On evaluating small-scale variations of hydrologic processes in time and space

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ON EVALUATING SMALL-SCALE VARIATIONS OF HYDROLOGIC PROCESSES IN TIME AND SPACE

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ABSTRACT

On Evaluating Small-Scale Variations of Hydrologic Processes in Time and Space

by

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This study focuses on representing the small variations in hydrologic properties at various scales. Numerical experiments are used to explore the effects of the distributions of various hydrologic processes at various scales. The model used in this study is a physically based distributed hydrologic model system (HMS). The HMS is implemented in a 7.29 km² sub-watershed within the Susquehanna River Basin in Pennsylvania. Rasterized data sets such as topographic data, soil data, and land use/land cover data are used as input to the HMS. Geographic Information System package is used to preprocess the raster data sets. Stochastic approaches are applied to account for small-scale variations of hydrologic properties that are traditionally viewed as homogeneous. 50 simulation runs are conducted at various spatial and temporal scales. The results show that 100 meters in space and 15 minutes in time are optimal scales for accurate and efficient simulations. Scale factor functions are developed based on the numerical experiments. Hydrologic responses at large scales can be predicted by a scale factor.
based on the simulated responses at small scales. This study provides an alternative to
the hydrologic simulations at different scales.
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CHAPTER 1

INTRODUCTION

In hydrologic simulation practices, the size of the simulated river basin can vary from several to millions of square kilometers, i.e., a scale from a small watershed to a continent. As noted by Minshall (1960) and Amorocho (1961), the watershed scale exerts an influence on simulated hydrologic responses. In distributed modeling, the size of the grid cells varies although a grid divides the underlying surface into a number of small cells. Studies show that the size of the cells can affect the simulated results (Zhang and Montgomery, 1994; Wolock and Price, 1994). The size change is also referred to as the scale change in hydrologic simulations. Problems raised in hydrology concerning the scale change are referred to as "scale problems" or "scale issues". Studies on scale problems in hydrology were intensified in the 80's. In recent years, there are urgent needs to better represent the small-scale hydrologic processes in regional and global climate models due to increased greenhouse effect and global warming concerns (Yu et al., 1999a, Yu et al., 1999b).

This study is intended to provide some understandings into the scale problems in hydrologic simulations with a distributed approach. Hydrologists have made persistent efforts in research on the scale problem, but the yields are not as fruitful as expected. As noted by Beven (2001), who developed the well-known TOPMODEL system, that "Yet, despite all the paper and internet traffic expended on this topic, there seems to have been
very little true progress. This should not really be surprising, in fact, since both aggregation and disaggregation problems are inherently impossible to resolve in hydrology …”.

1.1 Hypothesis

Hydrologic processes are extended both in time and space. Very commonly the observations are made at small temporal and spatial scales and estimates are needed for large scales. On the other hand, the chances of utilizing data from large-scale observations and simulations in small-scale predictions also exist (Jensen and Mantoglou, 1993). So there are interpolations and extrapolations in the operations, i.e., information transformations across scales.

One of the early observations is that at large watershed scales, the simulated runoff is not sensitive to rainfall intensity changes measured at individual gage stations (Amorocho, 1961). In 1982, Dooge concluded that linking phenomena at field scales (10 km²) and watershed scales (1,000 km²) was unresolved. Both cases show that applying directly small-scale observations to large-scale simulations will induce some bias. The reason for that is with the change of scales, the change of hydrologic heterogeneities (e.g., variations in precipitation, topography, land use/land cover, and soil properties) is not well accounted for.

Large-scale hydrologic properties are easy to obtain but difficult to use. With the increasing availability of remotely sensed topography, soil, and land use/land cover data at fine resolutions, it is now possible to estimate hydrologic properties in fine-scale simulations. Usually this estimation needs to implement hydrologic models to bridge the
scale gap. Lumped and distributed models are two common types of hydrologic models. Traditional lumped models tend to ignore the hydrologic heterogeneities within the simulated region. Therefore, lumped models are not capable of providing insight into the understanding of the effects of small-scale hydrologic heterogeneities. Physically based distributed hydrologic models emphasize processes in individual grid cells (Beven and Kirkby, 1979; Abbott et al., 1986; Arnold et al., 1989; Grayson et al., 1992; Paniconi and Wood, 1993; Yu and Schwartz, 1998). Because the grid size can be as small as 2 meters by 2 meters in the distributed model (Zhang and Montgomery, 1994), observations from both laboratory and field can be input directly into the model.

Thus it is hypothesized that distributed hydrologic models are superior to lumped models and capable of evaluating effects with regard to the scale change based on the above discussions. Scale functions are then developed for scaling hydrologic properties among different scales.

1.2 Objectives

The main objectives of this study are: (1) to understand the hydrologic responses in a 7.29 km² sub-watershed of the Susquehanna River Basin (WE-38, named after the outlet gauging station) to changes in parameter schemes, (2) to construct transportable schemes that can be used to upscale and downscale hydrologic variables in practice. The steps for achieving these objectives are to: (1) collect and compile field observed data for analysis and model use, (2) pre-process various data sets using a Geographic Information System (GIS) package (ARC/INFO), (3) implement the HMS in the WE-38 watershed, (4) develop parameterization schemes for different hydrologic processes in the WE-38...
watershed. (5) assess effects of small-scale spatial variability in precipitation and hydraulic parameters on the hydrologic responses, and (6) provide the methodology for applying the HMS at large scales (i.e., regional).

1.3 Uniqueness of the Research

Understanding the behaviors of various hydrologic processes at small scales directly helps better parameterize the various hydrologic variables in the adopted HMS: it provides a basis for applying the HMS at large scales without losing the fundamental physics; it also helps develop a general scheme for upscaling and downscaling various hydrologic processes among different scales. The research is unique because it develops scaling functions to be used in scaling hydrologic properties among different scales.

1.4 Methodology

Procedures performed in this study include: selecting grid resolutions for the digital elevation model (DEM) of the study area for the hydrologic simulations, comparing and selecting different parameterization schemes of various hydrologic processes, and implementing the HMS in the study area. Digital elevation, meteorological, soil, and land use/land cover data sets are collected to drive the HMS. The ARC/INFO package is used to preprocess digital elevation, soil, and land use/land cover data sets (Yu et al., 2001b). Stochastic methods are applied to account for small variations in hydrologic parameters that are traditionally viewed as homogeneous.

The DEM of the study area is used to evaluate effects of grid resolutions on the topographic features that include elevation, slope, and aspect. The original DEM (at a grid resolution of 15 m) of the study area was resampled into resolutions of 25 m, 50 m.
100 m. and 200 m by using ARC/INFO functions. The criterion used for selecting proper resolutions is that the topographic features show no significant deviation from those calculated from the original DEM. The criteria used are peak value, peak appearing time, and total volume of the hydrographs. Selected grid resolutions are also used to test-run the HMS to examine the effects of grid resolutions on the hydrologic responses. If the simulated hydrographs do not deviate significantly from the observed ones, the grid resolution is proper.

To implement the HMS, hydrologic features of the study area need to be extracted from the DEM of the selected resolution. An interactive command system called “GRID” in the ARC/INFO package performs such tasks. The hydrologic features include the conditioned DEM, flow directions, flow accumulations, basin boundary, and drainage and stream networks. A detailed processing procedure will be discussed in the following sections.

Different hydrologic features can be viewed as different data layers in ARC/INFO. These data layers together with the meteorological, land use/land cover, and soil texture data sets are input data sets for the HMS. Data obtained from the ARC/INFO processing procedures are in the form of grids. The format needs to be transformed into the binary storage format before the data can be input into the HMS.

Parameterization schemes are compared with the consideration of spatial and temporal variations in precipitation and hydraulic parameters. Preferential schemes are obtained and used for scaling hydrologic and hydraulic variables.
1.5 Expected Contributions

Small hydrologic heterogeneities in time and space are specifically important in simulating hydrologic responses to storms at fine scales. The most significant factors that need to be considered first are grid spacing, hydraulic parameters, and precipitation. It is envisioned that the smallest resolution of the grid cells will provide the most accurate representation of features of the landscape, hence they produce the most adequate results in the simulations. The smallest time interval in precipitation is likely to give the most precise results in the responses to small-scale storms (Zhang and Montgomery, 1994).

Variations of hydraulic parameters are important even in traditionally viewed homogeneous subzones. Possibilities of using a distributed approach to obtain accurate simulations are discussed. Scaling functions are developed to facilitate information transformations among different scales.

1.6 Structure of Thesis

Chapter 2 is an overview of the mechanisms of the major hydrologic processes simulated within the hydrologic model system. Based on the information in Chapter 2, Chapter 3 provides the formulations and implementations of the HMS. Chapter 4 describes the information of the study area and the data sets. Chapters 5 and 6 provide the specific analysis of the spatial and temporal distributions of the precipitation and hydraulic conductivity and the implications to the modeling. Chapter 7 provides a discussion of the topographic feature changes with the grid spacing. Chapter 8 describes the model parameterizations and discusses the model simulations and results. Chapter 8 also provides the details of the development of scale factor functions for the total runoff.
Chapter 9 presents a method for developing general scale factor functions. Chapter 10 draws conclusions and outlines possibilities for future study.
CHAPTER 2

MAJOR HYDROLOGIC PROCESSES

This chapter briefly overviews the major hydrologic processes and the mechanisms that are of particular importance to the HMS adopted in this study. This chapter also provides a terminology review.

2.1 Introduction

From a system's view, the hydrologic cycle consists of interrelated components such as precipitation, evaporation, runoff, soil water movement, groundwater movement, and streamflow. Though the intrinsic mechanism in hydrology may not be fully understood (Chow et al., 1988), a breakdown analysis will definitely help understand the structure of the model system.

2.2 Precipitation

When air mass is lifted into the atmosphere, it cools and condenses. Liquid or ice droplets are formed. The particles grow by collision and coalescence. When the particles are too heavy and the upward force can no longer hold them, the particles will drop in the form of rain or snow.
There are three types of precipitation: frontal, convective, and orographic. Frontal precipitation results when a warm air mass is lifted over a cooler air mass by frontal passage. Convective precipitation is caused with convective lifting where the air mass is heated at the ground surface. Orographic precipitation involves a mountain that the air mass is lifted over.

