PROGRESS TOWARD DEVELOPMENT OF A GIS BASED WATER QUALITY MANAGEMENT TOOL FOR SMALL RURAL WATERSHEDS: MODIFICATION AND APPLICATION OF A DISTRIBUTED MODEL.

by

J. Boll  
Assistant Professor

C.O. Stockle  
Professor

S.K. Young  
Graduate Research Assistant

E.S. Brooks  
Graduate Research Assistant

J.E. Hammel  
Professor

C.R. Campbell  
Research Assistant

P.A. McDaniel  
Associate Professor

Department of Biological and Agricultural Engineering  
University of Idaho

Department of Biological Systems Engineering  
Washington State University

Department of Plant, Soils and Entomological Sciences  
University of Idaho

Moscow, ID  83844-0904

Pullman, WA 99164-6120

Moscow, ID  83844-2339

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Summary:

This paper reports on the development of the hydrology component of a water quality management tool for the Palouse Region of the Pacific Northwest. We apply and modify the Soil Moisture Routing (SMR) model which was originally developed at Cornell University. Perched water table data from a small catchment near Troy, ID, serves as a basis for validation of the original version of SMR and subsequent modifications. Changes in the SMR model presented are based on field-derived values of the saturated hydraulic conductivity and the presence of three soil horizons. The importance of the saturated hydraulic conductivity in accuracy of model predictions is shown. A major finding of our study is that the SMR model is well-suited for use in the Palouse Region provided that saturated hydraulic conductivities determined in the laboratory are adjusted to represent field-derived conductivities. The best predictions were obtained when the saturated hydraulic conductivity was decreased approximately exponentially with depth.

Keywords:

hillslope hydrology, modeling, GIS, water balance, fragipan, water quality
INTRODUCTION

Water quality abatement through restoration and management of watersheds is receiving tremendous attention in the United States at all levels of government and in local communities. Since contributions of most point sources have been reduced to acceptable levels, the main emphasis is on the control of nonpoint sources originating from urban, forest, agricultural, and recreational lands. Agriculture is identified as a significant contributor of non-point source pollution. Current efforts by EPA to address non-point source pollution are formulated in sections 208, 303(d) and 319 of the Clean Water Act including the formulation of Total Maximum Daily Loads. In Idaho, 962 lakes and stream segments have been placed on the 303(d) list. In Washington State, this number approaches 700.

Abatement of non-point source pollution is most manageable and most cost-effective when problems are identified and prioritized at the level of small watersheds. To assist in objective decision-making at the local scale in Idaho and Washington State, scientists of the University of Idaho and Washington State University are developing a GIS-based problem-solving tool for small rural watersheds in the Palouse Region of the Pacific Northwest. This tool will combine an existing climate generator (ClimGen) and crop growth model (CropSyst; Stockle et al., 1994), with newly developed models for subsurface and surface hydrology, and sediment and pollutant transport. A geographic information system (GIS) using state-of-the-art user interfaces will provide the linkage between these models. This paper reports on the development of the hydrology component of this problem-solving tool.

Hydrologic conditions in the Palouse Region are dominated by low intensity rainfall, snowmelt, and transient frozen soils. In addition, seasonal perched water tables are typical in soils of the Palouse Region. Perching occurs due to very slowly permeable subsurface horizons (i.e., fragipans) that severely restrict vertical water movement. Perched water tables are potential environmental hazards because of their proximity to surface-applied agrichemicals and their ability to promote rapid, lateral water flow.

When soil surface temperatures are above 0 °C, conditions of low intensity rainfall, snowmelt and perched water tables generate saturation excess overland flow (Dunne, 1978; Frankenberger et al., 1998). Saturated areas, which expand and contract with time, often form in areas where subsurface lateral flow converges, the slope changes, or where depth to the restricting layer decreases. Runoff and nonpoint source pollution originate on these variable source areas. An important strategy for minimizing nonpoint source pollution, therefore, is to identify areas likely to become saturated so that potential pollutant loading can be minimized there. When soils are frozen, runoff is produced mainly by Hortonian flow or infiltration excess (Hillel, 1980). Under these conditions, the entire watershed likely will contribute runoff and nonpoint source pollution.

In shallow soils, like those in the Palouse Region, the assumption that the hydraulic gradient is equal to the land slope greatly simplifies distributed modeling on a watershed scale. Digital data give information about slope and soil parameters at any point, and land use and climate data allow a distributed estimation of evapotranspiration and Hortonian flow. Several models have used this assumption but vary in data requirements and calibration parameters (Frankenberger et al., 1998; Wigmosta et al., 1994; Grayson et al., 1992; Beven, 1989). Development of a problem-solving tool for management purposes requires a model with minimal data requirements and calibration. The Soil Moisture Routing (SMR) model by Frankenberger et al. (1998) appears to be well-suited for application to the Palouse Region. A similar model to SMR is SMoRMoD also was presented by Zollweg et al. (1996).

