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Farshad Baloochestani

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doi: <https://doi.org/10.57709/1059583>

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**ESTIMATION OF HYDRAULIC PROPERTIES OF THE SHALLOW AQUIFER  
SYSTEM FOR SELECTED BASINS IN THE BLUE RIDGE AND THE PIEDMONT  
PHYSIOGRAPHIC PROVINCES OF THE SOUTHEASTERN U.S. USING  
STREAMFLOW RECESSION AND BASEFLOW DATA**

**by**

**FARSHAD BALOOCHESTANI**

**Under Direction of Dr. Seth Rose**

**ABSTRACT**

The objectives of this research are to measure the aquifer properties (*S*, *T*, and *K*) of selected watersheds delineated to the U.S. Geological Survey gauging stations using streamflow recession and baseflow data and to describe the relations among the properties of shallow aquifers and the physical properties of the basins, such as slope, regolith type and thickness, and land use type. Geographic Information System (GIS) techniques are utilized to investigate critical physiographic controls on transmissivity and storage coefficients on a regional basis. Moreover, the effect of evapotranspiration on recession index is illustrated. Finally, a detailed quantitative comparison of results for the Piedmont and the Blue Ridge Physiographic Provinces in southeast of the U.S. is provided.

Recession index, annual groundwater recharge, and annual baseflow data were obtained from 44 USGS-gauging stations with drainage areas larger than 2 (mi<sup>2</sup>) and less than 400 (mi<sup>2</sup>). These gauging stations are located in Georgia and North Carolina. Analyses of data focused on

GIS techniques to estimate watershed parameters such as total stream length, drainage density, groundwater slope, and aquifer half-width. The hydraulic diffusivity, transmissivity, and storage coefficient of watersheds were computed using hydrograph techniques and the Olmsted and Hely, and Rorabaugh mathematical models.

Median recession index values for the Blue Ridge and Piedmont Provinces are 87.8 and 74.5 (d/log cycle), respectively. Median areal diffusivity values for the Blue Ridge and Piedmont are 35,000 and 44,200 (ft<sup>2</sup>/d), respectively. Median basin-specific estimates of transmissivity for basins in the Blue Ridge and Piedmont are 150 and 410 (ft<sup>2</sup>/d), respectively. The large values of transmissivity obtained for the Piedmont regolith may be attributed to the thick regolith, low values of basin relief, and voids that develop as a result of fracturing, foliation, weathering, and fractured quartz veins in the saprolite. Median basin-specific estimates of storage coefficient for basins in the Blue Ridge and Piedmont are 0.005 and 0.009, respectively. In general, the results from this study reveal great differences in basin-specific hydraulic parameters of the regolith material within the Piedmont compared to that of the Blue Ridge Physiographic Province.

INDEX WORDS:           Recession index, baseflow, aquifer properties, Blue Ridge and Piedmont, southeastern U.S., groundwater-surface water interaction, master recession curve, shallow aquifer system

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**FARSHAD BALOOCHESTANI**

A Dissertation Submitted in Partial Fulfillment of the Requirements for the Degree of

Doctor of Philosophy

In the College of Art and Sciences

Georgia State University

2008

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Office of Graduate Studies  
College of Art and Sciences  
Georgia State University  
May 2008

## DEDICATION

This dissertation is dedicated to my parents for their love, wisdom, support, high moral standards, and compassion for humanity.

## ACKNOWLEDGEMENTS

I am grateful to my supervisor, Dr. Seth Rose, for his excellent guidance, valuable comments, and kind support. I also thank Dr. W. Crawford Elliott, Dr. Hassan A. Babaie, and Dr. Jeremy E. Diem, whose comments led to important revisions.



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# 1 INTRODUCTION

## 1.1 *Broad Impact of This Study*

The areal transmissivity  $T$ , areal diffusivity  $T/S$ , storage coefficient  $S$ , and recession index  $K$  are fundamental hydraulic parameters that are useful in assessing the groundwater potential of an aquifer particularly in time of drought. These parameters can be used for optimal development and management of water resources in the study area. Despite the importance of the transmissivity and storage coefficients, they have rarely been quantified at regional scale, and the mechanisms controlling them are poorly understood. The traditional approach for determining aquifer properties is the well-test method which can be unreliable at the regional scale because hydraulic properties obtained vary in individual wells in which aquifer tests were conducted.

A clear understanding of shallow aquifer characteristics has a great importance for efficient development and management of surface water and groundwater resources, as well as for reducing the pollution of aquifers and connected surface waters. Surface waters such as rivers, lakes, ponds, reservoirs, wetlands, riparian areas, and estuarine areas are the integral part of ecological systems (EPA's Draft Report, 2003). In addition, these important sources of fresh water are utilized for drinking, recreation, irrigation, livestock, and industrial purposes. According to Alley et al. (1999), groundwater, one of the most important natural resources in the United States, accounts for nearly 40% of the U.S. public water supply, and is the source of much of water used for irrigation. In the view of some experts, groundwater seems to be the only option for the United States' potential future water supply (EPA's Draft Report, 2003). Moreover, it is well known that shallow groundwater discharge constitutes much of the streamflow of rivers in humid climates, thus indicating the close interaction between

groundwater and surface water (Wittenberg and Sivapalan, 1999). Groundwater also seems to play a major role in governing the surface water quality. According to Reay et al. (1992), the misinterpretation of data and error in water quality management strategies is probably a consequence of neglecting shallow groundwater discharge as a nutrient source to streams (Arnold et al., 2000). To illustrate, the Gerhart (1986) reported the complex link between shallow groundwater recharge and the nitrate pollution in Pennsylvania (Arnold et al., 2000). Moreover the importance of shallow groundwater recharge is emphasized by Krulikas and Giese (1995) who argue that groundwater resources are the main water supply for Florida, thus causing the legislature to consider planning of tax incentives to owners of high recharge areas (Arnold et al., 2000).