When rain reaches the underlying surface, part of it will be intercepted and held by vegetation. This portion of rainfall is called interception. Interception is a loss of the precipitation in terms of runoff.

The amount of precipitation during a unit time length is called rainfall intensity.

2.3 Evaporation and Evapotranspiration

Water has three phases: solid (ice), liquid, and gas (water vapor). These three phases can transform from one to another under specific conditions. Driven by solar radiation, water molecules both from the land surface and water bodies tend to become vapor. This is evaporation. If the water vapor comes directly from vegetation, then it is transpiration. Both evaporation and transpiration are water loss in a watershed. Traditionally they are called evapotranspiration.

If moisture is readily available at the evaporating surface, the evapotranspiration is called potential evapotranspiration. Actual evapotranspiration never exceeds potential evapotranspiration.

2.4 Overland Flow

After interception and evaporation, precipitation reaches the ground surface. Then the rain water is divided into several portions. Some penetrates into the soil. This part is
called infiltration. Some is detained in depressions on the ground surface. This part is called pond-filling. If there is still some water left, the water will flow along the slope by gravity. The flow is called overland flow.

Initially overland flow takes the form of sheet flow, but it tends to concentrate into small channels of various forms. This is called concentrated flow. Overland flow is the driving force for erosion and non-point source contamination.

Overland flow has two general mechanisms. If the flow occurs across a saturation surface and no water infiltrates, it is called saturation excess overland flow. If the rainfall intensity is greater than the infiltration rate, water cannot infiltrate as fast as rainfall is supplied. This type of overland flow is called infiltration excess overland flow or Hortonian overland flow.

2.5 Soil Water Movement

Infiltration into unsaturated soil is driven by two forces: capillary force and gravity. If the soil is saturated, then the capillary force no longer exists.

Part of the water infiltrates into soil becomes soil moisture, and part of the water continues moving downward until it reaches the underground water table. Soil water also moves laterally because it is driven by a lateral gradient. Laterally moving soil water that reaches the stream bank, seeps out of the bank and recharges the streamflow ultimately, is called through flow.

2.6 Groundwater Flow

When soil water reaches the groundwater table during a storm, a lateral gradient will be built. Groundwater will flow along the gradient. As the result, the groundwater table
Groundwater flow is saturation flow. When groundwater charges the streamflow, it is called return flow (or baseflow).

It is noted that in a watershed, the groundwater catchment boundary may not always correspond with the watershed boundary.

Groundwater flow is a media flow. Usually the flow rate is very slow as compared to surface flow. However, if soil contains macropores, groundwater will flow through these macropores very quickly. This kind of flow is called preferential flow. Preferential flow can be very significant in favorable environments where a large portion of the streamflow is groundwater preferential flow recharge in the early stage of a flood.

2.7 Streamflow

After reaching the river channel, water moves downstream in the form of a kinematic wave, diffusive wave, or dynamic wave. Water reaching the river channel can be in the form of overland flow, through flow, and return flow (or baseflow).

Streamflow is the main response of a watershed to precipitation. Streamflow is always the target of hydrologic simulations.
This chapter provides a detailed description of the theories and formulations implemented in the HMS. The theories, formulations, and implementations of a model system are viewed as the conceptual model. The conceptual model is the key to understanding the detailed structures and implementations of the HMS. An overview of the conceptual model also provides a conceptual validation for the model system.

3.1 Introduction

The hydrologic model system (HMS) adopted in this study integrates four modules or models to simulate different processes in the basin hydrologic cycle. These four models are the Soil Hydrologic Model (SHM), the Terrestrial Hydrologic Model (THM), the Groundwater Hydrologic Model (GHM), and the Channel Groundwater Interaction Model (CGI) (Yu and Schwartz, 1998; Yu et al., 1999a). The structure of the HMS is shown in Figure 1.

The HMS accommodates the spatial heterogeneity of various hydrologic properties by utilizing remotely sensed and digitized data sets such as DEMs, soil, vegetation, and other hydrologic parameters. The advantages of the HMS are that the hydrologic parameters required in the HMS have physical bases and can be related to measured
Figure 1. The structural diagram of the hydrologic model system (HMS) (From Yu et al., 1999a).
hydrologie processes in the field, and inputs and outputs are for individual grid cells. Thus the approach takes the full consideration of the geomorphologic features of the basin.

3.2 Soil Hydrologic Model - SHM

The SHM simulates the vertical movement of the soil water. Evapotranspiration, soil water content, and groundwater recharge are calculated in this model.

3.2.1 Evaporation and Evapotranspiration

Evaporation and evapotranspiration are simulated through the Penman-Monteith Model (Monteith, 1981):

\[
E = \frac{s(T_a)(K + L) + \rho_a C_a C_{at} [e_{at}(T_a)](1 - W_a)}{\rho_a \lambda_a [s(T_a) + \gamma]} \tag{1}
\]

\[
ET = \frac{s(T_a)(K + L) + \gamma \rho_a C_a C_{at} [e_{at}(T_a)](1 - W_a)}{\rho_a \lambda_a [s(T_a) + \gamma [1 + C_{at} / C_{can}]]} \tag{2}
\]

where: \( E \) is the evaporation, \( ET \) is the evapotranspiration, \( T_a \) is the air temperature, \( K \) is the net short wave radiation input, \( L \) is the long wave radiation input, \( \rho_a \) is the mass density of air, \( C_a \) is the heat capacity of air, \( C_{at} \) is the atmospheric conductance, \( e_{at}(T_a) \) is the saturation vapor pressure in air, \( W_a \) is the relative humidity, \( \rho_w \) is the mass density of water, \( \lambda_a \) is the latent heat of vaporization, \( s(T_a) \) is the slope of the relation between saturation vapor pressure and temperature, \( \gamma \) is a factor used in calculating Bowen ratio.

\( C_{can} \) is the canopy conductance, and:

\[
s(T_a) = \frac{25308}{(T_a + 237.3)^2} \exp\left(\frac{17.3T_a}{T_a + 237.3}\right) \tag{3}
\]
\[ c_a = 0.24 \] \hspace{1cm} \text{(4)}

\[ e_{w}(T_a) = 6.11 W_a \exp\left(\frac{17.3T_a}{T_a + 237.3}\right) \] \hspace{1cm} \text{(5)}

\[ \gamma = \frac{c_a P}{0.622 \lambda} \] \hspace{1cm} \text{(6)}

where \( P \) is the atmospheric pressure.

Evaporation and evapotranspiration calculated in SHM are used to estimate the water budget in GHM.

### 3.2.2 Soil Water Content

Soil water movement is governed with the continuity equation and momentum equation. The continuity is in the form:

\[ \frac{\partial \theta(z,t)}{\partial t} + \frac{\partial q(z,t)}{\partial z} = S(z,t) \] \hspace{1cm} \text{(7)}

where \( \theta \) is the volumetric water content, \( t \) is time, \( q \) is Darcy's flux, \( z \) is depth, and \( S \) is a source/sink term. For one-dimensional saturated flow in the vertical direction, the momentum equation is expressed as:

\[ q = -K \frac{\partial h}{\partial z} \] \hspace{1cm} \text{(8)}

where \( K \) is the hydraulic conductivity, \( h \) is the head of the flow. This equation is known as Darcy's Law or Darcy's Equation. For one-dimensional unsaturated flow in the vertical direction, the momentum equation is expressed as:

\[ q = -K \frac{\partial (\psi + h)}{\partial z} \] \hspace{1cm} \text{(9)}

where \( \psi \) is the hydraulic metric potential. Applying Darcy's Equation into the continuity equation yields:
\[
\frac{\partial q(z,t)}{\partial t} = \frac{\partial}{\partial z} \left[ K(z,t) \frac{\partial \psi(z,t)}{\partial \theta} \frac{\partial \theta(z,t)}{\partial z} \right] + \frac{\partial K(z,t)}{\partial \theta} \frac{\partial \theta(z,t)}{\partial z} \cos \chi \tag{10}
\]

where \( \chi \) is the terrain slope angle. The first term of the right hand side of this equation is solved using the Crank-Nicholson method while the second term is solved through forward-in-time-backward-in-space finite differencing (Capehart and Carlson, 1994).

The schemes relate \( \psi \) and \( K \) to a normalized volumetric water content, \( S_v \), as described as:

\[
S_v = \frac{\theta - \theta_s}{\theta_i - \theta_s} \tag{11}
\]

where \( \theta_s \) is the soil water content at saturation, and \( \theta_r \) is the residual soil water content, which is viewed as 0 in the model (Yu et al., 2001a).

### 3.2.3 Groundwater Recharge

The groundwater recharge is the soil water computed in Section 3.2.2 once it has reached the groundwater table.

In the simulation, for a specific cell and time step, input in the Richard's Equation (Equation 10) is the available overland flow water depth routed from neighboring cells plus the available precipitation. Infiltration and evaporation or evapotranspiration are treated as a source and a sink in the Richard’s Equation (Yu, 2000).

### 3.3 Terrestrial Hydrologic Model - THM

The procedures in this model are: partitioning the rainfall into infiltration and runoff, routing the excess runoff overland to nearby channels, and routing the flow in the channels to the watershed outlet.
Three schemes are implemented in the THM for partitioning rainfall into infiltration and runoff. The three schemes are: Uniform Loss Rate, Soil Conservancy Service (SCS) Curve Number, and Green-Ampt Method. This study adopts the Green-Ampt infiltration scheme as described in the next section. The THM uses a kinematic wave function to simulate the overland flow. It uses the Muskingum-Cunge routine to route the channel flow through DEM-derived channel networks to the watershed outlet.

3.3.1 Green-Ampt Infiltration Scheme

Compared to other infiltration-runoff schemes implemented in the HMS, Green-Ampt scheme is more physically based. The phrase "physically based" indicates that the governing equation is an analytical solution to the Richard's Equation (Equation 10). The Green-Ampt infiltration equation can be written as:

\[
 f_p = K(1 + \frac{SM}{F})
\]

where \( f_p \) is the infiltration capacity, \( K \) is the average hydraulic conductivity in the wetted zone, \( S \) is the difference in average capillary pressure before and after wetting, \( M \) is the difference in average soil moisture before and after wetting, and \( F \) is the cumulative infiltration (Mein and Larson, 1973). With this equation, one can partition rainfall into infiltration and runoff.