The SMR model is a simple distributed water balance model run on a daily time step, which can be used to determine the probability of saturation-excess overland flow occurring at any point
in the watershed (Frankenberger et al., 1997; Frankenberger, 1996; Frankenberger et al., 1998).
It was developed at Cornell University with the objective of aiding in management decisions in
potential runoff source areas of watersheds with shallow soils. An unique feature of the SMR
model is its ability to predict variable source areas and streamflow based on publicly available data
with minimal calibration.

The overall objective of this paper is to present the application and modification of the SMR
model to perched water table conditions of the Palouse Region in Idaho and Washington State. We
will first show how the original SMR model was applied without any modifications. Then we will
show how the SMR model was modified based on observations from a field site in Troy, ID.
Results in this paper will compare observed and simulated perched water table depths. This type
of validation is more challenging than previously done for the SMR model of Frankenberger et al.
(1998) who mainly used streamflow hydrographs to validate the model.

THE SOIL MOISTURE ROUTING MODEL

SMR is based on a hydrologic model for shallow soils of Steenhuis and Norman (1986),
adapted into the GIS GRASS by Caraco (unpublished report, 1992) and Zollweg (1994) who used
a similar model to predict streamflow event hydrographs. Frankenberger (1996) and
Frankenberger et al. (1998) further developed the model performing validation and sensitivity
analyses on a watershed in the Catskills Region of New York State. The model is written as a
sequence of commands within GRASS. The version reported by Frankenberger et al. (1998)
assumes one soil layer above the restrictive layer.

The soil types that best fit this model consist of a relatively thin, permeable soil layer over a
much less permeable fragipan, bedrock, or other restricting layer. This profile is typical of upland
soils in the glaciated regions of the United States where a fragipan or bedrock limits root growth
and water movement. The model is most effective in areas of steep topography, where slopes are
significant enough to be the main cause of lateral flow.

Water Balance

SMR is based on a water balance at each time step for each grid cell of the area of interest.
The original SMR model is run on a daily time step. Cells size is optional but is typically of
dimension 10 m to 30 m. Soil moisture content for each cell is predicted, and any moisture above
saturation results in surface runoff. Inputs to each cell are daily precipitation and lateral flow from
uphill cells, while outputs are lateral flow to downhill cells, percolation into the subsurface,
evapotranspiration, and surface runoff. For each cell, the water balance equation is written as

\[ D_i \frac{d\theta_i}{dt} = P(t)_i - ET(t)_i + \sum Q_{in,i} - \sum Q_{out,i} - L_i - R_i \]  

where

- \( i \) = cell address
- \( D_i \) = depth of root zone of the cell
- \( \theta_i \) = volumetric moisture content of the cell
- \( P \) = precipitation (rain + snowmelt)
- \( ET_i \) = actual evapotranspiration
- \( \sum Q_{in,i} \) = lateral inflow from surrounding upslope cells
- \( \sum Q_{out,i} \) = lateral outflow to surrounding downslope cells
- \( L_i \) = leakage out of the surface soil layer to bedrock
- \( R_i \) = surface runoff
- \( t \) = time
Calculation of the water balance is facilitated by the GIS, which keeps track of input parameters such as elevation, soil data, slope, land use and flow direction as well as the moisture stored in each cell at each time step using simple GIS commands. The model components in Eq. (1) are illustrated in Figure 1. They will be described briefly in the following sections. More details on the SMR model can be found in Frankenberger et al. (1998).

**Evapotranspiration**

![Conceptual hydrologic model](image)

**Precipitation**

Precipitation consists of all moisture inputs to the cell, including rainfall and snowmelt. Saturated hydraulic conductivities of soils where the model applies are generally higher than rainfall intensities, so all rainfall is assumed to infiltrate unless the soil is saturated or the area is disturbed, compacted or frozen.

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 calculate snowmelt:

\[ M = mt + k, \text{ if } t > 0 \degree C \] (2)

where \( M \) is snowmelt (cm/day) and \( t \) is average daily temperature (°C). The snowmelt factor \( m \) has values of 0.23 cm/°C in forested areas and 0.27 cm/°C in non-forested areas while the constant \( k \) is 0 in forested areas and 1.22 cm in non-forested areas (U.S. Army Corps of Engineers, 1960).

