Hydrologic assessment of an urban variable source watershed in the northeast United States

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Received 30 March 2006; revised 15 September 2006; accepted 10 October 2006; published 10 March 2007.

Effective control of nonpoint source contaminants in runoff from urbanized watersheds requires knowledge about the locations of runoff source areas under a variety of conditions. Physical monitoring of spatially variant processes, such as runoff production form variable source areas, is time-consuming and expensive. Thus modeling the processes may provide a valuable cost-effective alternative. In this paper we adapt and validate a variable source model for an urban watershed to predict areas of the landscape prone to elevated soil moisture levels and saturation excess runoff. We modified the soil moisture distribution and routing (SMDR) model to simulate hydrologic processes in an urban upstate New York watershed by considering the impact of impervious surfaces, hydraulic control structures (detention basins), and land use on the water balance. In our model, infiltration excess runoff from impervious surfaces infiltrates in the surrounding soils. Simulated and observed streamflow agreed well, and more importantly, the distribution of soil moisture levels and overland flow throughout the watershed were well predicted. Removing urban features from the model resulted in substantially lower peak stormflows than observed, and the base and interflows increased accordingly. Both modeled and measured distributed results indicated that the more urbanized areas of the study site had both higher soil moistures and runoff losses due to shallower soils, greater upslope contributing areas, and a larger area of impervious surfaces generating runoff. The model results helped identify processes that describe how urbanization impacts integrated and distributed hydrology, which provides useful information for targeted water quality management.


1. Introduction

Urbanization can dramatically alter the hydrologic response of a watershed. For example, studies in urban watersheds have documented increased streamflow, reduced time of concentration [Wissmar et al., 2004; Murdock et al., 2004], altered soil moisture levels [Carlson and Arthur, 2000; Arthur-Hartranft et al., 2003], increased runoff losses [Corbet et al., 1997], and reduced albedo [Schmid et al., 1991]. Furthermore, urban areas can alter the transport of nutrients and other contaminants to surface water bodies [Interlandi and Crockett, 2003], which result in increased primary productivity [Smart et al., 1981] and impaired water quality. There have been many studies assessing the impact of agricultural activity on nonpoint contaminant sources [Eckholm et al., 2000; Sharpley et al., 2001; Andraski and Bundy, 2003], but relatively few comprehensive studies characterizing the effects of urbanization on watershed scale hydraulic response and water quality [Rodriguez et al., 2003; Carle et al., 2005].

Many areas in the mountainous Northeast United States are characterized by shallow permeable soils overlaying a less permeable restricting layer [Dunne and Black, 1970]. Since the soil permeability of these areas is generally greater than the precipitation rate, runoff generation is predominantly by saturation excess. A saturated area may form during extended rainfall events where subsurface flow converges with the soil surface [Frankenberger et al., 1999], the slope flattens [Peters et al., 1995], the depth to the impervious layer decreases [Ogden and Watts, 2000], or the soil otherwise has little additional storage capacity [Dunne, 1978]. However, the spatial-temporal extent of these areas, termed variable source areas (VSA), is dynamic and not easily predicted. Urban areas can further confound predictions of the spatial extent of VSAs due to impervious surfaces and other man made structures that alter natural hydrologic processes. Therefore modeling distributed processes can be difficult. This is particularly true for most urban models, which are generally non/distributed or lumped in nature and therefore often fail to provide realistic estimates of distributed responses such as changes in soil moisture levels and transport.
moisture or runoff generating areas [Valeo and Moin, 2000]. The complexity and limitations of many urban hydrology models prompts formulation of a simple model that can be used with easily available data to produce results that can aid in assessing the impacts of urbanization on water resources.

[5] Numerous urban storm water models have been developed ranging from statistical regression based models [Driver and Troutman, 1989; Vaze and Chiew, 2003] to fully distributed models using variants of physically based equations (Darcy’s law, Hortonian flow, Richards equation, Laplace equation, etc.) approximating relevant physical manifestations of the processes. Regression models have the advantage of simplicity, often relating event loads to catchment characteristics through empirical equations [Vaze and Chiew, 2003], but cannot be confidently applied outside the region where they were created or outside their range of calibration. Process-based models are more complex, data intensive, expensive, and require considerable expertise on the part of the user [Bhaduri et al., 2001], but allow the user to model the process in a spatial temporal field. For example, the semidistributed storm water management model [Metcalf and Eddy Inc., 1971; Huber and Dickinson, 1988] is a commonly used urban storm water management tool that has extensive data requirements (1-hour rainfall, Manning’s roughness coefficients for all surfaces, depression storage for all surfaces, etc.) and requires extensive parameterization and calibration [Balascio et al., 1998], but generally provides reasonable catchment outlet results. A common problem with current urban water quality models, such as the long-term hydrological impact assessment model [Bhaduri et al., 2001; Lim et al., 2006], is that they use soil conservation curve number approach to predict runoff, which links runoff response to soils and land use in ways that fail to capture the dynamics of VSA hydrology. In VSA watersheds this approach may not adequately capture the spatial extent of contributing areas, as they are, in many cases, independent of soil hydrologic group and more controlled by topography.

[5] While integrated or watershed outlet model predictions of many models may agree with measurements, this does not mean that distributed hydrological processes are correctly captured. Indeed, models with different structures (i.e., fully distributed, semidistributed, lumped) can perform equally well in modeling streamflow, and other integrated responses [Franchini and Pacciani, 1991; Johnson et al., 2003]. For instance, the soil and water assessment tool (SWAT) currently has no water flow among hydrologic response units within a subbasin and therefore no capability to determine different sources (VSAs) of runoff and pollutants in the landscape [Arnold and Fohrer, 2005]. In some cases lumped models, i.e., TOPMODEL [Beven et al., 1984] are capable of delineating saturated areas or VSAs on a larger scale, but may not adequately capture soil moisture or runoff production [Beven, 1997] from heterogeneous landscapes such as in urban areas. Few models couple simplicity with a distributed output. Consider the distributed hydrology soil vegetation model (DHSVM) [Nijssen et al., 1997], the application of which is limited to areas where distributed meteorological inputs can be found or where error introduced by modeling the distributed inputs is not problematic. For instance, Storck et al. [1998] describe the minimum metrological inputs for DHSVM as direct beam and diffuse short-wave radiation, air temperature, relative humidity, wind speed, and precipitation. Additionally, DHSVM requires calibration of the integrated and distributed output, for which, in many watersheds, the data does not exist. When an integrated response variable such as streamflow is used to calibrate a distributed model more than one combination of parameters may result in the best fit, commonly called equifinity [Binley and Beven, 1991]. Grayson et al. [1992] state that distributed models with numerous calibration parameters should be used for hypothesis testing or research purposes, not for management applications. For management applications, distributed models should avoid calibration; utilize simple input data, and clearly state hydrologic assumptions [Brooks et al., 2007]. One of the few models that requires little or no calibration, uses easily available input data, and generates a spatially distributed output is the soil moisture distribution and routing model (SMDR) [Zollweg et al., 1996; Frankenberger et al., 1999; Kuo et al., 1999, Mehta et al., 2004].

[6] SMDR was developed as a tool to locate areas in the landscape contributing runoff from transient, shallow perched water tables with applications to shallow, sloping soils underlain by a restrictive layer. SMDR has been shown to provide accurate estimates of hydrologic processes such as streamflow and soil moisture distribution in forested and agricultural watersheds [Zollweg et al., 1996; Frankenberger et al., 1999; Walter et al., 2000; Johnson et al., 2003; Mehta et al., 2004; Gérard-Marchant et al., 2005], but has yet to be applied to urban areas.

[7] Because of the spatially heterogeneous nature of urbanized watersheds (i.e., roads, roofs, lawns, wooded areas, varying on a fine spatial scale) a distributed model is necessary to correctly characterize the distribution of hydrologic processes. The distributed nature of SMDR allows comparison of spatially explicit variables such as soil moisture and runoff generating areas, and can be used to assess the impact of urban development on water quality, allowing land managers to better target critical source areas. Municipalities would benefit from the urban version of SMDR, outlined here, as a tool to assess zoning regulations and reduce or better plan development in sensitive areas.