Three parameters are involved in the generation of runoff: effective rainfall intensity \( I \), saturated hydraulic conductivity \( K_s \), and infiltration capacity \( f_p \). The rainfall intensity is calculated through measured precipitation. The saturated hydraulic conductivity is assigned through reference to soil type and land use/land cover data. The infiltration capacity is estimated based on the saturated hydraulic conductivity, capillary pressure, soil moisture, and cumulative infiltration. For each time step, when \( I < K_s \),
all rainfall is infiltrated. When \( K_r < I < f_r \), water starts to pond on the surface. When \( K_r < f_r < I \), runoff is generated (Yu et al., 1999a):

\[
R = I - f_r
\]

where \( R \) is the generated runoff.

### 3.3.2 Kinematic Wave Overland Flow Routing

The kinematic function is derived with the assumption that the friction slope is equal to the bed slope and that the gravitational and shear forces are dominant in the Saint Venant Equations (Woolhiser and Liggett, 1967). The function takes the form:

\[
\frac{\partial Y}{\partial t} + \frac{\partial q}{\partial x} = 0
\]

and

\[
q = \alpha h^n
\]

where \( Y \) is the flow depth, \( t \) is time, \( q \) is the flow per unit width, \( x \) is the distance, and \( \alpha \) and \( n \) are flow geometry and surface roughness parameters, respectively. Using Manning’s resistance law, \( \alpha \) is expressed as:

\[
\alpha = \frac{1.49}{n} S^{0.5}
\]

where \( n \) is the Manning’s roughness factor, \( S \) is the slope, and \( m = 5/3 \) in the model (Johnson and Miller, 1997). To solve Equation 14, each grid cell and each time step are further subdivided into smaller increments. Then the following forward finite-difference scheme is employed:

\[
Y_{i+1,t} = q\Delta t + Y_{i+1,t-1} - c_n \left[ \frac{\Delta t}{\Delta x} \right] \left[ \frac{Y_{i+1,t-1} + Y_{i+1,1-t-1}}{2} \right]^{m-1} \times \left[ Y_{i+1,t-1} - Y_{i+1,t-2} \right]
\]
where $\Delta t$ and $\Delta x$ are the subdivisions of time and distance respectively. Precautions are needed to perform the computation. First, to guarantee the accuracy and stability of the kinematic wave routine, the following relationship is maintained in the model:

$$C \Delta t = \Delta x$$

where $C$ is the kinematic wave velocity over a cell. If the instability is anticipated, then the conservative form of the finite difference equation is used. Second, the kinematic wave assumptions remain valid only at slopes around 10%. With shallower slopes, hydrostatic forces may become important (Johnson and Miller, 1997). With steeper slopes, pressure forces may become important. Both will bring error into the solution.

The kinematic wave travel time is estimated using:

$$t_i = \frac{n^{3/5} L_{oa}^{1/5}}{i^{2/5} S_0^{3/10}}$$

where $t_i$ is the travel time, $n$ is the Manning’s roughness coefficient, $L_{oa}$ is the length of the overland flow path, $i$ is the rainfall intensity, and $S_0$ is the bed slope (Johnson and Miller, 1997).

### 3.3.3 Muskingum-Cunge Channel Routing

The Muskingum-Cunge method (Cunge, 1969) assumes no lateral inflow in a certain river reach and storage is equal to the difference between inflow and outflow:

$$I - O = \frac{\Delta S}{\Delta t}$$

where $I$ is the inflow, $O$ is the outflow, $t$ is time, and $S$ is the storage. $S$ is expressed as:

$$S = K[XI + (1 - X)O]$$
where $K$ is the travel time through the reach, $X$ is a weighting factor. If considering a lateral flow, $Q_{in}$, the differential equation of a diffusive transport is written as:

$$\frac{\partial Q}{\partial t} + c \frac{\partial Q}{\partial x} = \mu \frac{\partial^2 Q}{\partial x^2} + c Q_{in}$$

Where $Q$ is the channel flow, $c$ is the wave celerity, $Q_{in}$ is the lateral inflow, $x$ is the distance, and:

$$\mu = \frac{Q}{2BS_0}$$

where B is the width of the channel bottom. Solving Equation 22 yields the Muskingum-Cunge equations as (Miller and Cunge, 1975):

$$O_2 = C_1 I_1 + C_2 I_2 + C_3 O_1 + C_4 Q_{in}$$

where:

$$C_1 = \frac{\Delta t - 2KKX}{2K(1 - X) + \Delta t}$$

$$C_2 = \frac{\Delta t + 2KKX}{2K(1 - X) + \Delta t}$$

$$C_3 = \frac{2K(1 - X) - \Delta t}{2K(1 - X) + \Delta t}$$

$$C_4 = \frac{2\Delta t}{2K(1 - X) + \Delta t}$$

The coefficients of $K$ and $X$ is expressed as:

$$K = \frac{\Delta x}{c}$$

and,\

$$X = \frac{1}{2} \left( 1 - \frac{Q}{c \Delta x BS_0} \right)$$
respectively (Cunge, 1969).

In the model, the channel properties such as depth, width, velocity, and cross-sectional area are estimated using empirical relationships. With simulated flow rate, the flow velocity can be expressed:

\[ V = aQ^b \]

where \( V \) is the flow velocity, \( Q \) is the simulated flow rate, and \( a \) and \( b \) are coefficients that are watershed specific and need to be calibrated (Leopold et al., 1964).

Computations are performed for each grid cell and each time step. The outflow of each cell is input into the downstream cell along the DEM-derived drainage and channel networks. Eventually the flow is routed to the outlet cell of the watershed.

3.4 Groundwater Hydrologic Model - GHM

Groundwater movement is simulated through solving the second-order partial differential equations in the GHM:

\[
\frac{\partial}{\partial x} (T \frac{\partial h}{\partial x}) + \frac{\partial}{\partial y} (T \frac{\partial h}{\partial y}) = S \frac{\partial h}{\partial t} + Q_{nw} \]

where \( h \) is the hydraulic head, \( S \) is storativity, \( T \) is the transmissivity, \( t \) is time, and \( Q_{nw} \) is the net groundwater withdrawal rate, including groundwater recharge from infiltration, evaporation of shallow groundwater, withdrawal of groundwater from wells, and possible induced infiltration of groundwater from the stream network (Domenico and Schwartz, 1990). This equation is solved using an iterative alternating direction implicit method (Yu, 1997).

In the simulations, the material below each cell in the vertical direction is subdivided into a set of layers of thickness \( \Delta z \). The simulation domain is formed through combining
the cell dimensions in the x and y directions. The lateral no-flow boundary is assumed to be the basin boundary (topographic divides). This assumption is valid in the WE-38 watershed (Pionke and Urban, 1985). The bottom boundary is also assumed to be a no-flow boundary.

3.5 Channel-Groundwater Interaction Model - CGI

The CGI simulates the interactions between the channel flow and groundwater flow along DEM-derived channel networks through solving the Darcy's Equation (Yu and Schwartz, 1998). It is assumed that a layer of low permeability separates the groundwater system and the channel system at each stream cell.

The direction of the flow is determined through the relative level of groundwater and the channel stage. If the groundwater level (from GHM) is higher than the stream level (from THM), water flows to the channel, and vice versa. When the groundwater level is lower than the channel bed and there is no flow in the channel, there is no flow in either way. Because there are no real measurements on the permeability rates of the channels are available, the rates are treated as a calibration parameter (Yu, 2000).
CHAPTER 4

STUDY AREA AND DATA SETS

This chapter provides information about the study area, data sets, and processing procedures. Distributed data sets are essential for the distributed hydrologic modeling. The characteristics of the study area are also important to a simulation. A model is always designed for a specific climate region because the characteristics of precipitation and underlying surface are quite different for different climate regions. These differences lead to the difference in runoff generation mechanisms.

4.1 Study Area

The watershed WE-38 is a small watershed with a catchment area of 7.29 km². It is located 40 km north of Harrisburg, Pennsylvania along the East Mahantago Creek, in the Susquehanna River Basin (Figure 2). The stream network originates in the northern ridges of high elevation and outlets to the southern valleys of low elevation. Land use/land cover consists of about 57% cropland, 35% forest, and 8% permanent pasture. Vegetation is typically mature forests on the northern ridges, and pastures, agricultural crops in the middle and southern areas. No urban areas or industries occur within the watershed. Rotations of corn-oats-hay are common agricultural practices within the watershed.
Pionke and Urban (1985) ascribed the watershed a temperate and humid climate. Annual average precipitation is approximately 1090 mm and annual streamflow approximately 460 mm, of which 60-80% is groundwater baseflow (Gburek et al., 1986; Pionke et al., 1996). Annual evaporation loss is 479 mm. During most times of the year, potential evapotranspiration is less than precipitation. Groundwater and subsurface recharge occurs during the late fall, winter, and spring months. Recharge mainly comes from rainfall, not snow melt. Groundwater movement is coincident with surface water, both are from north to south.

Geographically the area belongs to the Appalachian Valley and Ridge physiographic province. Elevation ranges from 240 to 460 m (Pionke and Urban, 1985). In the southern part of the watershed, the major bedrock unit is the Late Devonian Trimmers Rock Formation. In the northern part, the major bedrock unit is the Late Devonian-Early Mississippian Catskill Formation. The Trimmers Rock Formation is mainly shale and crops out in near-horizontal strata at the watershed outlet. The stratal dip increases to the north to a maximum of 22°. Overlying the Trimmers Rock Formation is the Catskill Formation that consists of inter-bedded shale, siltstone, and sandstone which crops out from the middle to the northern divide of the watershed. These strata dip 22° at the mid-watershed and dip increases to 30° at the northern divide. Grain size coarsens progressively from south to north (Gburek and Urban, 1990).

The soil depth in the watershed ranges from 1 to 2 meters. Gburek and Urban (1985) found that below the soil, a 3-10 meters thick shallow weathered and fractured layer exists, which forms the transition zone between surface soil and bedrock. The groundwater table in the dormant season exists both in the soil zone and the transition zone. The soils on the ridge tops are highly permeable; nearly all rainfall will infiltrate.
The fine-textured soils adjacent to the streams have relatively low permeability, forms groundwater discharge zones in the dormant months.

