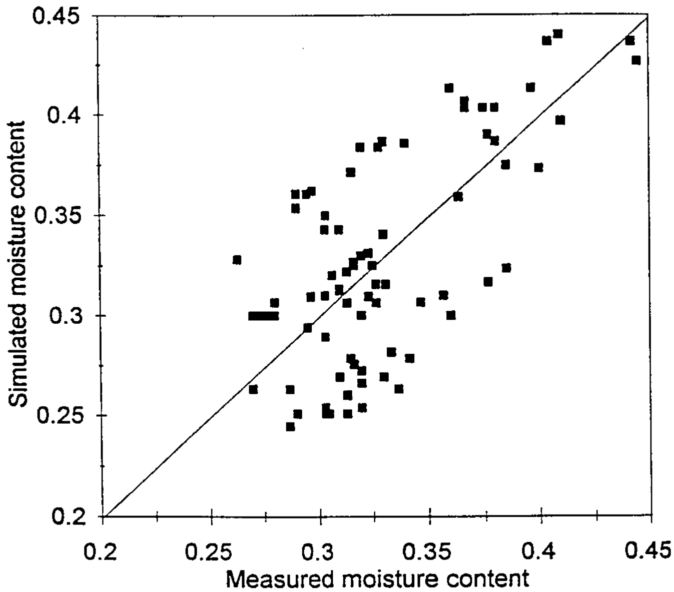


Figure 6. Predicted vs. measured soil moisture for all monitoring dates. Standard error is 0·044 cm^3/cm^3 and the r^2 is 0·45 over a range of soil moistures from 0·25 cm^3/cm^3 to 0·45 cm^3/cm^3

primary soil types, and two main land uses (cropland and pasture). The area upslope of the transect was forest.

A summary of statistical comparisons between predicted and measured soil moisture along the transect is presented in Table III. The sampling transect had considerable spatial variability in topography and, consequently, in moisture content under wet conditions. Steeper areas became much drier than flat areas when conditions were wet enough for a saturated layer to form over the bedrock or fragipan (6 May, 1994 and 18 January, 1995), while topography had little effect during drier times (28 October, 1994). The model correctly predicted the general pattern of soil moisture, particularly the increase in soil moisture from the top to the bottom of the slope. 6 May, 1994 showed particularly good agreement in downslope wetting; SMR simulated a 31% increase from 330 m to the bottom of the slope which was relatively close to the 26% increase measured (Figure 7). 28 October, 1994 showed the least statistical correlation between predicted and measured results because the total downslope change in measured soil moisture was only slightly higher than the standard error and almost all the change occurred in the bottom 30 m. The standard error gives a better indication of model and measurement agreement, 0·009 cm^3/cm^3, which is an order to magnitude lower than the measurement error.

Table III. Summary of soil moisture statistics: predicted vs. measured

	Simulated			Observed			ste[(a)]	$r^{2(b)}$
	Mean	Max	Min	Mean	Max	Min		
6 May 1994	0·333	0·38	0·29	0·334	0·39	0·27	0·027	0·91
28 Oct 1994	0·30	0·31	0·28	0·29	0·36	0·25	0·009	0·43
18 Jan 1995	0·38	0·45	0·32	0·34	0·45	0·24	0·03	0·87

Note: The units of Mean, Max, Min are cm^3/cm^3.
[a] Standard error of the predicted to measured soil moisture; cm^3/cm^3.
[b] Correlation coefficient between predicted soil moisture and the moving average of the measured soil moisture.