[8] The purpose of this study was to model the integrated and distributed hydrologic response of an urbanizing watershed with a variable source model using available data. To gain insight into the effect of urbanization, the urban watershed response was compared to the modeled watershed response prior to development. We examined the integrated (stream discharge) and distributed output (soil moisture and runoff losses from the landscape) to assess model accuracy, predictive ability, and management implications. We discuss the model results and the insight given by the model into the hydrology of this and similar urban watersheds.

2. Watershed Description

[9] The model was applied and validated on a 332 ha urbanizing watershed located in Ithaca and Lansing, New York, (42°48′N, 76°46′W) in the Appalachian Plateau physiographic province (Figure 1). The region is typified by steep hillslopes with flattened hilltops of glacial origins with shallow permeable soils, underlain by a restrictive
The climate is humid with an average annual temperature of 7.7°C and average annual precipitation of 1143 mm. Elevation in the watershed ranges from 250 to 350 m above mean sea level with an average slope of 8.5% and slopes as steep as 27% mainly near the watershed outlet. Slopes in the upper watershed are between 2 and 5%. Soils are generally deeper in the upper reaches of the watershed and underlain by bedrock at depth greater than 1 m while soil depth is less near the watershed outlet and underlain by fragipan at depths between 30 and 80 cm (e.g., coarse-loamy, mixed, active, mesic, to frigid Typic Fragu- depts, Lytic or Typic Dystrudepts common to glacial tills). The lower watershed is predominantly urban (40% of total watershed) while the upper watershed is forested (44%), water/wetland (8%), and pasture (8%) (Figure 1). The urban area is a mix of home landscapes (lawns, woods, and impervious areas) as well as parks, schools, and commercial development. Impervious surfaces comprise 24% of the lower watershed (Figure 2), (i.e., between the detention pond and outlet in Figure 1). The overall watershed outflow was measured with an Isco 6712 stream gauge with a 750 area velocity module (Isco Inc, Lincoln, NE) (Figure 1).

For validation of the distributed model response, soil moisture levels and runoff were monitored on 4-m² plots in the watershed from April 2003 to April 2005 (Table 1 and Figure 1). During, or directly following nine runoff events and four dry periods, the soil moisture content was measured at the (0–30 cm depth) from the 12 locations 1–12 in Figure 1, (three samples per plot, bulked) and compared to the modeled soil moisture level. Samples were weighed wet, dried at 104°C for 48 h, and reweighed to determine water content. The saturation degree was determined by dividing the volumetric water content by the porosity.

Because of the inherent variability in soil parameters a frame of reference is necessary for assessing model accuracy. Establishing error estimates for the soil moisture content of watershed soils was done by sampling moisture contents at 15 locations within a 10 × 10 m area, repeated at 42 locations in the watershed, and determining the moisture content standard error within each sampling area. The average within-cell standard error of the mean measured soil moisture was 0.026 cm⁻³, which translates to 1.32 mm of water for a representative soil. The standard error represents both spatial-temporal variation and sampling error. This was used as a frame of reference for assessing model predictions, i.e., if the standard error of the predicted to measured saturation degree, is within 0.026 cm⁻³ the
## Table 1. Land Use, Hillslope Position, and SMDR Required Soil Properties for the Landscape Sampling Locations in Figure 1

<table>
<thead>
<tr>
<th>Landscape Sampling Location (Land Use)</th>
<th>Slope Position</th>
<th>( \eta ), cm(^3) cm(^{-3})</th>
<th>( \rho_b ), g cm(^{-3})</th>
<th>( z ), m</th>
<th>( \theta_{FC} ), cm(^3) cm(^{-3})</th>
<th>( \theta_{WP} ), cm(^3) cm(^{-3})</th>
<th>( K_s ), m d(^{-1})</th>
<th>RFC, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 (grass right of way)</td>
<td>toe</td>
<td>0.57</td>
<td>1.4</td>
<td>0.30</td>
<td>0.41</td>
<td>0.21</td>
<td>2.3</td>
<td>1.1</td>
</tr>
<tr>
<td>2 (wooded scrub)</td>
<td>face</td>
<td>0.61</td>
<td>1.1</td>
<td>0.35</td>
<td>0.36</td>
<td>0.20</td>
<td>2.9</td>
<td>14.1</td>
</tr>
<tr>
<td>3 (grass well maintained)</td>
<td>face</td>
<td>0.52</td>
<td>1.3</td>
<td>0.65</td>
<td>0.34</td>
<td>0.19</td>
<td>6.7</td>
<td>1.1</td>
</tr>
<tr>
<td>4 (grass poorly maintained)</td>
<td>toe</td>
<td>0.57</td>
<td>1.1</td>
<td>0.49</td>
<td>0.39</td>
<td>0.21</td>
<td>2.7</td>
<td>1.1</td>
</tr>
<tr>
<td>5 (wooded successional)</td>
<td>toe</td>
<td>0.50</td>
<td>1.5</td>
<td>0.55</td>
<td>0.32</td>
<td>0.17</td>
<td>4.0</td>
<td>14.8</td>
</tr>
<tr>
<td>6 (grass well maintained)</td>
<td>face</td>
<td>0.49</td>
<td>1.4</td>
<td>1.05</td>
<td>0.20</td>
<td>0.10</td>
<td>6.1</td>
<td>15.4</td>
</tr>
<tr>
<td>7 (wooded successional)</td>
<td>face</td>
<td>0.50</td>
<td>1.3</td>
<td>0.85</td>
<td>0.25</td>
<td>0.14</td>
<td>7.9</td>
<td>13.7</td>
</tr>
<tr>
<td>8 (grass poorly maintained)</td>
<td>toe</td>
<td>0.49</td>
<td>1.4</td>
<td>1.05</td>
<td>0.20</td>
<td>0.10</td>
<td>6.1</td>
<td>15.4</td>
</tr>
<tr>
<td>9 (grass well maintained)</td>
<td>face</td>
<td>0.49</td>
<td>1.3</td>
<td>1.00</td>
<td>0.23</td>
<td>0.10</td>
<td>7.9</td>
<td>16.3</td>
</tr>
<tr>
<td>10 (wooded scrub)</td>
<td>flat</td>
<td>0.51</td>
<td>1.9</td>
<td>1.13</td>
<td>0.25</td>
<td>0.13</td>
<td>7.9</td>
<td>13.5</td>
</tr>
<tr>
<td>11 (seasonal wetland)</td>
<td>flat</td>
<td>0.51</td>
<td>1.9</td>
<td>1.13</td>
<td>0.25</td>
<td>0.13</td>
<td>7.9</td>
<td>13.5</td>
</tr>
<tr>
<td>12 (grass right of way)</td>
<td>face</td>
<td>0.49</td>
<td>1.3</td>
<td>0.95</td>
<td>0.22</td>
<td>0.10</td>
<td>7.9</td>
<td>26.4</td>
</tr>
</tbody>
</table>

1Landscape sampling location one is closest to the watershed outlet.

2Porosity (\( \eta \)), bulk density (\( \rho_b \)), depth (\( z \)), field capacity (\( \theta_{FC} \)), wilting point (\( \theta_{WP} \)), saturated hydraulic conductivity (\( K_s \)), and rock fragments (RFC).

Model predictions are at least as accurate as the measured within-cell variability and sampling error.

[12] The first nine sampling locations in Figure 1 (Table 1) also served as surface runoff collection locations. Runoff was collected on the down gradient side of the plot using an H flume and directed to a tipping bucket sensitive to 0.1 mm. Plots were bordered to prevent overland flow entering the from upslope areas (subsurface flow could move below boarders). Thus we have measures of runoff from known areas to compare with model predictions. Measured and modeled runoff depths were compared for 15 events, spanning over a range of event types (i.e., prolonged steady precipitation, snowmelt, and intense thunderstorms). The landscape sampling locations represent the range of typical northeast US urban landscapes, including fertilized lawns, unfertilized areas, wooded areas, and right of ways (Table 1).