4.2 DEM and Preprocessing

The original DEM of the study area was derived from aerial stereophotography photographs obtained by Photo Sciences, Inc. on April 21, 1994. The DEM has a horizontal resolution of 15 meters. The DEM was resampled to resolutions of 25 m, 50 m, 100 m, 150 m, and 200 m for model use. Statistics of each derived DEM are shown in Table 1.

Table 1. Statistics of the DEMs

<table>
<thead>
<tr>
<th></th>
<th>15 meters</th>
<th>25 meters</th>
<th>50 meters</th>
<th>100 meters</th>
<th>150 meters</th>
<th>200 meters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Data Columns</td>
<td>210</td>
<td>124</td>
<td>53</td>
<td>31</td>
<td>20</td>
<td>17</td>
</tr>
<tr>
<td>Data rows</td>
<td>220</td>
<td>132</td>
<td>66</td>
<td>33</td>
<td>22</td>
<td>16</td>
</tr>
<tr>
<td>Number of Grids</td>
<td>32372</td>
<td>11673</td>
<td>2920</td>
<td>749</td>
<td>322</td>
<td>195</td>
</tr>
<tr>
<td>Area (km²)</td>
<td>7.28</td>
<td>7.29</td>
<td>7.30</td>
<td>7.49</td>
<td>7.24</td>
<td>7.80</td>
</tr>
<tr>
<td>Area Correction Factor</td>
<td>1.001</td>
<td>1.000</td>
<td>0.998</td>
<td>0.973</td>
<td>1.007</td>
<td>0.935</td>
</tr>
</tbody>
</table>

The resampling procedure causes a small change in the watershed area in each derived DEM. Although the changes are not significant, the area must be corrected because the changes will introduce error in the model operations. Therefore the area correction factors are calculated and applied in the simulations.

Hydrologic properties of a basin are defined by the geological and topographical features of the ground surface. Among the many hydrologic features, the drainage
Figure 2. The DEM-derived WE-38 watershed (after Yu et al., 2000). Numbers inside a circle are locations of groundwater wells. Contours are in meters. Solid circles are rain gauge locations.
system is the most important one. The drainage system can be divided into land surface and drainage network. Before rainfall generated runoff reaches the stream network, it is defined as the overland flow. After runoff reaches the stream network, it is referred to as the channel flow. The overland flow and channel flow are significantly different in terms of flow rate, evaporation, erosion, and infiltration behaviors as well as the shallow groundwater-channel interaction. So it is critical that the drainage system is accurately represented in the distributed hydrologic modeling.

The process for developing the drainage system starts from an original DEM. First, topographical imperfections such as small sinks (e.g., small ponds) and peaks will be removed because sinks and peaks will hamper the flow development. Then the flow directions are calculated for each cell by comparing the elevation of the cell to elevations of surrounding cells. According to the flow direction of each cell, the number of cells from which water will flow into is aggregated. A threshold value (contributing area) is used to determine which cell is a channel cell and which cell is a land cell. Then the drainage network and watershed boundary can be delineated (Yu et al., 2001b).

The process is carried out using ARC/INFO functions. An interactive command system in ARC/INFO called "GRID" contains useful functions such as "FILL", "FLOWDIRECTION", "SNAPPOUR", "WATERSHED", "CON", and "GRIDFLOAT". The procedure "FILL" is used to remove sinks and peaks. "FLOWDIRECTION" is used to determine flow directions for each cell. "WATERSHED" is used to delineate drainage networks and watershed boundary. "SNAPPOUR" is used to determine the watershed outlet. "CON" is used to delineate stream networks. The products from these procedures are several data layers. These data layers are in the format of ARC/INFO grids. The function "GRIDFLOAT" is used to transform the grid format into binary storage format.
The data in the binary storage format can then be used to drive the HMS. Figure 2 shows the elevation and stream networks of the WE-38 watershed derived from the original DEM.

4.3 Land Use/Land Cover

An 1:7500 scale aerial photograph that covers the study area was obtained from the 1990 NASA Multi-sensor Airborne Campaign (MAC-HYDRO). Land use field boundaries are digitized from this photograph. Information on the specific land use types was obtained from interviews with farmers (Yu et al., 2000). Major crop types are corn, alfalfa, soybeans, and wheat. Traditional practices include strip farming and crop rotations. Because no records of such information are available, crop type distribution is expressed using areally averaged parameters. Originally eighteen land use types were identified in the study area. The eighteen land use types are reclassified into six types because some types have similar hydrologic properties (Figure 3). Information on crop types is used to describe the ability of each crop type to remove water through roots from the soil column. Information on crop growth stages is used to describe the variation in soil moisture utilization at various depths. In Figure 3, blank areas occur where data are not available. A mixed type was assigned for such areas.

4.4 Soil Types

Nine soil types were identified within the watershed. In this study, the nine soil types are reclassified into five soil categories according to the hydrologic properties (Figure 4). In the rainfall-runoff partitioning scheme, the saturated hydraulic conductivity and the
average capillary suction head are estimated through cross-referencing the soil textures and land use/land cover data.

4.5 Precipitation

In this study, 5-minute interval meteorological data were collected from four gauging stations within the watershed. The stations are marked as Stations 1, 2, 3, and 4 where Station 1 is located in the far north and Stations 2, 3, and 4 form an array from west to east in the southern part of the watershed (Figure 2). The watershed discharges at the southernmost boundary where Station 3 is the nearest gauging station. Two storms have been selected for the simulations. One occurred on July 23, 1997 (referred to as Storm 1) and another on August 26, 1997 (Storm 2). Each storm lasted 48 intervals of 5 minutes. Detailed data measured at each station are shown in Figures 5 and 6.
Figure 3. Land use / land cover of the WE-38 watershed (From Yu et al., 2000).
Figure 4. Soil map of the WE-38 watershed (From Yu et al., 2000).
Figure 5. Precipitation of Storm 1 observed at 4 stations.
Figure 6. Precipitation of Storm 2 observed at 4 stations.
CHAPTER 5

DISTRIBUTION OF PRECIPITATION

Spatial and temporal distributions of precipitation events are complex. The distributions depend on the type of the storm. Detailed study of the distributions is critical to capture the spatial and temporal characteristics of the storms in simulations.

5.1 Spatial and Temporal Distributions

Spatial and temporal distributions are examined using differences and similarities in the statistics of the storms among the gauging stations. Statistics of the storms include total volume, time averages and variances, and positions of the median. The formulations for these variables are:

\[ P_{\text{total}} = \sum_{i=1}^{n} P_i \]  \hspace{1cm} (33)

\[ \bar{P} = \frac{1}{n} \sum_{i=1}^{n} P_i \]  \hspace{1cm} (34)

\[ (S_P)^2 = \frac{1}{n} \sum_{i=1}^{n} (P_i - \bar{P})^2 \]  \hspace{1cm} (35)

\[ C_i = \frac{S_P}{\bar{P}} \]  \hspace{1cm} (36)
where $P_{\text{total}}$ is the total volume of precipitation of a station, $n$ is the number of time steps. $i$ is the time step variable. $P_i$ is the precipitation at time step $i$. $\bar{P}$ is the average precipitation. $(S_p)^2$ is the variance. $C_i$ is the coefficient of variation. $X_{\text{med}}$ is the position where the median value is located in the time series.

Time series statistics are calculated and shown in Tables 2 and 3.

Table 2. Statistical characteristics of Storm 1.

<table>
<thead>
<tr>
<th></th>
<th>Station 1</th>
<th>Station 2</th>
<th>Station 3</th>
<th>Station 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total Precipitation (mm)</td>
<td>26.4</td>
<td>29.2</td>
<td>28.4</td>
<td>29.3</td>
</tr>
<tr>
<td>Average Precipitation (mm)</td>
<td>0.55</td>
<td>0.61</td>
<td>0.59</td>
<td>0.61</td>
</tr>
<tr>
<td>Variance (mm$^2$)</td>
<td>0.0257</td>
<td>0.0305</td>
<td>0.0304</td>
<td>0.0306</td>
</tr>
<tr>
<td>Coefficient of Variation</td>
<td>0.291</td>
<td>0.286</td>
<td>0.295</td>
<td>0.287</td>
</tr>
<tr>
<td>Median Position (out of 48 intervals)</td>
<td>26</td>
<td>25</td>
<td>24</td>
<td>24</td>
</tr>
</tbody>
</table>

The differences among the total amounts and average values of the four gauging stations for Storm 1 are significantly less than those for Storm 2, which means that Storm 1 is more evenly distributed in space than Storm 2. Larger values of the coefficients of variation for Storm 2 in each station show that Storm 1 is more evenly distributed over time. The median positions calculated from the four stations for each storm demonstrate that rainfall for Storm 1 was more concentrated in the temporal center while rainfall for
Storm 2 was concentrated earlier than the temporal center. The detailed differences between Storms 1 and 2 definitely affect the geometry of the corresponding hydrographs for each storm.

Table 3. Statistical characteristics of Storm 2.

<table>
<thead>
<tr>
<th></th>
<th>Station 1</th>
<th>Station 2</th>
<th>Station 3</th>
<th>Station 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total Precipitation</td>
<td>25.2</td>
<td>29.2</td>
<td>32.8</td>
<td>34.7</td>
</tr>
<tr>
<td>(mm)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Average Precipitation</td>
<td>0.52</td>
<td>0.61</td>
<td>0.68</td>
<td>0.72</td>
</tr>
<tr>
<td>(mm)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Variance (mm^2)</td>
<td>0.0683</td>
<td>0.1186</td>
<td>0.1205</td>
<td>0.1368</td>
</tr>
<tr>
<td>Coefficient of Variation</td>
<td>0.502</td>
<td>0.564</td>
<td>0.510</td>
<td>0.514</td>
</tr>
<tr>
<td>Median Position (out of 48 intervals)</td>
<td>26</td>
<td>24</td>
<td>22</td>
<td>22</td>
</tr>
</tbody>
</table>

A time series analysis shows that during the storms, rainfall at the stations varies significantly. Figures 7 and 8 show the relationships between every pair of stations. For Storm 1, the largest correlation coefficient is 0.793, which happened between Stations 2 and 3. The smallest is 0.573, which is between Stations 1 and 3. In Storm 2, the closest relationship is 0.738, which exists between Stations 2 and 4. The least relationship is 0.496, which is between Stations 2 and 3. Together with the total volume of each station, we can see that these two storms are centered storms. Storm 2 is centered around Station 4. Storm 1 is more evenly distributed in space than Storm 2.