**Evapotranspiration**

Evapotranspiration is calculated for each cell as a function of daily potential evapotranspiration, vegetation and stage of growth, and moisture content in the cell:

\[ ET_i = PET(ET/PET)_i c_i \] (3)

where

- \( PET \) = potential evapotranspiration
- \( (ET/PET)_i \) = the ratio of actual to potential evapotranspiration based on soil moisture content
- \( c_i \) = a vegetation coefficient which varies throughout the year for each vegetation class.

The (ET/PET) ratio is calculated using the relationship developed by Thornthwaite and Mather (1955) of evapotranspiration at the potential rate when the matric potential is less than 1/3 bar. No
evapotranspiration takes place when moisture content is below wilting point, and a linear relationship is assumed between wilting point and 1/3 bar moisture contents.

**Subsurface lateral flow**

Shallow subsurface lateral flow, or interflow, is a key component of the water balance. The quantity of lateral flow out of each cell is calculated from Darcy’s Law, approximating the hydraulic gradient by the land slope, $\beta_i$, at each cell $i$:

$$Q_{out.i} = wK(\theta_i)D_i\beta_i$$

where

- $Q_{out.i}$ = lateral flow out of cell $i$
- $D_i$ = depth to restrictive layer for cell $i$
- $w$ = width of each cell
- $K(\theta_i)$ = hydraulic conductivity of the soil profile at cell $i$

The hydraulic conductivity depends on soil moisture content, $K(\theta)$. If the average moisture content through the soil profile, $\theta_i$, in cell $i$ is less than field capacity, $\theta_{fc}$, then $K(\theta_i)$ is calculated based on $\theta_i$ of the profile using an exponential relationship

$$K(\theta_i) = K_s \exp\left(-\alpha \frac{\theta_s - \theta_i}{\theta_s - \theta_r}\right)$$

where

- $K_s$ = saturated hydraulic conductivity
- $\theta_r$ = residual moisture content
- $\theta_s$ = porosity
- $\alpha$ = a constant equal to 13 (Bresler et al., 1978; Steenhuis and Van der Molen, 1986)

Because of the restricting layer at shallow depth, significant vertical water movement stops when the soil becomes unsaturated at the deepest part of the profile. Field capacity is therefore defined as the moisture content when the soil becomes just saturated at the restricting layer interface (Steenhuis et al., 1988). When soil moisture is above field capacity, $\theta_{fc}$, the moisture profile is simplified as a step function, with soil at field capacity overlying soil at saturation above the restricting layer. The hydraulic conductivity of the soil profile then is:

$$K(\theta_i) = (K_s - K(\theta_{fc}))(\frac{\theta_i - \theta_{fc}}{\theta_s - \theta_{fc}}) + K(\theta_{fc})$$

Lateral flow is divided among all cells that are downhill from a particular cell. A multiple flowpaths algorithm (Quinn et al., 1991) is used to allocate to each neighbor a portion of the total flow depending on (a) the elevation difference between it and cell $i$, and (b) the distance between the cells. For any neighbor $j$ of cell $i$,

$$P_{ij} = \frac{(Z_i - Z_j)/L_j}{\sum_{j=1}^{n}[(Z_i - Z_j)/L_j]}$$

where

- $P_{ij}$ = the portion of the total flow out of cell $i$ routed to neighbor $j$
- $Z_i$ and $Z_j$ = elevations of cell $i$ and its neighbor $j$, respectively
- $L_j$ = the distance from the center point of cell $i$ to neighbor $j$
- $n$ = the number of downslope neighbors of cell $i$. 
Leakage out of the root zone layer

If a saturated layer is present above the restricting layer or fragipan, water can percolate downward. An "effective conductivity", $K_{\text{sub}}$, of the restrictive layer is specified which limits the rate at which water can leak out of the root zone.

Compacted and frozen areas

Compacted and frozen areas may generate Hortonian or infiltration-excess overland flow, which are modeled in an approximate way by the SMR model by artificially changes the soil depth to near zero. These areas may include parts of a farmstead, intensively cropped fields where infiltration has been reduced by growing continuous row crops or other poor field practices, or when surface soil temperatures are below 0 °C. Algorithms for frozen conditions have not been field tested to date.

Surface runoff

The main cause of surface runoff is soil moisture exceeding the limited moisture storage in the soil profile. Moisture storage capacity is exceeded when inputs from precipitation and lateral flow in exceed outputs to evapotranspiration, lateral flow downslope and percolation by more than the available moisture storage in the soil profile.