Population growth and the consequent increase in water use for irrigation, industrial, and public use within the Blue Ridge and the Piedmont Provinces of southeast of the United States have raised concerns about the maintenance of water resources. Understanding of groundwater-streamflow interaction within these areas is required to: estimate the aquifer properties (e.g. storage coefficient and transmissivity), manage groundwater withdrawal and streamflow depletion, design water resources structures, model the rainfall-runoff system, quantify groundwater pollutant loading to streams and rivers, and design and implement stream-habitat protection and restoration programs. The result of this study combined with other studies in the Blue Ridge and Piedmont Physiographic Provinces of the eastern United States may assist the management of water resources by shedding light on the quality and quantity of these resources.

## ***1.2 Proposed Contribution of this Study***

Several studies (Olmsted and Hely, 1962; Hoos, 1990; Rutledge and Mesko, 1996; Nelms et al., 1997) have focused on stream recessions and aquifer properties of the Blue Ridge and

Piedmont Physiographic Provinces of the eastern United States. According to Rutledge and Mesko (1996) the hydraulic diffusivity  $\frac{T}{S}$  of the aquifer can be determined by the following equation:

$$\frac{T}{S} = \frac{0.933a^2}{K} \quad \text{Equation 1}$$

where  $T$  is areal transmissivity ( $\text{ft}^2/\text{d}$ ),  $S$  is storage coefficient (dimensionless),  $a$  is aquifer half-width (mi), and  $K$  is recession index (day). The authors used the initial estimates of aquifer half-width  $a$ , which were obtained by scanning the literature to solve Equation 1 for areal diffusivity. The authors applied a simplistic approach wherein the few results available for storage coefficient  $S$  in the literature were put into Equation 1 to estimate the areal transmissivity. The Rutledge and Mesko (1996) method, however, could be more accurate if the Geographical Information System (GIS) was employed for estimation of the aquifer half-width  $a$  and areal transmissivity  $T$ . This paper applies the Arc/Info 9.2 network analysis procedure for estimates of the aquifer properties of shallow aquifer systems in southeastern of the United States. In other words, this study extends the use of previous methods (Olmsted and Hely, 1962; Hoos, 1990; Rutledge and Mesko, 1996; Nelms et al., 1997) by implementing GIS to increase the accuracy and precision of estimates of the aquifer properties in the study area. Furthermore, GIS and streamflow analyses allow assessment of the aquifer properties on a regional scale.

### **1.3 Purpose and Scope**

Analyzing properties of shallow aquifers by means of streamflow records may shed light on the apparent inconsistency between small to moderate groundwater yield to wells and the high yield to streams. In addition, according to Rutledge and Mesko (1996), the presence of fractured-

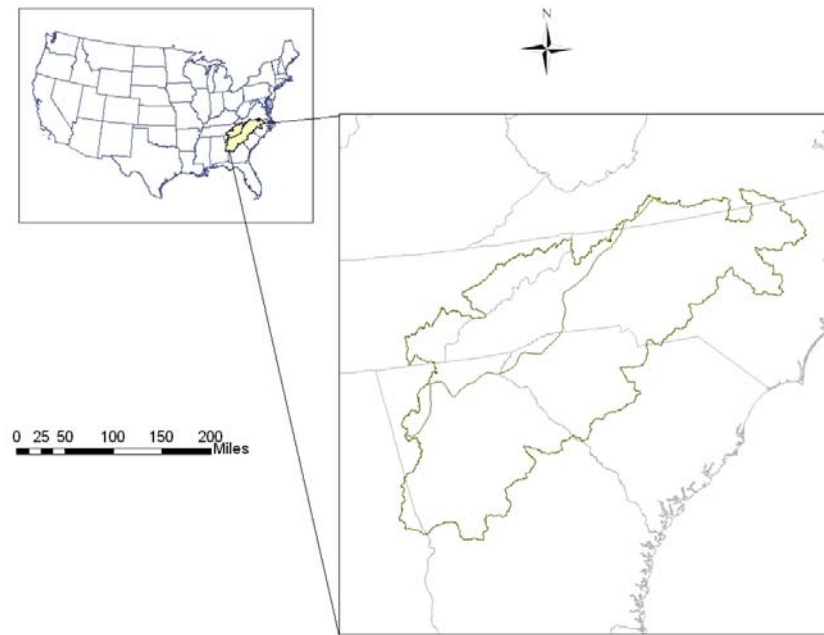
rock flow systems in the study area may add to the benefits of streamflow analysis for determining aquifer properties over traditional aquifer-test methods. Traditional pumping test analysis can be unreliable in these systems because well diameter, length, and pumping rates can be small relative to the scale of fractures. Because the scale of streams is large relative to the scale of most fractures, ground-water flow can be conceptualized by means of continuum mathematics. Hydraulic properties obtained from streamflow analysis are likely to be “average” values for the aquifer system. The principal problem with traditional aquifer-test methods in fractured rocks is the variability between the hydrogeologic conditions at the well site and idealized conditions in the equations used for analyzing test results (Rutledge and Mesko, 1996). Moreover, the traditional approach can be unreliable at regional scale because the hydraulic parameters obtained vary in different wells in which aquifer tests were conducted. The challenge to tradition well test method is the fact that the geologic settings in the region are extremely complex.