5.2 Spatial Interpolation

In order to obtain a good fit between simulated and observed responses, spatial variations of the rainfall need to be fully considered in the input. Without dependable
measurements of precipitation for each grid cell, we rely on rainfall data measured at points (four point measurements in the WE-38 watershed). Spatial interpolation techniques were used to assign rainfall for each grid cell. In the WE-38 watershed, each of the four gauge stations contributes values assigned to grid cells. Let $w_{i,x,y}$ represent the weight a grid cell $(x, y)$ received from Station $i$. Then for each time step, the rainfall at cell $(x, y)$ is:

$$P_{x,y} = \sum_{i=1}^{4} w_{i,x,y} P_i$$

(38)

where $P_i$ is the measured rainfall for Station $i$ in the same time step. Numerous methods may be used to estimate $w_{i,x,y}$. In this study, I chose the “inverse distance” method to interpolate the rainfall data, in which the weight can be expressed as:

$$w_{i,x,y} = \frac{d(x,y,i)^{-1}}{\sum_{j=1}^{4} d(x,y,j)^{-1}}$$

(39)

where $d(x,y,i)$ stands for the distance from cell $(x, y)$ to Station $i$ (Dingman, 1994).
Figure 7. Correlation coefficients of precipitation among stations for Storm 1. Units of precipitation are millimeters.
Figure 8. Correlation coefficients of precipitation among stations for Storm 2. Units of precipitation are millimeters.
CHAPTER 6

DISTRIBUTION OF HYDRAULIC CONDUCTIVITY

Hydraulic conductivity is proved to be highly variable in space but much less variable in time (Russo and Bresler, 1980; Mantoglou and Gelhar, 1987; Gburek and Urban, 1990). Because no direct measurements are available throughout the watershed, accurately estimating the value at each grid cell becomes impossible. Indirect estimations include pure random method, pure physical method, and a mixture of these two.

6.1 Pure Random Hydraulic Conductivity Field

With some information about the hydraulic conductivity field such as a mean value and the assumption of a distribution function, a random hydraulic conductivity field for the whole basin can be generated (Russo and Bresler, 1980; Sudicky, 1986). This approach has been adopted by many researchers (Dagan, 1979; Neuman and Yakowitz, 1979; Gutjahr, 1984; Yeh et al., 1985; Mantoglou and Gelhar, 1987).

6.2 Pure Physically Based Approach

The phrase "pure physically based" means that the hydraulic conductivity is obtained by implementing measurements or slug tests. However, because it is impossible to
measure the whole watershed, measurements can only be made in the laboratory or in parts of the field (Russo and Bresler, 1980; Sudicky, 1986; Gburek and Urban, 1990).

Slug tests show that the hydraulic conductivity is highly variable in the WE-38 watershed. It is even variable within the same geologic units (Gburek and Urban, 1990). Comparatively, the hydraulic conductivity is higher both in the soil and the transition zone than in the bedrock. Hydraulic conductivity is also greater at high elevations compared to low elevations, especially in the ridge areas relative to in the valley floors.

6.3 Physically Based & Random Approaches

In this study, a random distribution was adopted within a physical framework to assign hydraulic conductivity values for grid cells. By overlapping soil data and land use/land cover data layers, sub-zones of different soil type and land use/land cover can be obtained. Traditionally one tends to assign a value of a hydraulic parameter to each of those subzones, which will provide a uniform value for grid cells within individual subzones. In other words, each subzone is considered homogenous. However, a natural land surface tends to be heterogeneous because of small creeks, seams, small ponds or peaks, variations in elevation, compactness of the soil, foliage and roots, etc., which cause variations in hydraulic parameters. The fact that variations occur in hydraulic parameters within each subzone is comparable with variations in hydraulic parameters in samples from each textual class reported by Cosby et al. (1984). Accordingly, I propose adding variations into the hydraulic parameter in each subzone. In this study I adopt the mean and standard deviation values of a log-normal distribution reported by Cosby et al. (1984) for each subzone. Then the random standard normal deviates are used to generate
realizations and the hydraulic conductivity $K$ is made to vary in the grid cells within each subzone. Frequently $K$ is viewed as log-normally distributed (Neuman, 1982):

$$ Y = \ln K $$

(40)

$Y$ is a normally distributed variable. It can be calculated using:

$$ Y = \bar{Y} + S_Y Z $$

(41)

where $\bar{Y}$ is the mean value, $S_Y$ is the standard deviation for $Y$, and $Z$ is the random standard normal deviates obtained from established tables (Haan, 1977). Once a realization is generated for $Y$ using Equation 41, the corresponding $K$ can be solved for through Equation 40. In this study, the logarithm with a base of 10 was used. Therefore the calculated $Y$ was increased by a constant of 2.303. The process of assigning $K$ values can be viewed as a kind of downscaling of the spatial variability of the hydraulic conductivity, $K$, with a stochastic approach. Because it is impossible to measure parameters between grid cells within a traditionally viewed homogenous subzone, a stochastic approach is always adopted and considered reasonable (e.g., Dagan, 1984; Gelhar, 1986). Theoretically, numerous realizations of $K$ values for the grid cells in one subzone are possible. Optimized realizations in terms of overall effects exist (Yu and Schwartz, 1999). If a cell covers two different soil types, then an areally weighted parameter value is assigned.
CHAPTER 7

CHANGE OF TOPOGRAPHIC FEATURES WITH GRID SPACING

Accurate representation of the topographic features of a basin is the key to the physically based modeling. Large deviation from the actual attributes of a basin surface in a simulation will definitely cause inaccuracy. In this respect, small grid sizes have advantages over large grid sizes because the hydrologic features extracted from DEMs of small grid sizes are more accurate than those from DEMs of large grid sizes. This chapter presents a discussion of topographic changes with grid sizes.

7.1 Statistics

Table 4 and Figure 9 show the changes of statistics of elevation, slope, and aspect with grid sizes in the WE-38 watershed. For elevations, maximum values decrease 7% with grid size changes from 15 m to 200 m. Minimum values increase 11%. As a result, mean values show no major change. Correspondingly, slope changes also occur. The mean values of slope decrease 35% when grid size changes from 15 m to 200 m. The maximum values of slope decrease 36%. These results are consistent with previous studies (Jensen, 1993; Zhang and Montgomery, 1994; Yu et al., 2001b). Changes in elevation and slope in general reflect a smoothing effect on the land surface. It can be expected that in the simulations the computed overland flow velocity will decrease and
the concentration time will be prolonged. A negative trend of the mean value of aspect may be indicative of shortening of the flow path. This occurs because in DEMs of large grid sizes, detailed small-scale variations are neglected and a straightforward flow path will be extracted. The shortening of the flow path will reduce the Muskingum-Cunge channel routing travel time. These data tell us that the simulation is affected by the topographic change.

Table 4. Variation of elevation, slope, and aspect with grid sizes.

<table>
<thead>
<tr>
<th>Grid Size (m)</th>
<th>Elevation (m)</th>
<th>Slope (degree)</th>
<th>Aspect (degree)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean</td>
<td>Minimum</td>
<td>Maximum</td>
</tr>
<tr>
<td>15</td>
<td>288.1</td>
<td>215.1</td>
<td>497.1</td>
</tr>
<tr>
<td>50</td>
<td>288.2</td>
<td>216.1</td>
<td>489.9</td>
</tr>
<tr>
<td>100</td>
<td>289.6</td>
<td>215.2</td>
<td>487.5</td>
</tr>
<tr>
<td>150</td>
<td>288.2</td>
<td>227.1</td>
<td>491.8</td>
</tr>
<tr>
<td>200</td>
<td>290.8</td>
<td>239.0</td>
<td>462.0</td>
</tr>
</tbody>
</table>

7.2 Elevation

Figure 10 shows the accumulative distributions of elevation with grid sizes. Particularly the minor changes in the low elevation part. Deviations are obvious for elevations between 350 and 450 meters for the 100 and 200 m grid sizes, indicating that part of the range is increased in the large size grids.
Figure 9. Changes in statistics of elevation, slope, and aspect by grid sizes.

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Figure 10. Accumulative distribution of elevation.
7.3 Slope

Figure 11 shows the accumulative distribution of slope. The entire range of slope is affected, however, the most prominent change happens around 10 degrees. With the increase of grid size, the slope of the watershed decreases. For the 200 m grid size, the steep part with slope greater than 10° is not comparable to the rest of the lines. This difference may occur because of the small number of grid cells in the 200 m grid. Therefore, large deviations and distortions can be expected for large size grids.

7.4 Aspect

From the histogram of aspect (Figure 12), deviations from the histogram of the original DEM (15 m grid) can be seen. Increasing grid sizes makes the histogram less uniform. The predominant aspects of 135° and 315° are indicative of the the southeast and northwest directions of the flow path of the stream network.

7.5 Selection of Grid Size

From the accumulative distributions of both slope and elevation (Figures 10 and 11), it can be seen that curves for grid sizes larger than 150 meters are stepped and distorted. The reason for this distortion is that fewer grid cells occur in grids with size greater than 150 meters. Smaller grid sizes produce smoother accumulative distributions. Apparently grids of sizes larger than 150 meters produce discontinuous accumulative distributions for both slope and elevation. Grid sizes smaller than 100 meters have good qualities for the hydrologic simulations in the WE-38 watershed.
Figure 11. Accumulative distribution of slope.
Figure 12. Histogram of topographic aspect.
This chapter presents the model calibration and applications of the HMS in the WE-38 watershed. Grid size effects on the simulations of various hydrologic processes are examined in detail. The procedures for developing scale factor functions are also presented in this chapter.