Outputs

Many outputs can be derived from this GIS-based model: days of saturation, runoff amounts, stream flow, moisture storage in the profile, etc. All these outputs can be grouped for different cells at any time period desired (e.g., monthly, seasonally, yearly). In this study, we calculated perched water table heights from moisture storage by subtracting field capacity in the zone above the saturated layer from moisture storage in the profile.

MATERIALS AND METHODS

Site Description

The model was tested on a watershed, approximately 1 ha in size, near Troy, Idaho. The site is typical of the Palouse Region with a perched water table due to fragipan at shallow depth. A seasonal stream drains the watershed. Land use in the watershed is undisturbed grassland, part of the Conservation Reserve Program. The soils in the watershed are classified as the Santa Series, which consist of moderately well-drained soils with a moderately deep profile extending to a fragipan. They were formed in deep loess with small amounts of volcanic ash. The taxonomic class is coarse-silty, mixed, frigid Ochreptic Fragixeralfs. The profile contains three genetic horizons, A, Bw1, E, and a fragipan, Btxb. The A horizon is a yellowish brown silt loam that is dark brown when moist with a subangular blocky structure that is 0 to 38 cm deep. The Bw horizon is a brown silt loam with a prismatic structure and is 38 to 68 cm deep. The E horizon is a pale brown silt loam with a massive structure that is slightly hard and is 68 to 86 cm deep. The Btxb horizon, or fragipan, is a yellowish brown silt loam and silty clay loam soil with a coarse prismatic and medium angular blocky structure that is very hard, firm, and brittle and ranges from 86 to 165 cm deep. The mean annual precipitation for the eastern Palouse ranges from 500 mm in the west to over 830 mm in the east. More than 60% of the annual precipitation occurs from November to April with low intensity rainfall or snowmelt. Soils at the site were covered with snow most of the time during the testing period preventing soil frost.

Field data collection

A 10m x 15m grid was established in the test watershed for installation of piezometers. A tractor-mounted hydraulic core sampler was used to extract 8.9 cm diameter soil samples to the

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1 For ease of use in variables B is substituted for Bw in the remainder of the text
fragipan from each point on the grid. Each soil sample was used to identify genetic horizons and their thickness. In addition, the following soil properties were determined in the laboratory for each horizon: saturated hydraulic conductivity, porosity, field capacity, and wilting point. Fragipan conductivity ($K_{sub}$) was derived from Soil Survey report (Barker, 1981). Table 1 lists these properties.

Table 1. Soil physical parameters used in SMR model simulations.

<table>
<thead>
<tr>
<th>Input Soil Parameters</th>
<th>Soil Layer</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Lateral) Saturated Hydraulic Conductivity</td>
<td>A</td>
<td>160 cm/day</td>
</tr>
<tr>
<td></td>
<td>B</td>
<td>50 cm/day</td>
</tr>
<tr>
<td></td>
<td>E</td>
<td>20 cm/day</td>
</tr>
<tr>
<td>Saturated hydraulic conductivity</td>
<td>Restricting layer</td>
<td>0.01 cm/day</td>
</tr>
<tr>
<td>Saturated moisture content</td>
<td>A</td>
<td>51 %</td>
</tr>
<tr>
<td></td>
<td>B</td>
<td>43 %</td>
</tr>
<tr>
<td></td>
<td>E</td>
<td>38 %</td>
</tr>
<tr>
<td>Field capacity moisture content</td>
<td>A</td>
<td>38 %</td>
</tr>
<tr>
<td></td>
<td>B</td>
<td>28 %</td>
</tr>
<tr>
<td></td>
<td>E</td>
<td>27 %</td>
</tr>
<tr>
<td>ET reduction moisture content</td>
<td>A</td>
<td>38 %</td>
</tr>
<tr>
<td>Wilting point moisture content</td>
<td>A</td>
<td>11 %</td>
</tr>
<tr>
<td></td>
<td>B</td>
<td>8 %</td>
</tr>
<tr>
<td></td>
<td>E</td>
<td>8 %</td>
</tr>
<tr>
<td>Residual moisture content</td>
<td>A</td>
<td>8 %</td>
</tr>
<tr>
<td></td>
<td>B</td>
<td>8%</td>
</tr>
<tr>
<td></td>
<td>E</td>
<td>8%</td>
</tr>
</tbody>
</table>

For use in model simulations, average soil properties for the soil above the fragipan also were calculated as either a depth-weighted arithmetic mean. For example, the average field capacity moisture content was calculated as

$$\theta_{fc,avg} = \frac{\theta_{fc,A}d_A + \theta_{fc,B}d_B + \theta_{fc,E}d_E}{d_A + d_B + d_E}$$

(8)

where:

- $\theta_{fc,avg}$ = field capacity for the entire soil depth
- $\theta_{fc,A}$, $\theta_{fc,B}$ and $\theta_{fc,E}$ = field capacity of the A, B, and E soil horizons
- $d_A$, $d_B$ and $d_E$ = soil depth of the A, B, and E soil horizons

A similar equation was used to calculate the average saturated moisture content, $\theta_{s,avg}$, or the average saturated hydraulic conductivity, $K_{s,avg}$.