[13] The distributed measurements were compared to the distributed model predictions of soil moisture and runoff losses using visual methods (mapped area) and statistical methods including the Nash-Sutcliffe efficiency (E) [Nash and Sutcliffe, 1970], which measures a models predictive capacity against the observed mean; the coefficient of determination (\( r^2 \)), which measures how well modeled values correlate with observed values; and the standard error as described above.

## 3. Soil Moisture Distribution and Routing Model Overview

[14] SMDR is a physically based, fully distributed, water balance model for shallow, sloping soils underlain by a restrictive layer [Steenhuis and van der Molen, 1986; Zollweg et al., 1996; Kuo et al., 1999; Frankenberger et al., 1999; Mehra et al., 2004; Gérard-Marchant et al., 2005]. The SMDR framework is adapted to run on the geographic information system (GIS) GRASS [U.S. Army Construction and Engineering Laboratory, 1997; Neteler and Mitasova, 2002]. In current form SMDR incorporates two soil layers, the effects of macropore flow on soil hydraulic conductivity, the effects of impervious surfaces on the water balance, and control structures operating on a 10 m grid.

[15] Figure 3 illustrates the conceptual mechanisms of the water balance in the two-layered system. For each pervious grid cell at each time step in the watershed the water balance is:

\[
D \frac{d\theta}{dt} = R - E_o + \frac{\sum_{i=1}^{i} (Q_{in,i} - Q_{out,i})}{A} - P - SE_{out,i} + IER_{in,i}
\]

where \( i \) is the cell address, \( D \) is the soil depth (m), \( \theta \) is the soil moisture content (m\(^3\) m\(^{-3}\), \( t \) is the time step (d), \( R \) is precipitation (m d\(^{-1}\)), \( E_o \) is the actual evapotranspiration (m d\(^{-1}\)), \( Q_{in,i} \) is the lateral inflow from upslope cells (m\(^3\) d\(^{-1}\)), \( Q_{out,i} \) is the lateral outflow to surrounding cells (m\(^3\) d\(^{-1}\)), \( P \) is the percolation from the soil profile to the bedrock reservoir (m d\(^{-1}\)), \( SE_{out,i} \) is the saturation excess runoff from the cell (m d\(^{-1}\)), \( IER_{in,i} \) is the infiltration excess runoff from imperious surfaces (m d\(^{-1}\)), and \( A \) is the cell area (m\(^2\)).

[16] For each pervious cell, any moisture storage in excess of saturation becomes runoff. On selected dates corresponding to measured runoff events, SMDR created output maps of runoff depths for each grid cell. A time step of one day was used. Presented below are the crucial component processes in the order they occur in the model.

![Figure 3](image-url)
For greater detail on SMDR formulation, one is referred to Zollweg et al. [1996], Frankenberger et al. [1999], and Gérard-Marchant et al. [2005].

3.1. Precipitation

[17] Precipitation and or snowmelt inputs are added to the moisture content of each cell at the beginning of each time step. Precipitation falls uniformly over the entire watershed and infiltrates the soil unless falling on an impervious surface, in which case it becomes runoff. Precipitation is assumed to occur as snow when the mean daily temperature is below 0°C. Snowmelt occurs when the mean daily temperature is above 0°C, and is estimated using the U.S. Army Corps of Engineers [1960] temperature index method. An adiabatic lapse rate (0.00636°C m⁻¹) is used to adjust temperatures based on the elevation, allowing higher, colder areas to retain snowpack longer than lower, warmer areas. Once precipitation or snowmelt has been added to the cell, soil moisture is redistributed to both layers in proportion to available storage capacity. If the storage capacity is satisfied the soil layers begin saturating from the restricting layer up.

3.2. Subsurface Lateral Flow

[18] Subsurface lateral flows are calculated for each grid cell at each time step and redistributed to a cardinal and a diagonal downslope neighboring cell, according to the D∞ algorithm, [Tarboton, 1997] and Darcy’s law:

$$Q_{out,j} = \sum_{i=1}^{n} w D_i K(\theta) \left( \frac{dh}{dl} \right)_i$$  \hspace{1cm} (2)

where w is the width of the cell (m), D is the depth (m), and $K(\theta)$ is the hydraulic conductivity of the cell (m d⁻¹), j is the number of soil layers modeled, n is the counter, and (dh/dl) is the hydraulic gradient (i.e., kinematic approximation in the direction of flow [Hillel, 1998]), which assumes that the restricting layer, and thus the water table, can be approximated by the cell slope.

[19] The hydraulic conductivity used to compute the lateral flow is modeled as a dual porosity/permeability model [Jarvis and Larson, 2001], with two flow regimes separated by a critical moisture content, noted $\theta_B$, above which water flows predominantly through macropores, and below which water flows through the soil matrix only. Below field capacity ($\theta_{FC}$) the hydraulic conductivity $K(\theta)$ tends toward zero:

$$K(\theta) = 0 \quad \text{for} \quad \theta < \theta_{FC}$$  \hspace{1cm} (3)

[20] Since SMDR assumes uniform moisture content in the soil profile above the restricting layer and in reality there is a perched water table and thus a gradient in moisture content, we have chosen a conductivity function to represent this. To find the drainage limit, $\theta_p$, where the soil is just saturated and macropores are conducting flow we assume that downward movement of water is small compared to the lateral flux. Consequently we set the downward hydraulic gradient equal to zero. In other words, when the soil is just saturated at the impermeable layer the matric potential at the soil surface is equal to the depth of the soil. The moisture content at this matric potential is the drainage limit ($\theta_B$), and can be calculated with a Brooks-Corey [Brooks and Corey, 1964] type relationship using the $\theta_{FC}$ from the soil survey [U.S. Department of Agriculture, 1965], at −30 kPa or equivalent to a 3 m height [Gérard-Marchant et al., 2005].

$$\theta_B = \theta_{FC} \left( \frac{3}{z} \right)^{1/4} \quad \text{for} \quad z \leq 3$$  \hspace{1cm} (4)

where $z$ is the depth to the restricting layer (m), and the Brooks Corey exponent can be estimated from the slope of the moisture retention curve (in log space). The conductivity function at $\theta < \theta_B$ represents unsaturated flow and is defined as

$$K_B = K_S \exp \left( -\alpha \frac{\theta_S - \theta}{\theta_S - \theta_{FC}} \right) \quad \text{for} \quad \theta_{FC} \leq \theta < \theta_B$$  \hspace{1cm} (5)

where $K_B$ is the hydraulic conductivity at $\theta_B$, $\theta_S$ is the saturated moisture content (m³ m⁻³) and $\alpha$ is a constant in hydraulic conductivity equal to 13, appropriate for many soils, including these NY soils [Steenhuis and van der Molen, 1986], and $K_S$ is the saturated conductivity. Since soil survey measures of $K_S$ are made on disturbed samples the contribution of macropores is neglected. When the soil moisture content is above $\theta_B$ a portion of the soil profile above the restricting layer is saturated and macropores are active in transport. Thus the conductivity is equal to $m K_S$ where $m$ is a factor (typical range of 2 to 10, decreasing with depth of the soil profile) introduced to correct soil survey transmissivity for preferential flow in macropores under saturated conditions [Boll et al., 1998]. Thus, at $\theta > \theta_B$ we assume that the soil above the water table is at $\theta_B$ and below $\theta_S$ the conductivity function is

$$K(\theta) = K_B + (mK_S - K_B) \frac{\theta - \theta_B}{\theta_S - \theta_B} \quad \text{for} \quad \theta_B < \theta < \theta_S$$  \hspace{1cm} (6)

3.3. Evapotranspiration

[21] After lateral flows have been computed actual evapotranspiration ($E_a$) from the root zone is calculated using the Thornthwaite-Mather procedure [Thornthwaite and Mather, 1955; Steenhuis and van der Molen, 1986]. Above $\theta_{FC}$ $E_a$ is equal to the potential evapotranspiration ($E_p$) times a vegetation coefficient that varies through the year [Jensen, 1973]. Below $\theta_{FC}$, $E_a$ changes linearly with $\theta$ from $E_p$ at $\theta_{FC}$ to zero at the wilting point ($\theta_{WP}$).