The purpose of this study is to define the hydraulic properties of selected basins on the basis of baseflow characteristics. Furthermore, this study describes the relations among the properties of aquifers and the physical properties of the basins, such as relief, soil type, soil thickness, land use type, and curve number map. The study area will be grouped and ranked based on baseflow characteristics and hydraulic properties. Data from published reports and from streamflow records maintained by the USGS are used for these evaluations.

#### ***1.4 Description of Study Area***

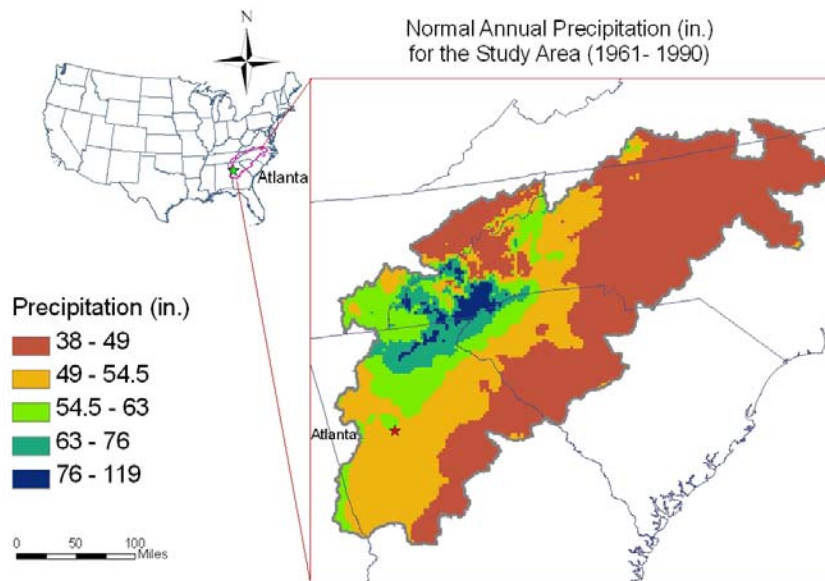
The Blue Ridge and Piedmont Physiographic Provinces of the study area encompass approximately 64,800 (mi<sup>2</sup>) in southeast of the United States (Figure 1-1). The study area is bordered by Valley and Ridge Physiographic Province on the west and by the Coastal Plain

Physiographic Province on the east. The entire study area is underlain by fractured rock aquifers locally covered by regolith consisting of soil, alluvium, and saprolite. Thickness of the regolith ranges from 0 to more than 150 (ft) throughout the study area (Swain and others, 1991).



**Figure 1-1: A generalized map of study area**

The Blue Ridge Physiographic Province encompasses approximately 15,700 (mi<sup>2</sup>) along a narrow northeast-tending belt between the Valley and Ridge and the Piedmont Physiographic Provinces, and consists of a chain of mountains and highlands underlain by metamorphic Proterozoic and Paleozoic rocks.



**Figure 1-2: The normal annual precipitation (in.) pattern for the study area (1961-1990). These data were extracted from the original data obtained from the PRISM site.**

The Piedmont Physiographic Province encompasses approximately 49,100 (mi<sup>2</sup>) along eastern part of the study area, and consists of gently rolling plain underlain by polydeformed and metamorphosed Proterozoic and Paleozoic rocks. A thick mantle of soil and weathered rock that overlies the fractured crystalline bedrock is a characteristic feature of the province (Swain and others, 1991). Large rift basins filled with sedimentary deposits of Triassic age exist in the Piedmont of North and South Carolinas. These basins are characterized by generally lower relief than surrounding Piedmont (Hack, 1989). Thin soils and shallow weathering profiles are characteristic features of the rocks in the Mesozoic basins.