Following core removal, piezometers constructed of PVC were installed at each core location and equipped with pressure transducers (PX26-005DV, OMEGA Technology) connected to CR-10 dataloggers (Campbell Scientific, Inc.). The transducers were calibrated in the laboratory. The piezometers were backfilled with fine sand at the base to promote good hydraulic contact and with Bentonite for the remainder to prevent vertical water movement along the piezometer (Young,
Perched water table heights were recorded twice daily since November, 1995. For purposes of testing the SMR model, data based on a 5m x 5m grid were obtained through interpolation. A 5m x 5m DEM was surveyed following installation of the piezometers. Figure 2 shows the piezometers in the watershed on a 10m x 15m grid as well as the 5m x 5m DEM.

Figure 2. Digital elevation model (DEM) and piezometer grid at the Troy, ID catchment

Weather data
During the period of this study, a weather station was not operational at the site. Precipitation data, therefore, were taken from two different locations. Daily rainfall, snow depth, air temperature, and pan potential evapotranspiration (PET) for the growing season were obtained from an independent observer at Helmer, Idaho, approximately 40 km northeast of Troy. These data are available from the University of Idaho’s Climatology Laboratory. PET data for the non-growing season were obtained from the weather station at Prosser, Washington. Figure 3 shows inputs of rain, snow water equivalent (SWE) and air temperature for the test period.
Saturated hydraulic conductivity derived from observed data

Besides laboratory derived values for saturated hydraulic conductivity of the soil layer above the fragipan, we derived depth-integrated or average saturated hydraulic conductivity values ($K_{s,avg}$) for each 10m x 15m grid cell from time series of field measured water table depth. These field-derived $K_{s,avg}$ values were determined using the “goal seek” tool in an Excel spreadsheet by iteratively solving the water balance equation (Eq. (1)) for one cell over a period of $\delta t$ days:

$$\begin{align*}
0 &= S_{i,t} - S_{i,t+\delta t} + \frac{\left(\sum Q_{in}\right)_{i,t} + \left(\sum Q_{in}\right)_{i,t+\delta t}}{2} \delta t - \frac{Q_{out,i,t} + Q_{out,i,t+\delta t}}{2} \delta t \\
&+ P(t) - ET(t) - L_i - R_i
\end{align*}$$

(9)

in which the left hand side of Eq. (1) is replaced by

$$S_{i,t} = \theta_{fc,avg} \left(d_A + d_B + d_E\right) + h_{i,t} \left(\theta_{s,avg} - \theta_{fc,avg}\right)$$

(10)

where

- $S_{i,t}$ = depth of water stored in the soil profile of cell $i$ at time $t$
- $h_{i,t}$ = depth of water observed in the piezometer of cell $i$ at time $t$

All other components in Eq. (9) were explained in the description of SMR above. When a cell consisted of both saturated soil and unsaturated soil, it was assumed that all unsaturated soil was at
field capacity moisture content. Flow width, \( w \), was set at 10 m and \( K_{\text{sub}} \) was kept constant at 0.01 cm/day. Flow routing was based on surface elevations using eq. (7).

All parameters in Eq. (9) are known from field or laboratory measurements and from public sources except for the \( K_{\text{s,avg}} \) which is part of Eq. (6). In this Equation, the actual moisture content of the cell is calculated as

\[
\theta_{i,t} = \frac{S_{i,t}}{d_A + d_B + d_E}
\]  

(11)

\( K_{\text{s,avg}} \) values were first determined at position (8,10) in the watershed where no upslope water was contributing. Cells on the upper perimeter of the piezometer grid where \( K_{\text{s,avg}} \) could not be calculated due to unknown upslope lateral inflow were assigned the \( K_{\text{s,avg}} \) equal to that found for position (8,10) assuming that the soil type and structure of those cells are similar to position (8,10). The flowpath algorithm of Eq. (7) then dictated the sequence of applying Eq. (9) to neighboring downslope cells provided data along the entire flow path were available. Due to missing data along some flow paths at positions (6,6), (5,4), (5,3), (4,6), (3,10), (3,9), and (2,5), \( K_{\text{s,avg}} \) values were calculated mostly for the upper section of the Troy, ID catchment.