3.4. Deep Percolation and Base Flow

[22] Percolation out of the soil profile through the bedrock or fragipan (Figure 3) occurs only if the soil profile moisture content is above $\theta_{FC}$ (i.e., vertical percolation thought the layer occurs only as saturated flow assuming a unit hydraulic gradient). Flow stops when the soil reaches $\theta_{FC}$. The rate at which water percolates out of the soil profile is controlled by the effective hydraulic conductivity of the underlying restricting layer. Effective conductivities used were 3 mm d⁻¹ for bedrock and 0.2 mm d⁻¹ for fragipan, within the published ranges for New York soils.
[Maidment, 1993; Frankenberger et al., 1999]. Below $\theta_{FC}$ percolation is negligible.

[23] Percolation, $P$, (m d$^{-1}$) is added to a bedrock reservoir, $R_S$, (m d$^{-1}$) at each time step:

$$R_S(t) = (R_S(t-\Delta t) + P(t)) - BF(t-\Delta t)$$

(7)

where $BF$ is the modeled base flow (m d$^{-1}$).

[24] Base flow is modeled using a linear reservoir method as

$$BF(t) = R_S(t) \left[ 1 - \exp\left(-a\Delta t\right) \right]$$

(8)

where $a$ (d$^{-1}$) is the recession coefficient, a property of the aquifer, and can be calibrated from the base flow recession curve, and $\Delta t = 1$ for a daily time step. The recession coefficient, $a$, varies between the summer and winter periods. The coefficient was estimated by assuming that areas of the watershed underlain by the more permeable bedrock do not contribute to base flow during the summer (i.e., areas of the watershed underlain by bedrock are drier), while winter base flow, occurring under more saturated conditions, is a combination of flow from areas underlain by bedrock and areas underlain by fragipan.

3.5. Impervious Surfaces

[25] Previous versions of SMDR implemented in forested and agricultural watersheds did not include the effects of impervious surfaces. Since impervious surfaces are important in modeling an urbanized area, and the extent of the impervious surfaces varies at a scale finer than the model resolution (i.e.,<10 m), impervious surfaces (roads, roofs, driveways, etc.) were digitized on a 1 m basis from a spatially justified digital orthophotograph quarter quadrangle (DOQQ) of the watershed (available on the NYS GIS Web site at http://www.nysgis.state.ny.us/). Each 1 m $\times$ 1 m cell of the DOQQ was then qualified as either impervious or pervious, and given an imperviousness degree ($ID_1$) of 1 or 0, respectively. The imperviousness degree of each 10 m $\times$ 10 m cell, $ID_{10}$, was then defined by applying a 10 $\times$ 10 moving window to the 1 m DOQQ:

$$ID_{10} = \frac{1}{100} \sum_{i=1}^{100} ID_1$$

(9)

where $n$ is the cell counter. Each 10 m $\times$ 10 m cell is assigned an $ID_{10}$, ranging from zero to 1, based on the extent of impervious surfaces in the 100 cells of the region (Figure 2).

[26] The maps of imperviousness degree and base land use were then combined to create a single map incorporating all land uses as well as a spatially justified estimation of the impervious surfaces (Figure 2). On the basis of the modified land use map, infiltration of precipitation or snowmelt on each 10 m $\times$ 10 m cell was adjusted by a factor of $1 - ID_{10}$, with the rest becoming runoff (i.e., if a cell is 30% impervious, 70% of rainfall or snowmelt infiltrates, and 30% becomes runoff). However, since storm sewers do not service the majority of the urban area, runoff from impervious surfaces is allowed to re-infiltrate in the neighboring downslope cells. Infiltration excess runoff (IER) from the impervious surface is routed to a cardinal and diagonal downslope neighbor with the same D∞ algorithm as subsurface lateral flows. If a neighboring cell has storage available a fraction of the IER infiltrates the cell until it becomes saturated, and the rest is routed to the next adjacent downslope neighbor where it can infiltrate if storage is available. In this manner all IER from the impervious surfaces is allowed to re-infiltrate given adequate soil storage. Thus only saturation excess runoff is routed to the watershed outlet once generated.

3.6. Detention Pond and Control Structures

[27] Detention ponds for flood control are common features in urban watersheds [Behera et al., 1999], such as this watershed, but will confound model predictions if not considered in the modeling process. A large detention pond (not capture by the digital elevation model (DEM) located at the stream inflow to the urbanized area (Figure 1) and a large culvert/weir retention basin at the watershed outlet were taken into consideration. The detention pond has a capacity of nearly 14,000 m$^3$, determined from site measurements, while the culvert weir retention basin has a capacity of 42,500 m$^3$. Runoff was detained in the basins based on the storage capacity:

$$S(t) = S(t-\Delta t) - O(t-\Delta t) + SE(t)$$

(10)

where $S(t)$ is the storage (m$^3$) at time $t$, $S(t-\Delta t)$ is the storage during the previous time step, $O(t-\Delta t)$ is the outflow from the basin during the previous time step (m$^3$), and $SE(t)$ is the runoff contributing to the basin in the current time step (m$^3$). Outflow from the basin is calculated as

$$O(t) = S(t) - S_{MAX}$$

(11)

where $S_{MAX}$ is the basin storage capacity (m$^3$) determined from site measurements. Calculations were made on a daily time step.

[28] Thus streamflow is the sum of surface runoff, lateral flow from cells adjacent to the stream (interflow), and base flow (contribution from percolation to the bedrock reservoir). Runoff is assumed to reach the outlet within the time step unless it has reinfiltrated in the landscape. New additions to SMDR described here include the base flow determination, the effects of impervious surfaces, the detention pond, and culverts on the mainstream channel.

4. Model Input Data

4.1. Weather

[29] During the growing season precipitation was measured in the watershed at 10-min intervals with two tipping bucket rain gauges (24 cm diameter). Temperature and winter snowfall and precipitation were measured hourly 3 km southeast of the watershed outlet. During the growing season (May–October) pan ET was estimated with a class A pan at a location 5.5 km south east of the watershed outlet. A pan factor of 0.8 was used to determine PET. Non-growing season ET was calculated using the Penman-Monteith method. All weather data was converted to daily values.
4.2. Base Maps

SMDR incorporates three primary maps; a DEM, soil, and land use map with associated descriptive data. All base maps were reprojected into the North American Datum 1927.

A 10 m resolution DEM was downloaded from the Cornell University Geospatial Information Repository (CUGIR) for the Ithaca East quadrangle. Construction in the watershed since the DEM was compiled changed elevation and altered flow near the watershed outlet. To remedy this, a 5 m grid was set up and elevation data was measured using a global positioning system with Wide Area Augmentation System (0.85 m accuracy) and the DEM modified correspondingly. Road drainage ditches not captured by the original DEM were digitized and the DEM modified accordingly.

Digital soil maps of Tompkins County in vector format were obtained from the State Soil Geographic (STASTGO) database (USDA-NRCS). These maps are digitized duplications of the original soil survey maps published in 1:24000 scales. The corresponding companion soil attribute tables [U.S. Department of Agriculture, 1965] were used to assign soil properties. Seven soil properties are required as input for SMDR, including porosity (η), bulk density (ρb), depth (z), field capacity (θFC), wilting point (θWP), saturated hydraulic conductivity (Ks), and percentage of rock fragments (RFC) (Table 1). When ranges of values were given the simple arithmetic mean was used.