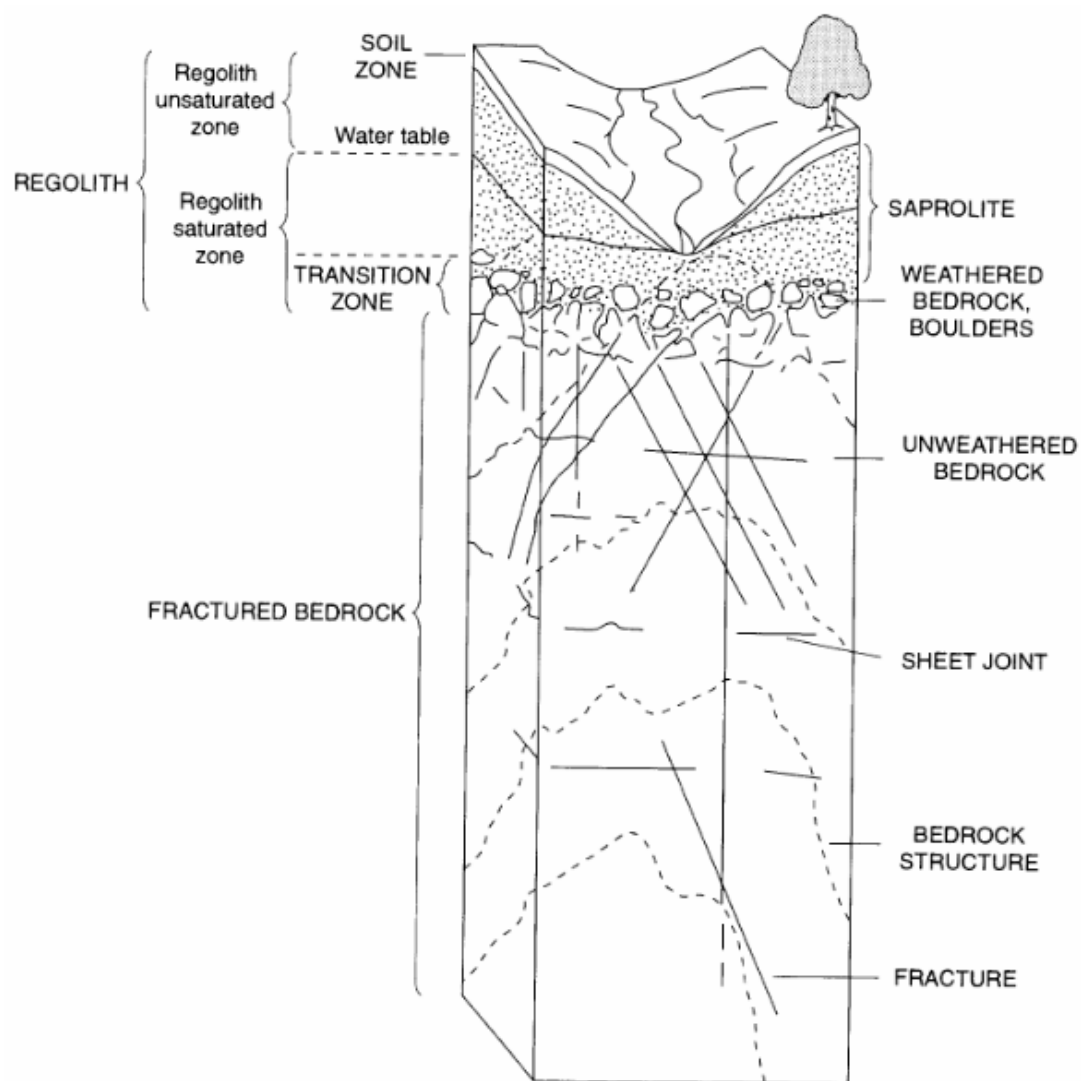
The climate of study area is moderate and can be described as humid-subtropical. The area is characterized by short, mild winters, and hot, humid summers. Precipitation varies with location and elevation but averages 51 (in/yr). In the Blue Ridge and Piedmont Physiographic



Provinces of the study area, the average annual precipitation is 57 (in/yr) and 49 (in/yr), respectively (Figure 1-2).

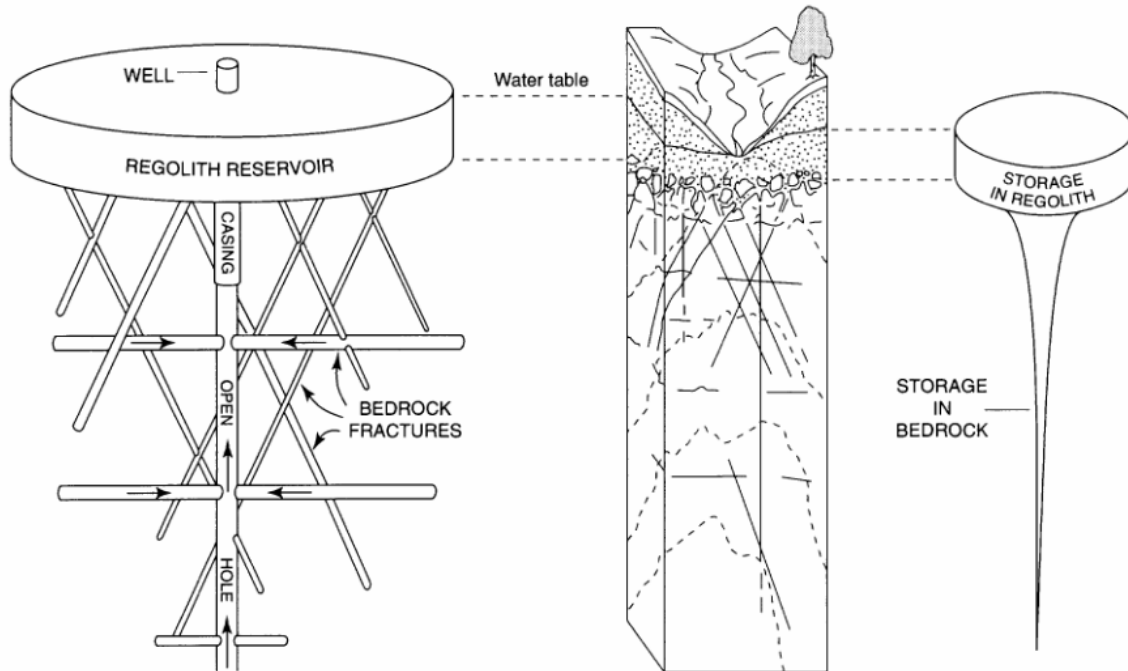
### ***1.5 The Regolith-Fractured Crystalline Rock Aquifer System***

In the study area, the metamorphic and igneous crystalline rocks are mantled by varying thicknesses of the regolith. An idealized sketch of the groundwater system (Figure 1-3) shows the following components of the system: (1) the unsaturated zone in the regolith, which generally contains the organic layers of the surface soil, (2) the saturated zone in the regolith, (3) the lower regolith which contains the transition zone between saprolite and bedrock, and (4) the fractured crystalline bedrock system.