To minimize error, calculations only were made during time periods which had negligible precipitation and runoff. For position (8,10) we evaluated 14 different drainages. For other positions \( K_{\text{s,avg}} \) values were calculated for two different drainage periods: 12-22-95 12:00 AM to 12-23-95 12:00 AM and 3-30-97 12:00 PM to 3-31-97 12:00 PM. Piezometers at positions (2,2) and (5,7) had data only for the second drainage (3-30-97).

Model Simulations

Cell sizes of 5 x 5 m required model simulations at an hourly time step as opposed to the original daily time step to avoid truncation of flow at high hydraulic conductivity values (i.e., at conditions close to saturation). Five simulations were performed:

Simulation I: This simulation applies the original SMR model to data from the field site at Troy, ID. All measured soil properties were averaged as described above (e.g., eq. (8)) to obtain one value for the soil above the restrictive layer.

Simulation II: In this simulation, we included a multiplier for the saturated hydraulic conductivity \( (K_s) \) in Eq. (6) based on results from the analysis of perched water table depths described above. Eq. (6) therefore was modified as follows:

\[
K(\theta_t) = (XK_s - K(\theta_{fc}))\left(\frac{\theta_t - \theta_{fc}}{\theta_s - \theta_{fc}}\right) + K(\theta_{fc})
\]

(12)

where \( X \) is the multiplier and all other parameters are as defined previously.

Simulation III: Simulation III was run after the original one-layer version of SMR was modified to include three soil horizons (A, B, and E) above the restrictive layer. In this modification, evapotranspiration is a function of the soil moisture in the A horizon, and perched water first develops in the E horizon, then the B and, finally, the A horizon.

Simulation IV: Simulation IV was similar to simulation III but included the same \( K_s \) multiplier (X) as in simulation II. This multiplier was the same for all three horizons.

Simulation V: This simulation is similar to the previous simulation except we decreased the multiplier X in order from horizon A, B to E approximating an exponential decay of the hydraulic conductivity with depth proposed by Beven (1982).
RESULTS AND DISCUSSION

We first show results of the analysis performed on perched water table data to obtain the $K_{s,avg}$. Next, we show results of model simulations and observed perched water table data from the Troy, ID site for the dates of November, 1995 through May, 1996. Table 2 summarizes characteristics of the five model simulations performed in this study. The X multiplier is defined in Eq. (12).

Table 2. Summary of model simulations I through V.

<table>
<thead>
<tr>
<th>Simulation no.</th>
<th>no. of layers above fragipan</th>
<th>X multiplier</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>II</td>
<td>1</td>
<td>5</td>
</tr>
<tr>
<td>III</td>
<td>3 (A, B, E)</td>
<td>A: 1, B: 1, E: 1</td>
</tr>
<tr>
<td>IV</td>
<td>3 (A, B, E)</td>
<td>A: 5, B: 5, E: 5</td>
</tr>
<tr>
<td>V</td>
<td>3 (A, B, E)</td>
<td>A: 10, B: 5, E: 2</td>
</tr>
</tbody>
</table>

Saturated Hydraulic Conductivities Derived from Observed Data

Saturated hydraulic conductivity ($K_{s,avg}$) values derived using the procedure described above for the soil at the Troy, ID site were found to be up to one order of magnitude larger than values obtained from the laboratory or reported in the Latah County Soil Survey (Barker, 1981). At position (8,10), where $K_{s,avg}$ values were derived from 14 different drainage periods, the average and standard deviation are approximately 400 cm/day and 110 cm/day, respectively (Table 3). Similar values were obtained for other positions along the grid. The depth weighted average of $K_s$ values for each horizon determined in the laboratory across the grid is approximately 77.1 cm/day.

Table 3. Field-derived $K_{s,avg}$ for position (8,10) at Troy, ID site using 14 different drainage periods.