The HMS has been successfully applied in the study of surface-water and groundwater interactions in the Big Darby Creek watershed in Ohio (Yu and Schwartz. 1998), in the study of the spatial distribution of water balance in the WE-38 watershed (Yu et al., 2000), as a nested submodel in the Pennsylvania State-National Center for Atmospheric Research Mesoscale Meteorological Model for studying the river-basin response to atmospheric forcing (Yu, 2000), the river-basin response to single-storm events (Yu et al., 1999b), and the subgrid hydrologic processes (Yu et al., 1999a; Yu, 2000), and in evaluating the spatial-temporal variation of soil moisture in the Susquehanna River Basin (Yu et al., 2001b). Most of the applications were made at a resolution of 1 km, however, in this study, the modeling is focused on much smaller scales.
8.1 Model Calibration

Model calibration follows schemes developed in Yu and Schwartz (1999). In this study, parameters that are needed for calibration include: kinematic wave travel time \( k \) and geometry factor \( x \) in Muskingum-Cunge routine scheme and the hydraulic conductivity of the channel bed in the Channel-Groundwater Interaction Model (CGI). The calibration target is the observed streamflow. A plausible range of the value for each parameter is given based on the physical meaning of the parameter and the type of watershed. Then the range is divided into several subdivisions. Various values of each parameter are used to run the HMS and obtain a good fit between observed and simulated streamflow. The first search is called a “coarse search”. If necessary, consecutive searches can be conducted by further dividing the obtained parameter sets into finer intervals and repeating the searching procedures until the desired fitness is obtained.

The actual model operation involves a two stage pre-simulation automatic calibration: a balancing period and a transient period (Yu et al., 1999a; Yu et al., 1999b; Yu et al., 2001b). The purpose of the balancing period is to let the models “forget” the initial condition, which is not always known. In the SHM, a balance is needed among the precipitation, evapotranspiration, runoff, soil-moisture contents, and subsurface drainage. The balance is reached when the standard deviation of soil-moisture contents is within 3% of the mean value. The balancing period depends on the selected balancing criteria, the temporal and spatial distributions of the driving hydrologic and meteorological forces, the given initial conditions, and seasonal factors. In the GHM and CGI models, the balancing is done among rain recharge, evaporation, groundwater storage, and baseflow. For example, if the baseflow is given five different sets of initial groundwater levels for each grid cell, the simulated hydrograph will converge to an equilibrium condition after
40 days (Yu et al., 1999a). If the initial groundwater level is higher than actual groundwater level, the groundwater flow system will release water to the streams to reach equilibrium. On the other hand, if the initial groundwater level is lower than actual groundwater level, the groundwater system will gain water from the streams.

In this study, the balancing period is designed to be 150 days considering the model operating time and for accommodation of the requirements of parameters in most models. A transient period follows the balancing period that further optimizes some of the parameters such as soil and groundwater hydraulic conductivity and storativity (Yu et al., 1999a). The transient period focuses on the overall water balance. The overland flow and channel flow timings can be calibrated in the transient period.

### 8.2 Parameterizing Precipitation Schemes

Effects of small variations in precipitation in space and time are studied using interpolated and averaged rainfall. The interpolating and averaging schemes used to assign rainfall values are discussed in Chapter 5. Figure 13 shows simulations with spatially averaged and interpolated rainfall for Storm 1 and Figure 14 represents Storm 2. Use of the averaged rainfall means that the spatial variation of precipitation is not considered in the simulation, whereas the interpolated rainfall is the case in which the variation of precipitation in space is reflected. Significant differences between the simulations of the spatially averaged and interpolated precipitation are observed in both events. In Figures 13 and 14, the dashed lines represent simulations with the spatially interpolated rainfall, the thin solid lines represent simulations with the spatially averaged rainfalls, and the heavy solid lines represent observed streamflow at the watershed outlet.
Figure 13. Spatial variations in precipitation for Storm 1. Upper bar chart is the spatially averaged rainfall, in millimeters.
Figure 14. Spatial variations in precipitation for Storm 2. Upper bar chart is the spatially averaged rainfall, in millimeters.
It can be seen that the dashed lines fit nicely with the observed, whereas the thin solid lines show poor fits.

Different behavior of the two storms can be expected. As discussed in Chapter 5, Storm 1 is more evenly distributed over the basin, whereas Storm 2 is more centered around Station 4. The simulated results using the spatially averaged rainfall reflect the difference. As expected, the deviation of the simulations with the spatially averaged rainfall from the observed hydrograph is much more pronounced for Storm 2 than Storm 1 (Figures 13 and 14). This difference clearly reflects that neglecting the spatial variability of the precipitation causes the simulated hydrograph to deviate much more for the storms with the high spatial variability. The total volume of the streamflow is also largely underestimated.

Figures 15 and 16 show the simulations with temporally averaged rainfall in time intervals of 5, 15, 30, and 60 minutes for the two storms. A large sampling interval of precipitation increases the deviations from a good simulation both on peak discharge and flood volume of the runoff. These deviations result from the averaging procedure for large time intervals, which reduce the precipitation intensity. In the Green-Ampt rainfall-runoff partitioning scheme, when the precipitation intensity is smaller than the hydraulic conductivity, no runoff will be produced. For large sampling intervals, both peak discharge and total volume of the simulated hydrograph have been reduced. It seems that a 15-minute or less sampling interval gives the most optimized results in peak values and total amounts for the two studied storms in terms of accuracy and required computing time.
Figure 15. The effect of variations in sampling interval of precipitation on the hydrograph for Storm 1. Upper bar chart is the spatially averaged rainfall, in millimeters.
Figure 16. The effect of variations in sampling interval of precipitation on the hydrograph for storm 2. Upper bar chart is the spatially averaged rainfall, in millimeters.
8.3 Parameterizing Schemes of Hydraulic Conductivity

The importance of the spatial variability in hydraulic conductivity, $K$, is examined in this section. Two different schemes are used to assign the $K$ value. One uses a uniform $K$ for each sub-zone and the other assigns a log-normally distributed $K$ as discussed in Chapter 6. The simulated results of the runoff are shown in Figures 17 and 18. In the uniform $K$ scheme, the hydrologic responses are significantly underestimated both on the peak discharge and the flood volume, which is the area under the curve. In addition, the rising points for the hydrographs are largely delayed. In contrast, the scheme of the spatially varied $K$ provides much better results in comparison to the observed hydrographs in both storm events. This better match occurs because small-scale heterogeneities are neglected in the uniform hydraulic conductivity scheme for each sub-zone and no runoff is generated unless the precipitation intensity is greater than the $K$ value in the scheme. In reality, numerous small spots in a sub-zone have a hydraulic conductivity that is smaller than the averaged value so even much smaller rainfall intensity will produce runoff. The comparisons suggest that the spatial variation of conductivity, $K$, plays an important role in the simulations of hydrologic processes.

8.4 Grid Size Effects

The calibrated HMS is used to study the effect of grid size on the hydrologic simulations within the WE-38 watershed. Keeping other parameters and approaches of the data sets unchanged, a range of grid sizes (50 m to 200 m) is selected for the simulations. Simulated results are shown in Figures 19 and 20. As expected, large size grids produce large deviations in the representation of the landscape, and thus, produce
Figure 17. Comparison of different schemes estimating hydraulic parameters for Storm 1. Upper bar chart is the spatially averaged rainfall, in millimeters.
Figure 18. Comparison of different schemes estimating hydraulic parameters for Storm 2. Upper bar chart is the spatially averaged rainfall, in millimeters.
large deviations in the hydrographs of the simulated processes. Both the peak discharge and the total amount of the streamflow at the watershed outlet are underestimated. A good indication of the quality of the small size grids is that the overall geometry of the simulated hydrograph mimics the observed one. The peak value, total amount, and the rising point are closest to the observed hydrograph. On the other hand, too small a grid size may not improve simulations. Figures 19 and 20 show the cases where the hydrograph of the 50 m grid is calibrated to the measured one. The peaks of the simulated hydrographs for 100 m and 200 m grids came earlier than observed. However, if the 200 m grid is calibrated to the hydrograph of the measured one, then both peaks for the hydrographs of 100 m and 50 m grids will be delayed and the peak for the 50 m grid hydrograph will be delayed further. Large volumes of total runoff for the 100 m and 50 m grids will also be expected. So the grid resolution of 100 m is suggested as a proper grid size for the WE-38 watershed in terms of accuracy and efficiency.

8.5 Scaling Functions and Applications

Figures 21 and 22 show the changes of the simulated total runoff over spatial and temporal extensions for the two storms. The runoffs are normalized to the observed one. Both cases show that total runoff decreases with the increase of spatial scales. The larger the scales are, the more the total runoff is reduced. These findings are consistent with previous studies (Finnerty et al., 1997; Koren et al., 1999).

Figures 21 and 22, the slope of the continuous lines changes sharply at the 100 m scale. Therefore simulated results with grid spacing larger than 100 meters deteriorate significantly. This deterioration may be more supporting evidence for proposing the 100
Figure 19. Grid spacing effects on the hydrologic simulation for Storm 1. Upper bar chart is the spatially averaged rainfall, in millimeters.
Figure 20. Grid spacing effects on the hydrologic simulation for Storm 2. Upper bar chart is the spatially averaged rainfall, in millimeters.
m grid for the simulations. Another observation is that the scale factor between simulations is based on different spatial and temporal scales. Scale functions can be built based on these findings. Let $M$ be this scale factor. Then:

$$r = M \times R$$

where $r$ is the simulated total runoff with one set of spatial and temporal scales and $R$ is the simulated total runoff with another set of scales. For the cases of Storms 1 and 2, scale factors are calculated and shown in Tables 5 and 6. These factors are scaled to the observed total runoff. For the purpose of prediction, we need to further divide $M$ into two parts, $M_s$ and $M_t$, where $M_s$ is the spatial scale factor and $M_t$ is the temporal scale factor. Assume a second order polynomial function for both spatial and temporal scales:

Table 5. Scale factors for Storm 1.