**Figure 1-3: Principal components of the groundwater system in the Piedmont Physiographic Province (from Harned and Daniel, 1992)**

As a general rule, the abundance of fractures and size of fracture openings decrease with depth. At depths approaching 600 (ft) and greater, the pressure of the overlying material, or lithostatic pressure, holds fractures closed, and the fracture porosity can be less than 1 percent (Daniel, 1989). Because of its larger porosity, the regolith functions as a reservoir that slowly feeds water downward into fractures in the bedrock (Figure 1-4). These fractures serve as an intricate interconnected network of pipelines that transmit water either to springs, wetlands, streams, or wells.



**Figure 1-4: The reservoir-pipeline conceptual model of the Piedmont groundwater system and the relative volume of groundwater storage within the system (from Heath, 1984 and Daniel, 1996)**

## **1.6 Groundwater Flow System**

The aquifer in the study area is a two-part system consisting of regolith and fractured crystalline bedrock. The water table typically lies in the saprolite zone. The depth of groundwater flow is difficult to define (LeGrand, 2004). The water from precipitation replenishes the aquifer through the land surfaces higher than adjacent stream valleys. The infiltrated water slowly moves downward through the aeration zone. In the saturation zone, the water moves vertically and laterally and discharges as seepage into areas where saturated zone is near land surface. In the regolith, the groundwater movement is through intergranular flow; however, the relict rock fabric and structure can influence the groundwater movement. In the bedrock, the groundwater movement is controlled by the fractures and schistosity, and the flow paths from recharge to discharges areas commonly are more circuitous than those in the regolith (Daniel and Dahlen, 2002). In the Piedmont where values of basin relief are low, it seems that the local flow systems primarily extend through the shallow regolith aquifer, although some deep local flow systems may develop through the fractured bed rock. In the Blue Ridge the local flow systems may extend towards deeper levels of the fractured bedrock due to the more pronounced relief in the basins of this physiographic province.

## 2 LITERATURE REVIEW

### ***2.1 Computer Simulation Models (Reductionist) Versus Hydrograph/Recession Techniques (Holistic)***

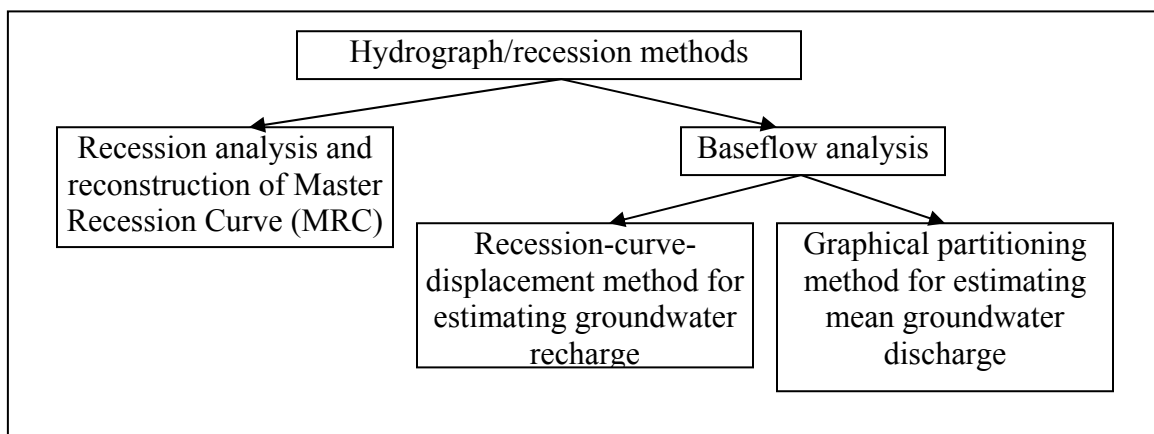
Table 2-1 provides a comparison of the two general types of methods utilized for estimating the groundwater recharge and discharge (groundwater balance): computer simulation/percolation models and hydrograph/recession techniques. In percolation models, rainfall at the surface is measured. This is followed by the estimation of infiltration, redistribution, evapotranspiration, percolation of the residual water through the vadose zone to groundwater table, and discharge of groundwater aquifer to streams as baseflow (Wittenberg and Sivapalan, 1999). Wittenberg and Sivapalan (1999) believe that for the purpose of the water resources assessment of large areas, the percolation models are inappropriate. This is because the lack of accuracy in the measurement of precipitation and the derivation of catchment-average rainfall, together with the errors due to the uncertainty of soil parameters, might result in somewhat unsatisfactory estimations of groundwater recharge and discharge (Wittenberg and Sivapalan, 1999). In addition, an integral and costly part of percolation models is monitoring the movement of water through the vadose zone by tensiometers, tracers, and weighing lysimeters. Due to the high cost of monitoring the water balance, percolation models are preferably used in dry regions (Arnold and Allen., 1999). One common example of these computer models is referred to as Soil and Water Assessment Tool (SWAT), which is a complex and conceptual method with precise parameterization (Arnold and Allen., 1999). Arnold and Allen (1999) have presented a complete description of SWAT model components, and the reader is referred to this paper for further information on the subject.

Table 2-1: A comparison of hydrograph/recession techniques and percolation model.