<table>
<thead>
<tr>
<th>From Date</th>
<th>To Date</th>
<th>Initial Piezo Depth cm</th>
<th>Final Piezo Depth cm</th>
<th>Total Days</th>
<th>Predicted $K_{sat}$ cm/day</th>
</tr>
</thead>
<tbody>
<tr>
<td>12/4/95</td>
<td>12/9/95</td>
<td>26</td>
<td>14</td>
<td>5</td>
<td>304.1</td>
</tr>
<tr>
<td>12/16/95</td>
<td>12/28/95</td>
<td>34</td>
<td>7</td>
<td>12</td>
<td>303.1</td>
</tr>
<tr>
<td>1/10/96</td>
<td>1/13/96</td>
<td>47</td>
<td>32</td>
<td>3</td>
<td>386.3</td>
</tr>
<tr>
<td>1/16/96</td>
<td>1/19/96</td>
<td>47</td>
<td>29</td>
<td>3</td>
<td>452.8</td>
</tr>
<tr>
<td>3/8/96</td>
<td>3/10/96</td>
<td>31</td>
<td>26</td>
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Avg. = 395.6
Std. dev = 113.0
N = 14
Figure 4 shows the ratio of field-derived $K_{s,avg}$ values for the 3-30-97 drainage and $K_{s,avg}$ from the laboratory. The predicted $K_{s,avg}$ ranges from 3 to 13 times larger than the $K_{s,avg}$ from the laboratory.

![Table with data]

Locations without adequate information to calculate $K_{s,avg}$ are shaded gray.

Figure 4. Ratio of field-derived $K_{s,avg}$ for 3-30-97 drainage and $K_{s,avg}$ from laboratory values (equal to approximately 77.1 cm/day across the grid).

Field-derived $K_{s,avg}$ values are larger than $K_{s,avg}$ values from the laboratory due to presence of preferential subsurface lateral flow paths. Typically, $K_{s,avg}$ values from the laboratory are based on soil core measurements representing mostly matrix flow. When upscaling to a larger catchment, macroscopic elements such as large pores play an important role. In a nearby watershed two orders of magnitude larger than the Troy, ID catchment, drought flow analysis by Brutsaert and Nieber (1997) and Brutsaert and Lopez (1998) resulted in $K_{s,avg}$ estimates 100 times larger than those in Soil Survey reports. Grayson et al. (1992), Wigmosta et al. (1994) and Frankenberger et al. (1998) also found that model simulations improved dramatically when increasing the saturated hydraulic conductivity by two orders of magnitude.

The field-derived $K_{s,avg}$ values are only approximate values as a result of three assumptions made in the analysis procedure: (i) a constant saturated hydraulic conductivity of the restrictive layer ($K_{sub}$), (ii) routing using surface elevation, and (iii) a constant flow width of 10 m. We briefly discuss these assumptions below.

$K_{sub}$ may vary from the constant value of 0.01 cm/day across the site due cracks and other preferential pathways. For example, if $K_{sub}$ at position (5,8) was 1.0 cm/day then the field-derived $K_{s,avg}$ for the 12-22-95 drainage reduced from 812 cm/day to 486 cm/day. Alternately, assuming
that $K_{s,\text{avg}}$ from the laboratory is correct, $K_{\text{sub}}$ must equal approximately 2.0 cm/day at position (5,8), an unlikely increase by two orders of magnitude.

Although we do not know the accuracy of the routing algorithm in SMR at the Troy, ID catchment, results of routing based on surface elevation and based on elevation of the restrictive layer appear similar (result not shown). If there are errors in the routing procedure, some cells will be predicted to receive more water and some less water than what actually occurs resulting in under or over-prediction of $K_{s,\text{avg}}$. Since field-derived $K_{s,\text{avg}}$ for position (8,10), which is not affected by the routing algorithm, is similar to the field-derived $K_{s,\text{avg}}$ at downslope positions errors due to routing probably are minimal.

A flow width of 10 m for actual grid sizes of 10m x 15m slightly under-predicted the field-observed $K_{s,\text{avg}}$ values. The 10 m width at the Troy, ID catchment faces south and north, which is the predominant direction of flow. In most cases, however, the actual flow width of the grid cell should range somewhere between 10 m and 15 m. If the flow width was assumed at a constant of 15 m during our analysis, then the field-derived $K_{s,\text{avg}}$ values would have been 3/2 larger than when using the 10 m width. The $K_{s,\text{avg}}$ values at the Troy, ID catchment still would be up to one order of magnitude larger than values from the laboratory.

**Simulations I through V**

Simulations are shown in Figures 5 to 9 comparing observed and predicted perched water table depths for three representative positions in the piezometer grid. Position (2,3) is near the bottom of the catchment, position (4,5) is near the center and position (8,10) is at the top of the catchment. Table 4 quantifies the goodness-of-fit between observed and predicted using the Nash-Sutcliffe efficiency coefficients (Nash and Sutcliffe, 1970) and corresponding normalized standard error for each simulation. The Nash-Sutcliffe coefficient (NS) appears to correlate best with visual observations and will be referred to below. When NS equals one, the predicted and observed data are identical.