4.3. Model Parameterization

Most model parameters were taken directly from the sources described above as averages of ranges given in tables with no calibration. Other parameter values were based on previous application of the model in the Catskill Mountains of New York. The drainage limit, θB, was based on the θFC in the soil survey and introduced to account for differences between laboratory derived permeability measurements and observations in undisturbed field soils [Boll et al., 1998]. The horizontal conductivity multiplier, m, used to convert Soil Survey laboratory derived permeability measurements to field values, was assumed to decrease exponentially with soil layer thickness from 10 for the shallowest soils to 2 for the deepest [Gérard-Marchant et al., 2005]. The effective conductivities of the bedrock and fragipan were selected a priori as 3 mm d⁻¹ and 0.2 mm d⁻¹ respectively, which are within the published ranges [Frankenberger et al., 1999; Maidment, 1993]. Only very limited calibration is required for parameters with no estimate available. The reservoir fraction, a, was calibrated to account for differences in the base flow regimes during the growing and dormant seasons using 2003–2004 stream discharge data. The calibrated a = 0.45 for summer (May–October) and 0.25 for winter (November–April) periods.
4.4. Model Assessment of Urban Impacts on Hydrology

[35] With the validated model (full urban model) we subsequently investigated the impacts of urbanization on watershed hydrologic response for a 2-year period by simulating the watershed without the effect of the detention pond (using the full urban model) and without development (using the predevelopment model). To model the watershed predevelopment the imperviousness degree (ID) was set to 0 and the land use/land cover (LULC) to forest in the urbanized area. To remove the effect of the detention pond from the full urban model, all streamflow was routed directly to the outlet without delay. No other input parameters were altered.

5. Results and Discussion

5.1. Validation

[36] Measured and modeled streamflow was compared at the watershed outlet for the full urban model. In addition to integrated watershed response, we investigated the distributed responses because correctly simulating processes internal to the watershed, such as runoff generation, is as critical to assessing the impact of urbanization as stream discharge. Responses were evaluated on event, seasonal, and yearly basis for a 2-year period.

5.2. Integrated Watershed Response

[37] Runoff from saturated areas and subsurface flow from the watershed were summed at the watershed outlet to predict streamflow. The graphical comparison of the modeled and measured streamflow over the course of the study (April 2003 to April 2005) for the full urban model is shown in Figure 4b. The model was able to capture the dynamics of most events well ($E = 0.79$, $r^2 = 0.81$) (Table 2 and Figure 4b). Both base flow and stormflow were correctly predicted with a slight over prediction of peak flows and a slight under prediction of low flows (Table 2), however, all statistical evaluation criterion indicted the model predicted well. Seasonal model statistics for the full urban model were better for summer than winter periods (Table 2), a direct result of the difficulty in predicting the
Table 2. Summary of Seasonal and Yearly Measured Streamflow Compared With the Full Urban Model, the Full Urban Model With No Detention Pond, and the Predevelopment Model at the Watershed Outlet

<table>
<thead>
<tr>
<th>Period</th>
<th>Minimum, mm d⁻¹</th>
<th>Mean, mm d⁻¹</th>
<th>Maximum, mm d⁻¹</th>
<th>Eᵇ</th>
<th>r²ᶜ</th>
<th>MAE⁺, mm d⁻¹</th>
<th>MCE⁺, mm d⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Measured</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Summer 2003</td>
<td>0.32</td>
<td>1.63</td>
<td>13.55</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Winter 2003–2004</td>
<td>0.46</td>
<td>1.94</td>
<td>10.96</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Summer 2004</td>
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<td>1.43</td>
<td>18.28</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Winter 2004–2005</td>
<td>0.45</td>
<td>1.87</td>
<td>23.87</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Summer</td>
<td>0.32</td>
<td>1.53</td>
<td>18.28</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>23.87</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Entire period</td>
<td>0.32</td>
<td>1.70</td>
<td>23.87</td>
<td></td>
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<td></td>
</tr>
<tr>
<td><strong>Full Urban Model</strong></td>
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</tr>
<tr>
<td>Summer 2003</td>
<td>0.35</td>
<td>1.58</td>
<td>16.54</td>
<td>0.81</td>
<td>0.83</td>
<td>0.32</td>
<td>0.55</td>
</tr>
<tr>
<td>Winter 2003–2004</td>
<td>0.41</td>
<td>1.99</td>
<td>11.54</td>
<td>0.71</td>
<td>0.74</td>
<td>0.21</td>
<td>0.90</td>
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<tr>
<td>Summer 2004</td>
<td>0.33</td>
<td>1.61</td>
<td>18.77</td>
<td>0.85</td>
<td>0.88</td>
<td>0.49</td>
<td>0.88</td>
</tr>
<tr>
<td>Winter 2004–2005</td>
<td>0.31</td>
<td>1.78</td>
<td>22.31</td>
<td>0.75</td>
<td>0.78</td>
<td>0.32</td>
<td>0.85</td>
</tr>
<tr>
<td>Summer</td>
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<td>1.59</td>
<td>18.77</td>
<td>0.83</td>
<td>0.86</td>
<td>0.32</td>
<td>0.71</td>
</tr>
<tr>
<td>Winter</td>
<td>0.31</td>
<td>1.89</td>
<td>22.31</td>
<td>0.74</td>
<td>0.77</td>
<td>0.27</td>
<td>0.88</td>
</tr>
<tr>
<td>Entire period</td>
<td>0.31</td>
<td>1.73</td>
<td>22.31</td>
<td>0.79</td>
<td>0.81</td>
<td>0.36</td>
<td>0.80</td>
</tr>
<tr>
<td><strong>Full Urban Model No Detention Pond</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Summer 2003</td>
<td>0.35</td>
<td>1.88</td>
<td>18.14</td>
<td>0.50</td>
<td>0.71</td>
<td>0.38</td>
<td>0.90</td>
</tr>
<tr>
<td>Winter 2003–2004</td>
<td>0.41</td>
<td>1.75</td>
<td>15.68</td>
<td>0.63</td>
<td>0.72</td>
<td>0.42</td>
<td>0.98</td>
</tr>
<tr>
<td>Summer 2004</td>
<td>0.33</td>
<td>1.72</td>
<td>20.03</td>
<td>0.62</td>
<td>0.74</td>
<td>0.47</td>
<td>0.87</td>
</tr>
<tr>
<td>Winter 2004–2005</td>
<td>0.31</td>
<td>1.72</td>
<td>25.82</td>
<td>0.51</td>
<td>0.60</td>
<td>0.32</td>
<td>0.97</td>
</tr>
<tr>
<td>Summer</td>
<td>0.33</td>
<td>1.80</td>
<td>20.03</td>
<td>0.56</td>
<td>0.71</td>
<td>0.32</td>
<td>0.89</td>
</tr>
<tr>
<td>Winter</td>
<td>0.31</td>
<td>1.73</td>
<td>25.82</td>
<td>0.56</td>
<td>0.66</td>
<td>0.37</td>
<td>0.98</td>
</tr>
<tr>
<td>Entire period</td>
<td>0.31</td>
<td>1.77</td>
<td>25.82</td>
<td>0.56</td>
<td>0.68</td>
<td>0.40</td>
<td>0.95</td>
</tr>
<tr>
<td><strong>Predevelopment Model</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Summer 2003</td>
<td>0.40</td>
<td>1.57</td>
<td>11.72</td>
<td>0.70</td>
<td>0.57</td>
<td>0.51</td>
<td>0.83</td>
</tr>
<tr>
<td>Winter 2003–2004</td>
<td>0.57</td>
<td>1.60</td>
<td>10.50</td>
<td>0.71</td>
<td>0.56</td>
<td>0.54</td>
<td>0.97</td>
</tr>
<tr>
<td>Summer 2004</td>
<td>0.46</td>
<td>1.37</td>
<td>12.54</td>
<td>0.76</td>
<td>0.61</td>
<td>0.52</td>
<td>0.89</td>
</tr>
<tr>
<td>Winter 2004–2005</td>
<td>0.52</td>
<td>1.60</td>
<td>12.54</td>
<td>0.72</td>
<td>0.70</td>
<td>0.52</td>
<td>1.00</td>
</tr>
<tr>
<td>Summer</td>
<td>0.52</td>
<td>1.58</td>
<td>12.54</td>
<td>0.60</td>
<td>0.54</td>
<td>0.55</td>
<td>1.00</td>
</tr>
<tr>
<td>Winter</td>
<td>0.52</td>
<td>1.64</td>
<td>12.54</td>
<td>0.64</td>
<td>0.61</td>
<td>0.54</td>
<td>0.92</td>
</tr>
<tr>
<td>Entire period</td>
<td>0.46</td>
<td>1.65</td>
<td>12.54</td>
<td>0.64</td>
<td>0.61</td>
<td>0.54</td>
<td>0.92</td>
</tr>
</tbody>
</table>