	Advantages	Disadvantages
<b>Computer simulation models (e.g. percolation models)</b>	<ul style="list-style-type: none"> <li>• According to Arnold et al. (2000), one of the great advantages of percolation model is that it can simulate the management and climate scenarios. Climate scenarios involve such variables as changes in precipitation, temperature, radiation, humidity, and <math>CO_2</math>; whereas, management scenarios may monitor parameters such as cropping systems, tillage, irrigation, fertilization, and reservoir management (Arnold et al., 2000). The authors also state that models may account for nutrient and pesticide simulations.</li> <li>• Due to the high cost, models are preferably utilized for small areas (e.g. sub-basins) and drier climates (Arnold et al., 2000)</li> </ul>	<ul style="list-style-type: none"> <li>• The need for many input data (e.g. weather, soil, land use, geology, and topography), makes the models difficult to apply.</li> <li>• Wittenberg and Sivapalan (1999) believe that for the purpose of the water resources assessment of large areas, the percolation models are inappropriate. This is because, the lack of accuracy in the measurement of precipitation and the derivation of catchment-average rainfall, together with the errors due to the uncertainty of soil parameters might result in somewhat unsatisfactory estimations of groundwater recharge and discharge (Wittenberg and Sivapalan, 1999).</li> <li>• According to Arnold et al. (2000), employment of percolation model is costly.</li> </ul>
<b>Hydrograph/recession techniques</b>	<ul style="list-style-type: none"> <li>• Hydrograph techniques are easy to apply.</li> <li>• These techniques are inexpensive because they entirely rely on streamflow data available free of charge on the United State Geological Survey (USGS) web site.</li> <li>• Hydrograph/recession techniques provide reliable estimates of groundwater balance, if they are utilized in large areas with sub-humid to humid climates (Arnold et al., 2000)</li> </ul>	<ul style="list-style-type: none"> <li>• When these methods are applied to small areas, results are unsatisfactory due to the fact that field conditions commonly do not meet the assumptions needed to use these techniques (Mau and Winter, 1997).</li> </ul>

Hydrograph/recession techniques used in this study are an alternative approach of estimating groundwater balance. These techniques utilize streamflow data to separate the total streamflow into baseflow and quickflow (overland flow), where baseflow is identified as the groundwater discharge into a stream. In hydrograph techniques, the procedural difficulties associated with small-scale phenomenological equations (e.g. percolation model) are avoided, thus rendering them effective for estimating the groundwater balance of large areas (Wittenberg and Sivapalan, 1999). As a result of effectiveness of hydrograph/recession techniques, many investigators have utilized them to compute the groundwater balance of various basins (e.g. Chapman and Maxwell, 1996; Rutledge and Mesko, 1996; Nelms, et al., 1997; Chapman, 1999, and others).

As shown in Figure 2-1, hydrograph/recession methods used in the literature are of two types: (1) the recession analysis and the construction of master recession curve (MRC), and (2) the baseflow analysis. Baseflow analysis itself is divided into two techniques: the recession-curve-displacement method for estimating groundwater recharge, and the graphical partitioning method for computing mean groundwater discharge. The stream hydrograph and a few of the hydrograph/recession techniques are described in the following sections.



**Figure 2-1: Various types of hydrograph/recession techniques**

## 2.2 Stream Hydrographs and Recession Analysis

Figure 2-2 shows a stream hydrograph which is the plot of streamflow discharge of a river at a single location against time. A stream hydrograph showing a recharge event can be broken into four components: overland flow, baseflow, direct precipitation, and interflow.

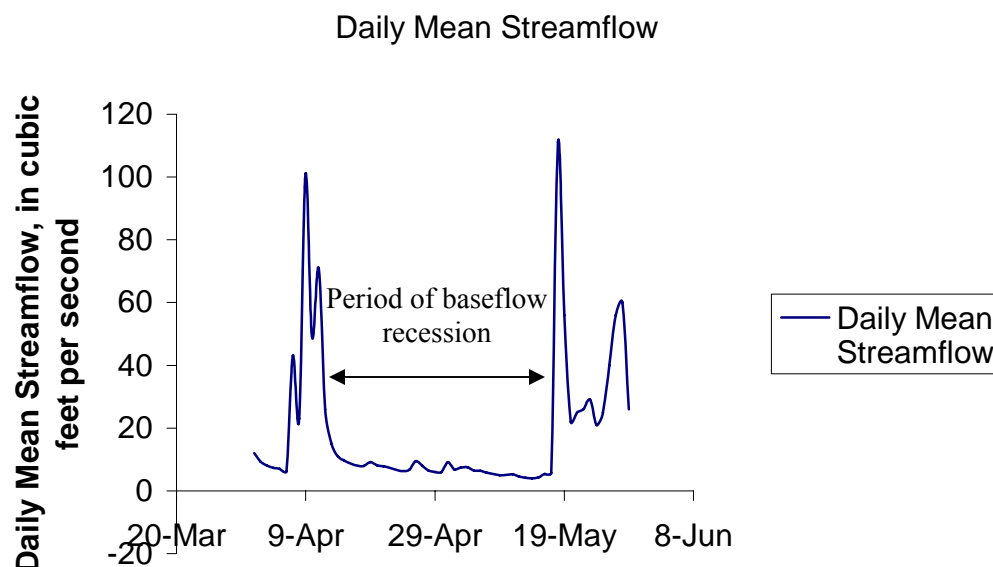
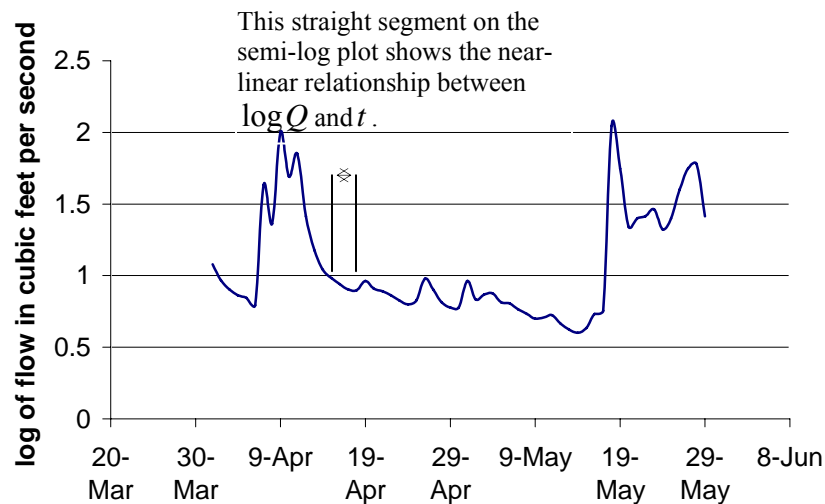