As the observed data show (e.g., Figure 5), perched water tables respond very fast to precipitation inputs. Drainage following these peaks is quite variable across the site. For example, at position (2,3) drainage is very rapid, while at positions (4,5) and (8,10) drainage is gradual with sudden breaks in the recession curves. Overall, the observed data show three characteristic drainage stages from November through May. From about November to January, the soil layers above the fragipan drain (nearly) to the fragipan depth (see position (2,3)), probably due to larger cracks in the E horizon and/or fragipan. When these cracks close, from January to late April, drainage is stalled near the interface of the B and E horizons. After April, when evapotranspiration increases, drainage occurs to the fragipan depth. We now briefly discuss results of each simulation.

<table>
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<tr>
<th>Simulation</th>
<th>Position (2,3)</th>
<th>Position (4,5)</th>
<th>Position (8,10)</th>
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<tr>
<td></td>
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<td>Norm. SE</td>
<td>NS</td>
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<tr>
<td>I</td>
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<td>V</td>
<td>-0.57</td>
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Simulation I: the original SMR model under-predicted the amount of drainage at all three positions shown in Figure 5 except after March 1, 1996, at position (4,5).

Simulation II: after increasing $K_s$ by a factor of 5, SMR mostly over-predicted drainage at all three positions. Interestingly, at positions (2,3) and (8,10) SMR reasonably predicted drainage between November and January and after April.

Simulation III: the three-layer version of SMR greatly under-predicted drainage at all positions. Nash-Sutcliffe efficiency coefficients in Table 4 indicate that the original SMR model performed better than the three layer model, probably because the average $K_s$ for one layer drains the E horizon faster.

Simulation IV: the three-layer model with a uniform multiplier of 5 across the soil profile visually improved predictions of drainage at all three positions, especially at position (8,10). Nash-Sutcliffe efficiency coefficients confirm the visual observation. Note, however, that drainage at positions (2,3) and (4,5) was over-predicted from January to April, and that drainage between November and January is not well-predicted at positions (2,3) and (8,10).

Simulation V: the assumption of a strong decay in $K_s$ with depth appears to provide the best prediction of drainage at all positions. Visual observations are in agreement with the Nash-Sutcliffe coefficients in Table 4. Note again that early drainage in November and December is not well-predicted at positions (2,3) and (8,10).

Results of the simulations clearly show that the SMR model is well-suited for soils of the Palouse Region, especially when field-derived $K_s$ values are used. Differences between observed data and SMR predictions of all simulations indicate, however, that $K_s$ values vary with time as larger pores in the overlying soil and/or fragipan open and close. In addition, results from simulation V indicate that in soils of the Palouse Region some exponential decay of $K_s$ with depth exists (Beven, 1982). Our inability to consistently simulate the exact shape of recession curves also suggests that variability of $K_s$ across the catchment may be significant. Analysis of data from 1997 and 1998 is currently underway to quantify the temporal and spatial variability of $K_s$ at the Troy, ID catchment, as well as the conductivity of the fragipan and underlying strata.
Figure 5. Predicted (+) versus observed (●) water table depth above the restricting soil layer based on original SMR model with one soil layer (Simulation I).
Figure 6. Predicted (+) versus observed (●) water table depth above the restricting soil layer based on the SMR model with one soil layer and a multiplier $X = 5$ (Simulation II).
Figure 7. Predicted (+) versus observed (●) water table depth above the restricting soil layer based on the SMR model with three soil layers (Simulation III).
Figure 8. Predicted (+) versus observed (•) water table depth above the restricting soil layer based on the SMR model with three soil layers and a multiplier $X = 5$ for each layer (Simulation IV).
Figure 9. Predicted (+) versus observed (●) water table depth above the restricting soil layer based on the SMR model with three soil layers and multipliers $X = 10, 5, \text{ and } 2$ for the A, B, and E layer, respectively (Simulation V).
CONCLUSIONS

Results from this paper show that the simple GIS-based distributed SMR model is well-suited for perched water table conditions of the Palouse Region. From observed data we confirmed findings by others elsewhere that the true saturated hydraulic conductivity for a small catchment can be at least one order of magnitude greater than values determined in the laboratory. Therefore, use of conductivity values from soil survey reports in hillslope hydrology models should be done with caution.

Accuracy of model predictions is highly sensitive to the saturated hydraulic conductivity. Comparison of results from model simulations to observed data highlights the temporal and spatial variability of the saturated hydraulic conductivity across the catchment. Predictions of perched water table depths were most consistent when field-derived hydraulic conductivities were used while approximating an exponential decay with depth.

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