*aSummer is May–October. Winter is November–April.*

*bNash-Sutcliffe efficiency comparison with measured streamflow.*

*cCoefficient of determination comparison with measured streamflow.*

*dMean absolute error comparison with measured streamflow.*

*eMean cumulative error comparison with measured streamflow.*

Table 3. Summary Statistics for Measured and Modeled Degree of Soil Saturation From 13 Events Using the Full Urban Model Taken From the Landscape Sampling Locations

<table>
<thead>
<tr>
<th>Event</th>
<th>Minimum, cm³ cm⁻³</th>
<th>Mean, cm³ cm⁻³</th>
<th>Maximum, cm³ cm⁻³</th>
<th>Eᵇ</th>
<th>r²ᶜ</th>
<th>SE⁺, cm³ cm⁻³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jul 2003ᵇ</td>
<td>0.23</td>
<td>0.59</td>
<td>0.89</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mar 2004ᵇ</td>
<td>0.54</td>
<td>0.78</td>
<td>0.91</td>
<td>0.55</td>
<td>0.84</td>
<td>0.97</td>
</tr>
<tr>
<td>Apr 2004ᵇ</td>
<td>0.42</td>
<td>0.64</td>
<td>0.81</td>
<td>0.51</td>
<td>0.69</td>
<td>0.88</td>
</tr>
<tr>
<td>Aug 2004ᵇ</td>
<td>0.33</td>
<td>0.61</td>
<td>0.92</td>
<td>0.34</td>
<td>0.62</td>
<td>0.97</td>
</tr>
<tr>
<td>Overall</td>
<td>0.23</td>
<td>0.60</td>
<td>0.92</td>
<td>0.25</td>
<td>0.62</td>
<td>0.97</td>
</tr>
</tbody>
</table>

*aSee Figure 1 for sampling locations. Events were combined for statistical comparison if sampling dates were consecutive.*

*bNash-Sutcliffe efficiency.*

*cCoefficient of determination.*

*dStandard error of predicted to measured degree of saturation.*

*eEvent consisted of six sampling dates.*

*fEvent consisted of one sampling date.*

*gEvent consisted of one sampling date.*

*hEvent consisted of five sampling dates.*
5.3. Distributed Watershed Response

Determining the spatial distribution of hydrologic predictions is essential in identifying VSAs, and the optimum locations for management practices enacted to enhance water quality in urbanized watersheds. Areas with a high saturation degree are much more likely to produce runoff than areas with a low saturation degree, thus these areas are important to identify. In this section we first assess the distributed soil moisture response and then the runoff response in the watershed.

5.4. Soil Moisture

Normalized soil moisture levels (i.e., saturation degree) from 12 locations in the watershed (Figure 1) were used to assess the model’s ability to locate saturated areas in the landscape. A summary of the saturation degree predictions and statistical comparison to measured values are shown for 13 dates in Table 3 and two time series in Figures 5 and 6.

- The 18–24 July 2003 sampling dates (Figure 5) were generally well modeled (Table 3) with a slight overestimation, particularly near the watershed outlet under saturated conditions, a result of sampling after the storm when interflow drainage had lowered the water content of the shallowest soils (Table 1). The standard error between modeled and measured values (0.008 cm$^3$ cm$^{-3}$) is less than the within-cell sampling error and variability (0.026 cm$^3$ cm$^{-3}$) indicating the predictions are accurate.

- The effect of an intense thunder storm causing IER on some soils can be seen during the 29 August 2004 event (Table 3 and Figure 6), where measured and modeled saturation degrees were not well correlated due to the model assuming all the precipitation infiltrated the soil when, in fact, some ran off directly. On the following days, 31 August and 1 September 2004, where rainfall intensities were low, the saturation degree was well captured (Figure 6). The March 2004 event was generally well captured for a winter

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Figure 5. Measured and modeled saturation degree from landscape sampling locations in the watershed and corresponding SMDR output map of the saturation degree using the full urban model for 18 July 2003, 20 July 2003, 22 July 2003, and 24 July 2003, during which 148 mm of precipitation fell. The sampling locations appearing as numbers on the first map and as dots in the subsequent maps correspond to the number on the x axis of the adjacent graph.
event \( (E = 0.49, r^2 = 0.94, \text{ standard error} = 0.012 \text{ cm}^3 \text{ cm}^{-3}) \), probably due to uniformly saturated soils in the majority of the watershed. There was very little change in the moisture content over the three months from December to March 2004. The April 2004 event was accurately modeled as well \( (E = 0.51, r^2 = 0.91, \text{ and standard error} = 0.009 \text{ cm}^3 \text{ cm}^{-3}) \), again due to a narrow range of high soil moisture levels in the watershed.

In general, the saturation degree was accurately modeled with the majority of modeled points falling inside the acceptable error estimates established for watershed soil moisture levels. Figure 7 shows plot of the measured and modeled saturation degree for all events modeled. The modeled saturation degree shows a small bias at high and low moisture contents. The highest saturation degrees occurred consistently in the area of the watershed with the shallowest soil, most variable slopes, and highest impervious degree (landscape sampling locations 1–6 in Figures 5 and 6), a result of the low storage capacity in the shallow soils and reinfiltration of runoff produced on impervious surfaces.

5.5. Landscape Runoff

At nine of the 12 landscape sampling locations in the watershed (Figure 1) runoff was measured and used to assess model predictions of runoff. SMDR predicted distributed runoff losses well for these landscape sampling locations during both the growing season and winter events (Table 4). Similar to saturation degree and streamflow predictions, summer events were slightly more accurately modeled than winter events (Table 4). A plot of measured and modeled runoff events fall about the 1:1 with some scatter, but no systematic deviation (Figure 8). Three distinct periods in July 2003, August 2004 and March–April 2005 and are used below to illustrate the model’s performance.

![Figure 6](image1.png)

*Figure 6.* Measured and modeled saturation degree from landscape sampling locations in the watershed and corresponding SMDR output map of the saturation degree using the full urban model for 29 August 2004, 31 August 2004, and 1 September 2004 during which 94 mm of precipitation fell. The sampling locations appearing as numbers on the first map and as dots in the subsequent maps correspond to the number on the x axis of the adjacent graph.

![Figure 7](image2.png)

*Figure 7.* Measured and modeled saturation degree \( (\theta) \) in samples taken from 12 locations (Figure 1) and 13 sampling dates in the watershed using the full urban model.
Table 4. Summary Statistics for Measured and Modeled Runoff Losses Using the Full Urban Model From 15 Runoff Events Taken From the Landscape Sampling Locations⁸

<table>
<thead>
<tr>
<th>Event</th>
<th>Measured</th>
<th>Modeled</th>
<th>E⁹</th>
<th>r²d⁹</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean, mm</td>
<td>Maximum, mm</td>
<td>Mean, mm</td>
<td>Maximum, mm</td>
</tr>
<tr>
<td>Jul 2003⁹</td>
<td>2.0</td>
<td>24.3</td>
<td>1.9</td>
<td>23.3</td>
</tr>
<tr>
<td>Aug 2004¹</td>
<td>4.4</td>
<td>33.6</td>
<td>4.1</td>
<td>31.7</td>
</tr>
<tr>
<td>Mar–Apr 2005⁶</td>
<td>9.8</td>
<td>40.5</td>
<td>10.1</td>
<td>39.8</td>
</tr>
<tr>
<td>Overall</td>
<td>4.2</td>
<td>40.5</td>
<td>4.2</td>
<td>39.8</td>
</tr>
</tbody>
</table>

⁸See Figure 1 for sampling locations. Events were combined for statistical comparison if sampling dates were consecutive.