<table>
<thead>
<tr>
<th></th>
<th>25 meters</th>
<th>50 meters</th>
<th>75 meters</th>
<th>100 meters</th>
<th>150 meters</th>
<th>200 meters</th>
</tr>
</thead>
<tbody>
<tr>
<td>5 minutes</td>
<td>1.18</td>
<td>1.17</td>
<td>1.14</td>
<td>1.11</td>
<td>1.03</td>
<td>0.96</td>
</tr>
<tr>
<td>15 minutes</td>
<td>1.16</td>
<td>1.15</td>
<td>1.13</td>
<td>1.09</td>
<td>1.00</td>
<td>0.91</td>
</tr>
<tr>
<td>30 minutes</td>
<td>1.13</td>
<td>1.12</td>
<td>1.11</td>
<td>1.06</td>
<td>0.98</td>
<td>0.85</td>
</tr>
<tr>
<td>60 minutes</td>
<td>1.04</td>
<td>1.03</td>
<td>1.02</td>
<td>0.99</td>
<td>0.92</td>
<td>0.76</td>
</tr>
</tbody>
</table>

Table 6. Scale factors for Storm 2.

<table>
<thead>
<tr>
<th></th>
<th>25 meters</th>
<th>50 meters</th>
<th>75 meters</th>
<th>100 meters</th>
<th>150 meters</th>
<th>200 meters</th>
</tr>
</thead>
<tbody>
<tr>
<td>5 minutes</td>
<td>1.36</td>
<td>1.32</td>
<td>1.27</td>
<td>1.21</td>
<td>1.08</td>
<td>0.81</td>
</tr>
<tr>
<td>15 minutes</td>
<td>1.33</td>
<td>1.31</td>
<td>1.26</td>
<td>1.19</td>
<td>1.06</td>
<td>0.78</td>
</tr>
<tr>
<td>30 minutes</td>
<td>1.31</td>
<td>1.28</td>
<td>1.21</td>
<td>1.16</td>
<td>1.03</td>
<td>0.69</td>
</tr>
<tr>
<td>60 minutes</td>
<td>1.26</td>
<td>1.24</td>
<td>1.19</td>
<td>1.12</td>
<td>1.00</td>
<td>0.61</td>
</tr>
</tbody>
</table>

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Figure 21. Spatial and temporal changes of total runoff for Storm I.
Figure 22. Spatial and temporal changes of total runoff for Storm 2.
where \( m_i \) is a function of spatial scales and \( m_r \) is a function of temporal scales. Then use data in Tables 5 and 6 to calibrate these functions. For example, a temporal scale function of 5 minutes and a spatial scale function of 25 meters for Storm 1 can be built up. The calibrated functions are:

\[
M_i = -0.0331 \times m_i^2 + 0.095 \times m_i + 1.0274
\]

and:

\[
M_r = -0.0004 \times m_r^2 - 0.007 \times m_r + 1.0973
\]

where:

\[
m_i = 1 + \log_2 \left( \frac{\text{length}}{25} \right)
\]

and:

\[
m_r = \frac{\text{time}}{5}
\]

where \( \text{length} \) and \( \text{time} \) are spatial and temporal readings in meters and minutes, respectively. With these functions, one can predict what a scale factor would be at a certain scale. If one wants to know the factor when spatial scale comes to 400 meters at a temporal scale of 5 minutes, then one uses Equation 47 with length as 400. A value of 16 meters for \( m_i \) results. By substituting 16 for \( m_i \) in Equation 45 and one finds the value of \( M_i \) is 0.67. So the prediction is that at scales 400 meters and 5 minutes, the simulated total runoff will be 0.67 times of the observed. Similarly, at a spatial scale of 25 meters, one can calculate the scale factor of 2 hours. From Equation 48 one determines that \( m_r \) is 24 minutes. Substitute 24 for \( m_r \) in Equation 46 and one finds that \( M_r \) is 0.69. The
resulting prediction is that at scales of 25 meters and 120 minutes, the simulated total runoff will be 0.69 times of the observed.

In fact, with interpolation, scale functions can be determined for any spatial and temporal scales. It must be noted that these scale functions are extremely case sensitive, however the method can be universal.
CHAPTER 9

TOWARD SCALE FACTOR PREDICTION

In order to develop proper scale functions for scale factor predictions, it is necessary to classify storms into categories. Each category is unique in terms of weather conditions, total volume, previous effective rainfall, etc. It is also necessary to choose a proper number of storms in each category and run the simulations. Once the scale factors are obtained and plotted together, the method proposed in this study can be applied to derive scale functions. This chapter is aimed at illustrating such an approach.

9.1 Storm Classification

Establishing the criteria for storm classification is rather arbitrary and complicated. First, the contributing elements must be determined. Second, the criteria for subdividing each contributing element into different groups need to be established. For example, one cannot easily define preferable weather conditions that will generate a certain type or amount of rainfall. Another example is to decide the number of subdivisions within the volume factor. Tremendous effort is involved in selecting criteria and handling each of the criteria in order to obtain reasonable patterns. Even when the criteria are determined and the groups are subdivided, one still needs a reasonable number of storms in each
category to run the simulations and obtain scale factors. So abundant data sets are necessary for performing the required task.

In this study, without detailed analysis of how the scale factors behave with the impact of contributing criteria, a third storm is selected with the assumption that the three storms are within the same category. A general scale function can be developed for this category based on the simulated scale factors.

9.2 A Third Storm

The third storm occurred on August 10, 1998 (referred to as Storm 3) in the WE-38 watershed. Precipitation measurements from the four gauge stations are shown in Figure 23. Average rainfall over the four gauge stations is 16.8 mm. The storm lasted 29 intervals of 5 minutes. Relationships of the data for every pair of stations are shown in Figure 24. Statistics of Storm 3 are calculated and listed in Table 7.

Table 7. Statistical characteristics of Storm 3.

<table>
<thead>
<tr>
<th></th>
<th>Station 1</th>
<th>Station 2</th>
<th>Station 3</th>
<th>Station 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total Precipitation (mm)</td>
<td>17.8</td>
<td>16.4</td>
<td>18.0</td>
<td>15.0</td>
</tr>
<tr>
<td>Average Precipitation (mm)</td>
<td>0.61</td>
<td>0.57</td>
<td>0.62</td>
<td>0.52</td>
</tr>
<tr>
<td>Variance (mm²)</td>
<td>0.9366</td>
<td>0.9264</td>
<td>1.055</td>
<td>0.8967</td>
</tr>
<tr>
<td>Coefficient of Variation</td>
<td>1.579</td>
<td>1.703</td>
<td>1.651</td>
<td>1.832</td>
</tr>
<tr>
<td>Median Position (out of 29 intervals)</td>
<td>13</td>
<td>13</td>
<td>12</td>
<td>20</td>
</tr>
</tbody>
</table>

Compared to Storms 1 and 2, Storm 3 shows a number of differences. First it can be seen from Table 7 that Storm 3 is a much smaller storm in terms of total volume. Both
Figure 23. Precipitation of Storm 3 observed at 4 stations.
$\rho_{12} = 0.598$  

$\rho_{13} = 0.659$

$\rho_{14} = 0.02$  

$\rho_{23} = 0.832$  

$\rho_{24} = 0.0006$  

$\rho_{34} = 0.036$

Figure 24. Correlation coefficients of precipitation among stations for Storm 3. Units of precipitation are millimeters.
Storms 1 and 2 have about twice the volume of Storm 3. Second, the variances and coefficients of variation of the four stations for Storm 3 are much larger than those for Storms 1 and 2, indicating that Storm 3 varies temporally to a higher degree than both Storms 1 and 2. In Figure 24, the closest relationship is between Station 2 and 3 where the correlation coefficient is 0.832. The correlation coefficients between Stations 1 and 2 and Stations 1 and 3 are smaller, but not significantly small. However, the correlation coefficients of Station 4 and the other three stations for Storm 3 are much smaller than the correlation coefficients among the other three stations. This feature indicates that Storm 3 is highly variable in space when compared with the other two storms. Finally, the median position of Station 4 for Storm 3 is farther away from the temporal center as compared with the other three stations. The change in the median position indicates that the storm center for Storm 3 moved from somewhere close to Stations 1, 2, and 3 to a point near Station 4 during the storm period.

The differences between Storm 3 and Storms 1 and 2 may provide additional information that can be helpful in developing general scale factor functions.

9.3 Synthetic Scale Factor Functions

Model calibration and input data processing described in Chapter 8 for Storms 1 and 2 were used here for Storm 3. Simulations were performed at the temporal scales of 5, 15, 30, and 60 minutes. Simulated hydrographs for the 50 m, 100 m, and 200 m spatial resolutions and 5 minutes temporal resolution are shown in Figure 25 where the simulated hydrograph for the 50 m grid is calibrated to the observed streamflow. The results are comparable with the simulated results for both Storms 1 and 2 (Figures 19 and
20). It can be seen from Figure 25 that the peaks for larger grid sizes come earlier than the observed peak. The peak discharges and total volumes for small grid sizes are larger than large grid sizes. It is noted that the first peak in the precipitation does not correspond with a peak in the measured or simulated hydrographs. The reason for this lack is that the previous effective rainfall for Storm 3 is quite limited in the WE-38 watershed. No storm events were recorded for the previous 30 days in the four gauge stations. So the total amount of rainfall for the first peak of the precipitation satisfied the initial abstraction. As shown in Figure 25, the simulated hydrographs captured the fact nicely.

In order to develop synthetic scale factor functions for a storm category, it is necessary to have scale factors from a number of storm events in the category. The word "synthetic" here means averaging. It is expected that the averaged scale factors from a storm category reflect the general trend of that category. So the functions developed based on the averaged scale factors can be used for prediction. The following is an example for developing one such function. Comparisons of the synthetic function with functions developed from single storm events were also made in terms of prediction accuracy.