Figure 2-2: May 2003 arithmetic hydrograph for Holiday Creek near Andersonville, VA

That part of a stream hydrograph showing the gradual depletion of streamflow discharge during periods of little or no precipitation is referred to as a recession curve, or baseflow recession hydrograph (Figure 2-2). This recession curve is composed of surface flow, interflow, and baseflow; and that, it is only the lower part of a recession curve that represents the baseflow (groundwater discharge). In other words, the lower segment of a recession curve is the plot of groundwater discharge against time. As shown in Figure 2-2, if there was no groundwater recharge, the groundwater discharge (baseflow) feeding the stream would become zero due to the drop in water table; a lower water table results in a lower baseflow rate. The baseflow will



continue approaching zero until the groundwater reservoir is replenished, thus raising the water table (or until the hydraulic gradient becomes zero).



**Figure 2-3: April-May 2003 semi-log hydrograph for Holiday Creek near Andersonville, VA**

The baseflow recession for a drainage basin is a function of watershed characteristics (e.g. overall topography, drainage pattern, soils, and geology) as well as aquifer characteristics such as storage coefficient and transmissivity. When the effects of groundwater abstraction and evapotranspiration are negligible, the baseflow recession provides valuable information about the storage properties of a groundwater reservoir. According to Tallaksen (1995), many investigators have derived average values of recession constant for various physiographical provinces; and these recession constants are expressed as indices somewhat representing the storage properties of hydrological regions. There are many ways by which the baseflow recession curve can be expressed; however, the most frequently used analytical expressions can be classified as linear and nonlinear models.

### 2.2.1 Linear model

The arithmetic-scale graph of groundwater depletion against time is a curve expressed by a simple exponential equation in the following form:

$$Q_t = Q_0 e^{-t/K} \text{ Equation 2}$$

where  $Q_t$  is the streamflow at some time after the recession started ( $\text{ft}^3/\text{s}$ ),  $Q_0$  is the flow at the start of recession ( $\text{ft}^3/\text{s}$ ),  $t$  is the time since the start of recession (days), and  $K$  (days/log cycle) is a constant (Boussinesq, 1877; Horton, 1933; Maillet, 1905). This equation is referred to as a linear model because the semi-logarithmic plot of  $\log Q$  versus  $t$  generates a straight line defined by the following expression:

$$t = k_1 \times \log Q + k_2 \text{ Equation 3}$$

where  $t$  is time in days,  $\log Q$  is the logarithm of flow in ( $\text{ft}^3/\text{s}$ ), and  $k_1$  and  $k_2$  are coefficients that are determined by linear regression (Rutledge, 2004). In this equation the absolute value of  $k_1$  is accounted for by recession constant; which is the reciprocal negative value of the slope.

Additionally, the previously mentioned exponential function implies a linear relationship between groundwater discharge and aquifer storage properties. Table 2-2 presents the parameters of the exponential equation and their dimensions, as well as alternative definitions of recession constant in the literature.

**Table 2-2: Parameters of the recession equation, as well as alternative definitions of recession constant ( $K$ ).**

$Q_t = Q_0 e^{-t/K}$		
$Q_t$	Flow (groundwater discharge) at some time $t$ after the recession started	(L <sup>3</sup> /t; ft <sup>3</sup> /s)
$Q_0$	Flow at $t=0$ (when the recession started)	(L <sup>3</sup> /t; ft <sup>3</sup> /s)
$t$	Time since the start of recession	( $t$ ; days)
$K$	<ul style="list-style-type: none"> <li>• is referred to as recession constant or recession index</li> <li>• is the retention constant that represents the storage lag-time</li> <li>• is the storage delay factor (Singh and Stall, 1971)</li> <li>• represents average response time in storage (Wittenberg and Sivapalan, 1999)</li> <li>• is the time (number of days) which is required for groundwater discharge to drop though one log cycle after the recession started</li> <li>• is the reciprocal negative value of the slope of semi-log plot in which <math>\log Q_t</math> against <math>t</math> yields a straight line</li> </ul>	( $t / \log \text{Cycle}$ ; days/log cycle)

### 2.2.2 Non-linear model

According to Wittenberg (1999), in most cases the semi-log plot of  $\log Q$  vs.  $t$  still remains concave indicating a non-linear storage-discharge relationship in which a decrease in streamflow is likely to cause an increase in  $K$ . For further information on non-linearity of storage-outflow relationship one may refer to papers of Wittenberg (1999), Werner and Sundquist (1951), and Mishra et al. (2003).