⁹Minimum values for both measured and modeled runoff losses were zero for all events.

¹⁰Coefficient of determination.

During the event in July 2003 (Figure 9) a 17 mm, low-intensity storm on 18 July was followed by a higher-intensity 22 mm storm on 20 July and a large 109 mm storm with low intensity on 22–24 July. The watershed moisture contents increased correspondingly during this period (Figure 6). Initially, on 18 July, the moisture content in the watershed ranged from dry (0.30 cm⁻³ cm⁻³), mainly in the upper areas, to wet (0.65 cm⁻³ cm⁻³), in the lower areas; as the storm progressed, the whole watershed became wetter especially, the upper portion of the watershed (Figure 5). Measured and modeled runoff for the 18 and 20 July portions of the event were highest from the more urbanized area of the watershed on the shallowest soils (Figure 9). On 18 July, landscape sampling location 1 had the highest runoff (2 mm or ~10% of the 17 mm rainfall) (Figure 10). On 20 July both plot 1 and plot 5, located in the urbanized area, had the highest runoff losses (~10% of the 22 mm rainfall). The model under predicted runoff from the 20 July portion of the event (Figure 9), perhaps because small portions of the landscape produced infiltration excess runoff (maximum intensities were 43 mm h⁻¹). The 109 mm rain on 22–24 July, all with intensities <10 mm h⁻¹ produced predominately saturation excess runoff, the distribution of which was captured very well (Figure 9), particularly the losses from large variable source area in the urban area (landscape sampling locations 1–5). The upper area of the watershed started to contribute runoff on 22 July (Figure 9), as it saturated (Figures 6 and 10).

The final and largest runoff event of the study occurred in March and April 2005 consisting of snowmelt in addition to 111 mm of precipitation (Figure 11). The initial portion of the event (29 March to 1 April 2005, Figure 11), consisted almost entirely of snowmelt, and was under predicted on all but one landscape sampling location (landscape 8). As noted by other researchers [Frankenberger et al., 1999; Mehta et al., 2004; Gérard-Marchant et al., 2005] SMDR is most prone to errors when the temperature fluctuates around freezing, which occurred during this period. On the following days (2–4 April 2005), where there was minimal snowmelt and the majority of the precipitation occurred, the runoff losses were generally well captured (Figure 11). Since the soils of the upper area of the watershed were saturated considerable runoff was generated in the area during the event (Figure 11).

For all events, the majority of the runoff is modeled as occurring in the lower urban area of the watershed, consistent with measured runoff losses, while the upper undeveloped area displayed low measured and modeled runoff. These runoff losses are coincident with the higher soil moisture levels, shallower soils, and higher fraction of impervious surfaces in the more urbanized area of the

Figure 8. Measured and modeled runoff depth taken from nine locations (Figure 1) and 15 sampling dates in the watershed using the full urban model.
watershed. In all cases, except snowmelt, the variable source area in the urban area of the watershed produced the majority of the runoff (landscape sampling locations 1–5 in Figures 9, 10, and 11).

5.6. Model Assessment of Urban Impacts on Hydrology

[47] Having successfully validated SMDR for urban (above) and rural watersheds [Frankenberger et al., 1999; Johnson et al., 2003; Mehta et al., 2004; Gérard-Marchant et al., 2005], we used the model to assess the specific impact of urbanization on watershed hydrology.

[48] Predevelopment model results were compared to full urban model results and to the measured data (e.g., Figures 4b and 4c, respectively). Interestingly, the Nash-Sutcliffe efficiency statistics (Table 2) and visual inspection of the streamflow hydrograph for the predevelopment model (Figure 4c) suggested that the model captured overall stream response dynamics well, however, the linear correlation between modeled and observed runoff (Figure 4c, inset) revealed a systematic bias in the model predictions. The predevelopment model predicted less stormflow (2.6 versus 3.2 mm d\(^{-1}\)) and increased lateral subsurface flow (2.2 versus 1.8 mm d\(^{-1}\)) compared to the full urban model.

Figure 9. Measured and modeled runoff for five runoff events in July 2003 using the full urban model. Event included intense thunderstorm precipitation (20 July 2003) and low-intensity sustained precipitation (total precipitation was 148 mm). SMDR output maps of the distributed runoff generation in the watershed are for the July 2003 event. The sampling locations appearing as numbers on the first map and as dots in the subsequent maps correspond to the number on the x axis of the adjacent graph.
This was a direct result of impervious surfaces, which in the full urban model prevented the infiltration and shed the storm water to the surrounding landscape, increasing the soil moisture adjacent to the impervious surfaces. This, in turn, increased runoff from the land near impervious surfaces. This pattern is clearly visible in Figures 5 and 6, where the soil was unsaturated beneath the impervious surfaces and substantially more saturated in the adjacent soils. Intuitively, these areas with higher soil moisture levels are especially prone to runoff losses as seen with the full urban model (Figures 9 and 10). In the predevelopment model, soil was less saturated and there was no simulated runoff on these locations (Figure 12) because the lack of impervious surfaces allowed infiltration. Additionally, because removal of impervious surfaces allowed infiltration across the whole watershed in the predevelopment model, summer evapotranspiration was generally higher than in the full urban model, although these differences were smaller than those seen with respect to overland runoff and subsurface flows.

Removing the effect of the detention pond from the full urban model increased simulated peak flows by an average of 16% (Figure 4d), although the statistical analysis indicated that the modeled streamflows using the full urban model with no detention pond are correlated fairly well with the measured streamflows ($r^2 = 0.60–0.74$) the poor efficiency and visual inspection of Figure 4d show that the model is severely over predicting stormflows (Table 2). Removing the detention basin worsened simulated streamflows particularly for winter snowmelt events (Table 2) when the upper part of the watershed contributed substantial runoff. During the winter, the soil in the upper watershed was usually near saturation because rainfall exceeded evaporation and there was little interflow due to the shallow slopes (2–5%). Thus, during snowmelt the soils in the upper watershed produced considerable saturation excess runoff (spring in Figure 4b and Figure 11). During the growing season when the evaporation exceeded rainfall, the soil is drier in the upper area (as can be seen in Figures 5 and 6) and runoff was minimal (Figures 9 and 10) and thus the absence of the detention basin, which is located above the urban area, had little impact on the model results. The summer events were dominated by the runoff from the urban area, i.e., the impact of the impervious surfaces on soil moisture and subsequent runoff.

[49] During large storm events on initially unsaturated soils the differences in soil moisture levels were as high as 50% between different areas of the watershed. For instance, the VSA in the urbanized area that frequently saturated during rainfall had a lower soil moisture level during dry periods than other areas of the watershed (18 July 2003 in Figure 5 and 29 August 2004 in Figure 6). This is due to the influence of steeper slopes on the lateral drainage of the water from the shallower soils [Hessel et al., 2003], and the
Figure 11. Measured and modeled runoff for six runoff events in March and April 2005 using the full urban model. Events consisted of snowmelt and precipitation (111 mm). SMDR output maps of the distributed runoff generation in the watershed for the March and April 2005 event. The sampling locations appearing as numbers on the first map and as dots in the subsequent maps correspond to the number on the x axis of the adjacent graph.

Figure 12. Soil moisture and runoff losses for 23 June 2003 for the predevelopment model.
fact that there is less water infiltrating the soil, due to the impervious surfaces. These soil moisture dynamics are important to consider when characterizing the runoff distribution from urban areas. Consider Figures 5 and 6 for soil moisture and Figures 9 and 10 for runoff losses. At the beginning of each storm the soil was initially unsaturated, and able to assimilate substantial quantities of precipitation. As the storm progressed the soil became progressively wetter, and runoff began to occur. As can be seen in Figures 5 and 6 some places in the urban area were saturated and, correspondingly, produced the largest runoff losses (Figures 9 and 10). The majority of the runoff from the June 2003 and August 2004 events was produced after the soil saturated, very little was produced prior to saturated conditions despite 60–80 mm of precipitation (Figures 9 and 10). The shallow soils in the urban area were unable to assimilate this excess water and, as a result, the soils remained saturated and substantial runoff was produced. However, during the same period the deeper soils in the upper watershed were able to infiltrate more of the precipitation, and produced very little runoff (Figures 5, 6, 9, and 10). The generally higher soil moisture levels and runoff losses in the urban area can be attributed to the shallow soils and the process of the impervious surfaces shedding water to adjacent areas.