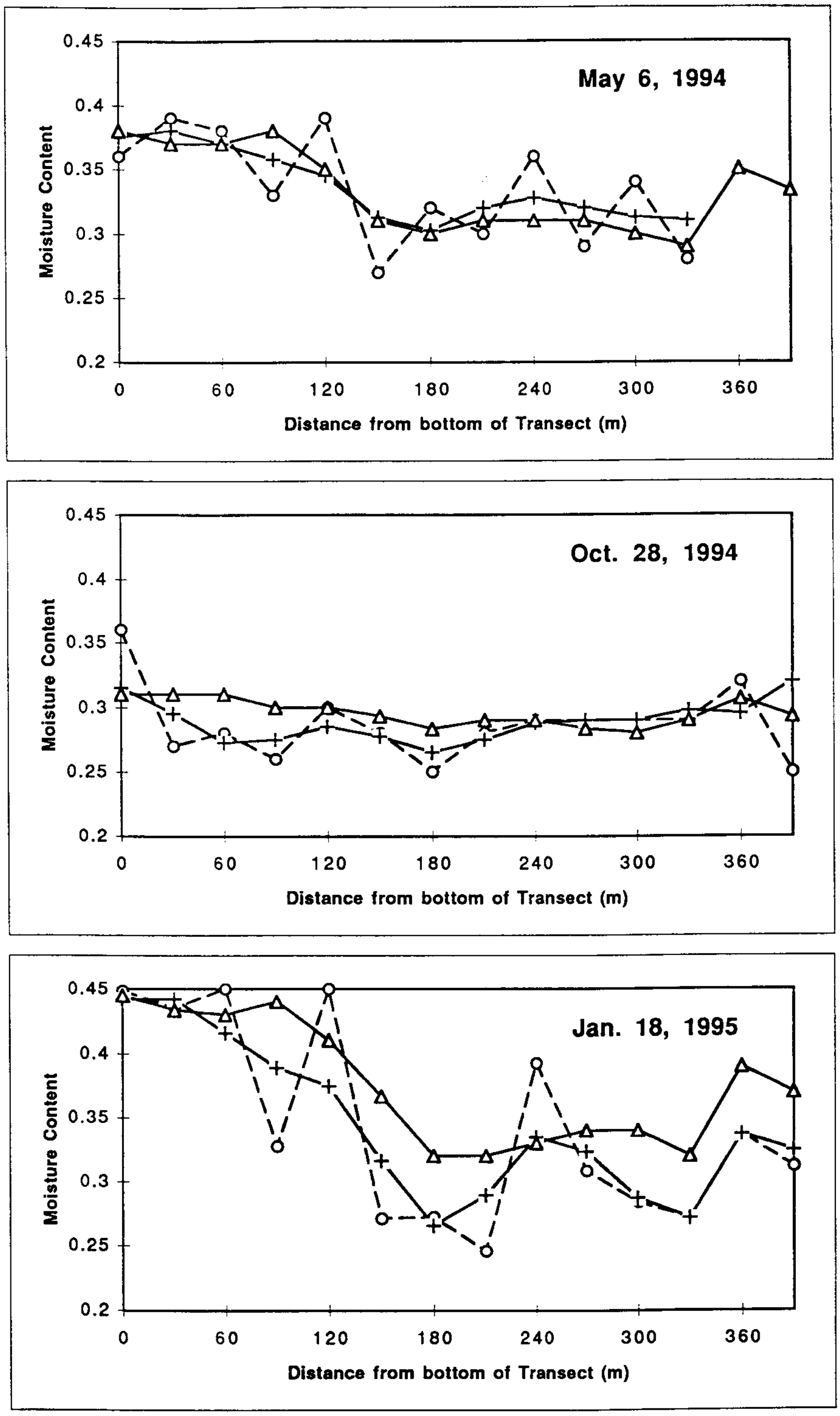


Figure 7. Predicted and measured soil moisture along a transect for three sampling dates: 6 May 1994 (wet), 28 October 1994 (dry), and 18 January 1995 (wet-winter). ○ = observed; Δ = predicted with SMR; + = moving average of observations.

SMR also correctly predicted the observed dip in soil moisture between 204 m and 120 m from the slope bottom. On all three dates, the dip was evident in both the predicted and measured soil moistures, although the predicted results underestimated the magnitude of the dip. This may be due to scaling issues. Large cell sizes will tend to remain wetter because cell slope will decrease, thus decreasing the rate of subsurface lateral flow out of a cell (Kuo *et al.*, 1998).

Some localised variations were not consistently predicted by SMR, such as the rapid change in moisture content between 90 m and 120 m. This observed trend may be due to a localised anomaly such as a 'trench' in the underlying bedrock or fragipan which, as McDonnel *et al.* (1996) showed, may have significant impacts on hydrologic predictions over small areas on the same order of magnitude as a SMR cell. In general, the SMR model predicted a smoother transition in soil moisture along the transect than was measured, although the standard error of predicted values for all three days were below the error in measured average cell moisture content, 0·04 cm^3/cm^3. Using a moving average to moderate the localised effects of microtopography on soil moisture gives good correlations, as shown in Figure 7 and Table III. 28 October 1994's correlation coefficient, r^2, is low even with the moving average because, as stated earlier, changes in moisture content along the slope are on the same order as errors in measurement.

CONCLUSIONS

The widespread availability of digital geographic data, particularly digital elevation models, opens new opportunities for using distributed models in watershed planning. At the same time, variability of soils, as well as lack of subsurface data, makes the use of complex models unrealistic in the field. The Soil Moisture Routing model uses elevation, soil, and land use data in a simple way to estimate soil moisture and runoff in watersheds with shallow soils. Its use of GIS to keep track of all parameters and state variables makes it easy to use and modify for different conditions. Results from the integrated output (stream flow) and the distributed output (soil moisture) show that it adequately predicted stream flow (runoff) and soil moisture distribution in the Crowe Road Watershed.

Despite its promise for use as a management tool, many sources of uncertainty exist in SMR predictions. Beven and Binley (1992) have listed the sources of error in hydrologic models as: (1) deficiency of model structure, (2) input data or boundary condition error, and (3) error associated with measurements used in model calibration. Deficiencies in model structure include the simplification of soil moisture relationships, infiltrated precipitation instantaneously distributed through the soil profile, no re-infiltration of runoff generated upslope, deficient snowmelt algorithm, absence of soil frost model, and the simplification of the bedrock hydraulics. Uncertainty in input data is a large source of error in the model; however, we believe that the simplified structure of the model is in balance with the level of data available for most watersheds. The greatest source of uncertainty lies in measuring distributed hydrology in a way that allows for meaningful comparisons between predicted and measured values. Though it was only briefly discussed in this paper, scale issues may be a significant source of model error and are definitely an uncertainty at this point.

Although refinements and further research are needed, SMR, in its present form, has proved useful in assisting with management decisions in the Catskills region, such as location and timing of manure spreading on farms to minimise contamination of the New York City reservoirs. It is also being used to evaluate the most effective 'best management practices'.

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