The 5 minute temporal scale is used for developing the scale factor functions. Calculated scale factors for the three storms at different spatial scales are listed in Table 8 and plotted in Figure 26. Similar to the approaches for Storms 1 and 2 (Tables 5 and 6), the scale factors for Storm 3 are normalized to the observed total runoff. Average scale factors for the three storms are also calculated and listed in Table 8.
Figure 25. Grid spacing effects on the hydrologic simulation for Storm 3. Upper bar chart is the spatially averaged rainfall, in millimeters.
Figure 26. Scale factor changes for the three storms.
In Figure 26, scale factors for the three storms form different lines. Scale factors for each storm can be used to develop a unique function, indicating that they are case specific. On the other hand, all the points form a consistent band, strengthening the trend of scale factors derived from Storms 1 and 2. The averaged scale factors derived from the three storms representing the general trend of the three storms. A synthetic scale factor function was developed based on the average scale factors at the temporal scale of 5 minutes. Using the same method as described in Chapter 8, the synthetic function is fitted as:

### Table 8. Scale factors at the temporal scale of 5 minutes.

<table>
<thead>
<tr>
<th></th>
<th>25 meters</th>
<th>50 meters</th>
<th>75 meters</th>
<th>100 meters</th>
<th>150 meters</th>
<th>200 meters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Storm 1</td>
<td>1.18</td>
<td>1.17</td>
<td>1.14</td>
<td>1.11</td>
<td>1.03</td>
<td>0.96</td>
</tr>
<tr>
<td>Storm 2</td>
<td>1.36</td>
<td>1.32</td>
<td>1.27</td>
<td>1.21</td>
<td>1.08</td>
<td>0.81</td>
</tr>
<tr>
<td>Storm 3</td>
<td>1.45</td>
<td>1.37</td>
<td>1.23</td>
<td>1.09</td>
<td>0.97</td>
<td>0.79</td>
</tr>
<tr>
<td>Average</td>
<td>1.33</td>
<td>1.29</td>
<td>1.21</td>
<td>1.14</td>
<td>1.03</td>
<td>0.85</td>
</tr>
</tbody>
</table>

\[
M_i = -0.0140 \times m_i^2 + 0.0042 \times m_i + 1.3387 \\
\]

A synthetic function is better than a function developed from a single storm in a statistical manner. For example, to predict the scale factor of Storm 3 at the spatial scale of 800 m, using different functions will produce different results. Assume a function developed using information directly from Storm 3 produces the most accurate result. The assumption is valid because the scale factors are case specific. For the sake of comparison, scale functions at the temporal scale of 5 minutes for the three storms were developed:
\[
M_i = -0.0089 \times m_i^2 + 0.0168 \times m_i + 1.1742 \tag{50}
\]
\[
M_i = -0.0261 \times m_i^2 + 0.0808 \times m_i + 1.2884 \tag{51}
\]
\[
M_i = -0.0069 \times m_i^2 - 0.0849 \times m_i + 1.5534 \tag{52}
\]

Equations 50-52 are developed based on information from Storms 1, 2, and 3 respectively. The calculated factors are shown in Table 9. Taking the scale factor obtained from Storm 3 as the standard value, then the differences are also listed in Table 9. It can be seen that prediction errors for the functions developed from Storms 1 and 2 are 1.49 and 1.57 times greater than the synthetic function, respectively. This comparison indicates the advantage of using a synthetic function. Table 10 listed the predicted total runoff for both 400 m and 800 m grid. For the simulations using the 800 m grid, the synthetic function performs better than both functions developed from Storms 1 and 2. For the simulations using the 400 m grid, the synthetic function performs better than the function developed from Storm 2, but poorer than the function developed from Storm 1. The reason for such a behavior of the synthetic function can be explained by storm similarity. It is expected that the synthetic function is better for capturing features of storms similar to Storms 1 and 2. However, as discussed above, Storm 3 shows a
number of differences from both Storms 1 and 2. So the synthetic function does not perform well in predicting Storm 3. The performance of the synthetic function proves the importance for storm classification.

Table 10. Accuracy of total runoff prediction.

<table>
<thead>
<tr>
<th></th>
<th>Equation 49</th>
<th>Equation 50</th>
<th>Equation 51</th>
<th>Equation 52</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total Runoff for 400 m Grid (m$^3$)</td>
<td>1437</td>
<td>1396</td>
<td>1516</td>
<td>1401</td>
</tr>
<tr>
<td>Absolute Errors from 1401 (m$^3$)</td>
<td>36</td>
<td>5</td>
<td>115</td>
<td>-</td>
</tr>
<tr>
<td>Total Runoff for 800 m Grid (m$^3$)</td>
<td>1282</td>
<td>1315</td>
<td>1310</td>
<td>1215</td>
</tr>
<tr>
<td>Absolute Errors from 1215 (m$^3$)</td>
<td>66</td>
<td>100</td>
<td>95</td>
<td>-</td>
</tr>
</tbody>
</table>
CHAPTER 10

CONCLUSIONS AND FUTURE STUDIES

This chapter provides conclusions from this study and perspectives for future studies as well. The author believes that many research topics exist with regard to scale problems in hydrology. This study can be a basis for the next phase of research tasks.

10.1 Conclusions

From above chapters it is suggested that the small-scale distributions of hydrologic properties in a watershed are critical for the hydrologic simulations of various hydrologic processes. In this study, the primary concern is the effects of topographic features and detailed distributions of precipitation and hydraulic conductivity.

It is understood that when the grid scale increases from 15 meters to 200 meters, the topographic features of the WE-38 watershed show a number of changes. First, maximum values of the elevations decrease 7%, minimum elevation increases 11%. The resultant mean values of elevation show no significant change. Maximum slope decreases 36% while averaged slope decreases 35%. Generally the topographic change with the increase of the grid scale reflects a smoothing effect on the land surface.

It is also understood that the selected storms in the WE-38 vary significantly in both space and time. To account for the variations, an interpolation scheme is applied.
Simulated results demonstrated a significant difference between the simulations with the spatially interpolated rainfall and the simulations with spatially averaged rainfall for the selected storms. Because the degree of spatial variations in Storm 1 is lower than Storm 2, the differences in the results between averaged and interpolated rainfall for Storm 1 is much less than Storm 2. Simulations with different time intervals show that smaller sampling intervals perform better than large sampling intervals both on the peak discharge and the total volume. This difference occurs because large time intervals reduce the precipitation intensity, which is important in the Green-Ampt rainfall-runoff partitioning scheme. A decrease in precipitation intensity reduces both the chance to produce runoff and the amount of the runoff produced.

Using the available soil type and land use/land cover data, the effect of spatial variability in hydraulic conductivity on the simulations is studied. In this study, sub-zones of different soil and land use/land cover type were first worked out. Then a log-normal distribution was selected for assigning hydraulic conductivity (K) values for each grid cell in each sub-zone. Simulations with uniform K values and log-normally distributed K show that the hydrologic responses are largely different. The peak discharge and total volume are significantly underestimated with the simulation of using uniform K values. In addition, the rising points for the simulated hydrographs are largely delayed. The better match of the log-normally distributed K indicates that the log-normally distributed scheme provides better representation of the spatial variation of the hydraulic conductivity than the uniform scheme. It is suggested that a log-normally distributed hydraulic conductivity K is adopted in the study instead of a uniform scheme.
Effects of different grid scales on the hydrologic simulations are studied. First, a small grid scale was applied to calibrate the model. Then a range of grid scales was used to test the model. Simulations show that compared to small grid scales, large grid scales tend to underestimate both the simulated peak discharges and total volumes. This may be because the spatial interpolation scheme for the precipitation also has a smoothing effect for large grid scales. The interpolation scheme becomes more and more like the averaging scheme when the grid scale increases. In this way, the overall rainfall intensity is reduced and less runoff is produced. The interpolation scheme deteriorates to the average scheme when the scale size increases to the extent that the whole WE-38 watershed becomes one cell. In addition to the underestimation of both peak discharges and total volumes, the peaks for simulations with large grid scales came earlier. This may be because of the shortening of the flow path for the large grid scales. Simulated flood travels faster with shorter flow paths.

A general expectation is that the smaller the grid size is, the better the results are. However, this study shows that too small a grid size may not improve the simulation. It is believed that the lack of detailed information in the subsurface soil properties and rock formations in the simulations hampered this study. First, a constant soil depth and a similar rock formation throughout the watershed are assumed in this study. Hence constant hydraulic conductivities are adopted both in the Soil Hydrologic Model and Groundwater Hydrologic Model. This will produce deviations in the simulations of soil water and groundwater movement. The same problem happens in the Groundwater-Channel Interaction Model. In the simulations, over 80% of the precipitation infiltrates in the subsurface. So the water movement underground plays an important role. More
accurate representations of subsurface water movement are highly required for the simulations. Second, the surface soil data was digitized from a map of 1:7500. Generalizations were made during the processes. These generalizations could introduce errors in the simulations. The effort of adopting a stochastic approach in the representation of the hydraulic conductivity largely improved the simulation, however, the detailed direct information is highly recommended. Third, as shown in Figures 7 and 8, the precipitations for Storms 1 and 2 are highly variable over time and space. To account for the variations, the spatial interpolation scheme of inverse distance is applied. But one problem with the mathematical function interpolation scheme is the "smoothing effect". High-resolution measurements such as radar precipitation data are suggested for better performance.

After all these uncertainties, it can be concluded that small heterogeneities, both spatial and temporal, are specifically important in simulating hydrologic responses to storms at fine scales. Among the most significant factors, grid spacing, and spatial and temporal distributions of precipitation and hydraulic parameters need to be considered first.

In the study, scale functions have been constructed for predictions. However, it must be understood that these functions are extremely case specific. It is also noted that the assumption for a second-order polynomial function is not mathematically guaranteed. More solid research needs to be done to draw such a conclusion.

Chapter 9 presents a more generic approach for scale factor predictions using a synthetic function. A synthetic function is statistically better than a function developed from a single storm. It is necessary to develop a synthetic function from a number of
storms in a category identified by similar weather condition, total volume, previous effective rainfall, etc.

Results show that the 100-m resolution grid provides a satisfactory control of topographic features of the WE-38 watershed and is appropriate to the level of overall data conditions. Hence, it produced the most adequate results in the simulations. The spatial and temporal distribution of precipitation and the conductivity, $K$, in the selected grid scale are key factors affecting the simulations. Better understanding of distributions of other hydrologic processes in model inputs are also recommended.

10.2 Future Studies

This study mainly addresses small spatial and temporal variability in different hydrologic processes and the effects on hydrologic responses. A small watershed is used to study the effects. It is expected that the same mechanism exists in a much larger watersheds. So next step study will concentrate on applying the methodology in the study to a number of large watersheds and explore the scaling effects. It is also noted that the WE-38 watershed is located in a humid region. Also the scaling effects in arid and semi-arid regions need further studies. On the other hand, studies show that the scaling effects are different for different hydrologic processes (Finnerty et al., 1997; Koren et al., 1999). Future studies should also aim at constructing rigorous scale functions for different hydrologic processes.
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