In this watershed all of the IER from impervious surfaces could potentially reinfiltrate in the surrounding landscape unless there was insufficient water storage capacity in the surrounding soil. This can be seen in Figures 9 and 10 when, during the first day of the June 2003 and August 2004 events, runoff was only created on impervious surfaces and infiltrated in the surrounding soil, as indicated by the lack of stream response (Figure 4b). As the events progressed and the soils continued to wet they were able to assimilate less and less of the IER from the impervious surfaces, and thus caused a large and rapid increase in streamflow (Figure 4b). By comparison, the predevelopment model predicted lower peak stream flows, and shallower rising and receding hydrograph limbs (Figure 4c), consistent with the increased infiltration when impervious surfaces are not present. During snowmelt or events on saturated soils very little of the IER from impervious surfaces was able to infiltrate because of lack of available soil storage (Figure 11) causing high runoff losses from most of the watershed. Since the vast majority of precipitation falling on impervious surfaces is expected to end up as runoff, and this watershed has a very limited storm water management system, the bulk of the runoff is directed to the surrounding soil where it may or may not infiltrate the soil. When it cannot, runoff losses from the urban area increase dramatically. In fact during some events, particularly when the soil was saturated in the spring, nearly all the IER from impervious surfaces and the precipitation falling on saturated soils ended up as runoff.

6. Model Improvements

While streamflow predictions were generally very good, improvements could be made, particularly during the winter. Visual inspection of Figure 4b, shows that full urban model captures rain events well during the winter (December 2003 and January 2005), but in February 2004 and 2005 snowmelt events were underestimated by 20–30%. Previously the shortcoming during the winter were noted when temperatures oscillated between freezing and non-freezing [Frankenberger et al., 1999; Mehta et al., 2004]. In urban areas snowmelt predictions are complicated further by salt applications, which cause melt on impervious surfaces and are not reflected in temperature indices. Predictions are further complicated by the effect of snow plowing on melt dynamics. For instance, snow piled by plowing has a greater density and a smaller surface area, and thus lower melting rates than undisturbed snow, which may reduce daily melt volumes and extend the melt period [Valeo and Ho, 2004]. The reduced albedo associated with urban areas [Taha, 1997] may also increase the actual ground surface temperature causing increased snowmelt. Additionally, in urban areas snowmelt appears to be dominated by net radiation fluxes, as opposed to sensible heat flux, turbulent exchange, and heat exchange at the soil-snow interface, as in natural systems [Sundin et al., 1999; Valeo and Ho, 2004]. More representative watershed temperatures and the inclusion of solar radiation may improve snowmelt estimates. Including these processes in the model is possible but calibration would be required as some of these input parameters are not easily available.

SMDR does not correctly capture IER from dry soil during periods of very high rainfall intensity (the model assumes all rain infiltrates until the soil water capacity is filled), for example the thunderstorms on 20 June 2003 (Figures 4b, 5, and 9). To loosely correct for this problem, SMDR can be modified to consider short-duration rainfall amounts, e.g., hourly rainfall, and calculate runoff from areas where the permeability is lower than the rainfall intensity. Subsequently, the rest of the storm, i.e., the portion not contributing to IER could be handled using SMDR current algorithms. Of course, this introduces difficulties for watersheds for which short-duration rain data are unavailable. This effect is lessened when thunderstorms occur on saturated soils where the rainfall intensity is not as important, and is generally well captured by the model in this application (September 2003, Figure 4b). It should be emphasized that SMDR is designed specifically for well-vegetated watersheds with high-permeability soils, slopes greater than 2%, and a shallow restricting layer.

The reinfiltration of runoff from impervious surfaces may not be applicable in urban watersheds that are storm sewered or otherwise convey storm water from impervious areas via subsurface pipes. Inclusion of these structures should be straightforward with this model conceptualization for watersheds where the location of storm water systems is georeferenced.

7. Conclusions and Implications

In this paper we developed and tested an urban version of SMDR, a fully distributed, process-based model in an urban watershed. This full urban model agreed well with field measurements of watershed discharge, distributed soil moisture, and distributed runoff generation. Distributed simulations of soil moisture and runoff showed that water accumulated at the bottom of hillslopes (and similar at slope breaks) and adjacent to large expanses of impervious surfaces.

Both the full urban model results and the field measurements indicated that the urban portions of the
watershed produced the greatest runoff, which may increase contaminant losses from these areas. Runoff losses from landscapes in the lower portion of the watershed (i.e., sampling locations 1–5) were consistently higher than losses from other areas of the watershed. That this area coincides with the more heavily developed part of the watershed is not likely coincidental, because impervious surfaces shed water to the surrounding landscape, resulting in a hyperpropensity for saturation and saturation excess runoff in the adjacent soils. This has serious implications for the impact of urban areas on surface water quality, in the sense that there is little chance for soil physical, biological, chemical processes to remediate contaminants transported via runoff when the majority of precipitation becomes runoff.

[57] These results have shown that the full urban model version of SMDR can provide valuable insight into distributed urban watershed hydrology. The implications for watershed managers are clear, there are certain areas in urban watersheds that produce most of the runoff and are thus especially susceptible to pollutant loss. For instance the VSA producing large runoff losses in the urban area of the watershed in this study coincides with many fertilized home lawns, which may be a source of excessive nutrients to the stream. The full urban model version of SMDR can help predict runoff producing areas on a fully distributed basis and this information can help watershed managers target storm water management strategies at parts of the landscape particularly prone to generating runoff including assessing the suitability of certain activities in high-runoff-generating areas of the watershed.

**Notation**

- $a$: fraction of bedrock reservoir as base flow (d$^{-1}$).
- $A$: grid cell area (m$^2$).
- $BF$: base flow (m$^3$ d$^{-1}$).
- $D$: soil depth (m).
- $dH/dt$: hydraulic gradient (m m$^{-1}$).
- $E_a$: actual evapotranspiration (m d$^{-1}$).
- $E_p$: potential evapotranspiration (m d$^{-1}$).
- $ID_{10}$: imperviousness degree of each10 m cell.
- $ID_1$: imperviousness degree of each 1 m cell.
- $IER_{in,l}$: infiltration excess runoff from impervious surfaces (m d$^{-1}$).
- $K$: hydraulic conductivity (m d$^{-1}$).
- $K_B$: hydraulic conductivity at $\theta_B$ (m d$^{-1}$).
- $K_s$: saturated hydraulic conductivity (m d$^{-1}$).
- $K(C)$: unsaturated hydraulic conductivity (m d$^{-1}$).
- $m$: transmissivity factor.
- $n$: number of soil layers.
- $O$: outflow from the detention pond (m$^3$ d$^{-1}$).
- $P$: percolation (m d$^{-1}$).
- $Q_{in,l}$: lateral inflow from surrounding cell (m$^3$ d$^{-1}$).
- $Q_{out,l}$: lateral outflow to surrounding cell (m$^3$ d$^{-1}$).
- $R$: rainfall (m d$^{-1}$).
- $R_B$: bedrock reservoir storage (m$^3$ d$^{-1}$).
- $SE_{out,l}$: saturation excess runoff (m d$^{-1}$).
- $S_{MAX}$: detention pond storage capacity (m$^3$).
- $S$: detention pond storage (m$^3$).
- $t$: time step (d$^{-1}$).
- $w$: grid cell width (m).

**Acknowledgment.** Thanks to the USDA for partial funding support for this research.

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