Table 2-3 provides a comparison between the linear and non-linear models of storage-discharge relationship.

**Table 2-3: A comparison of two linear and non-linear recession models in the baseflow analysis**

<i>Linear model</i>	<ul style="list-style-type: none"> <li>• In the linear model, the arithmetic graph of streamflow versus time is expressed by the following simple exponential equation:  <math display="block">Q_t = Q_0 e^{-t/K}</math> where <math>Q_t</math> (ft<sup>3</sup>/s) is the discharge at some time after the start of recession, <math>Q_0</math> (ft<sup>3</sup>/s) is the discharge at the start of recession, <math>t</math> (days) is time, and <math>K</math> (days/log cycle) is the recession constant.</li> <li>• According to Chapman (1999), for shallow aquifers, when the stream bed and underlying impermeable bedrock do not intersect each other, the linear model becomes applicable.</li> <li>• The linear model can be fitted satisfactorily to short recessions (e.g. less than 10 days). In humid climates, due to the frequent interruption of recession period by precipitation during rainy season, the linear model is likely to be applicable in such regions.</li> </ul>
<i>Non-linear model</i>	<ul style="list-style-type: none"> <li>• According to Wittenberg (1999), in the non-linear model, the arithmetic graph of streamflow versus time can be expressed by the following equation:  <math display="block">Q_t = Q_0 \left[ 1 + \frac{(1-b)Q_0^{1-b}}{ab} t \right]^{1/(b-1)}</math> where <math>Q_t</math> and <math>Q_0</math> (m<sup>3</sup>/s) are discharge at time <math>t</math> and discharge at the <math>t=0</math>, respectively, and <math>a</math> and <math>b</math> are constants. Wittenberg (1999) suggests a standard value of 0.5 for <math>b</math>, while he believes that <math>a</math> may contain information about porosity and hydraulic conductivity of aquifers.</li> <li>• According to Chapman (1999), for very shallow aquifers where stream bed and underlying bedrock intersect one another, the non-linear model is more applicable.</li> <li>• Wittenberg (1999) and Chapman (1999) believe that the non-linear model can be fitted satisfactorily to long recession periods (recessions more than 10 days). Dry regions usually favor such long recession periods.</li> <li>• Due to the higher flexibility of non-linear model, for the same basin, the baseflow estimated by linear model is lower compared to the non-linear model that gives higher value (Chapman, 1999).</li> </ul>

## 2.3 Recession Selection Algorithms

In practice, the recession curve is composed of three segments representing surface flow, interflow, and baseflow, respectively; therefore, one of the difficulties is the determination of start and end of a baseflow segment whose reciprocal of the slope is referred to as recession

constant. According to the linear model, the recession constant should be independent of time and flow. However in practice, the semi-log graph of  $\log Q$  against time is a curve composed of three segments; only the lower part of this curve is a straight line representing purely groundwater discharge. As a result, for the purpose of estimating the recession constant or inverse slope of the baseflow segment, the start and end of the baseflow segment must be determined. However, due to the difficult methodology involved in locating the start and end of a baseflow segment, estimating inverse slope of the baseflow segment is not straightforward. In other words, the recession analysis methods suffer from the lack of a consistent way of selecting a recession segment from a continuous flow record. However, there are two ways of determining the start and end of a baseflow segment: one is based on defining a constant value for each of these points, and the other one is based on a variable value. Based on the hydrological characteristics of a catchment, a constant value can be defined for each of these points. In contrast, in the second method, which is proposed for this study, a variable starting value can be defined as pure groundwater discharge at a given time after a rainfall event, and will take on different values for each event. Additionally, in humid to sub-humid climates, a recession period, or recession length, between 5 and 10 days is usually chosen. In this method the underlying assumption is that both precipitation and interflow components of a recession curve are relatively negligible; as a result, surfaceflow and baseflow constitute major components of a recession curve. The start of baseflow or the point at which the surfaceflow ceases can be estimated from either of the two following empirically based equations (Linsley et al., 1975):

$$N = A^{0.2} \quad \text{Equation 4}$$

where  $N$  is the number of days between the storm peak (peak of stream-hydrograph) and start of the baseflow (end of surfaceflow),  $A$  is the drainage basin area in square miles,

or

$$N = 0.827A^{0.2} \quad \text{Equation 5}$$

where  $A$  is the drainage basin area in square kilometers. It should be noted that Equations 4 and 5, due to their empirical nature, are dimensionally incorrect.

In most cases, more non-linearity can be observed at the upper part of a recession curve, as compared to the lower part, for the following reasons. Almost immediately after the peak, the surface flow component may still be significant resulting in non-linearity of the recession curve. The time (in terms of day) after the peak at which surface flow seems to cease can be estimated from Equation 5. Additionally, after the cessation of surface flow, the recession curve still tends to be non-linear due to the instability of the water table profile. The time period since the recharge event, or critical time  $t_c$ , during which the profile of water table distribution is unstable can be estimated from the following equation which is obtained by combining various equations from Rorabaugh (1964) and Rorabaugh and Simons (1966):

$$t_c = 0.2144K \quad \text{Equation 6}$$

where  $t_c$  is critical time and  $K$  is recession index, or amount of time during which groundwater discharge drops through one log cycle after the critical time. Finally, it should be noted that the minor peaks caused by negligible recharge are not taken into account, as it relates to estimation of critical time. This is because the period of non-linearity caused by a minor peak may be much shorter than critical time.

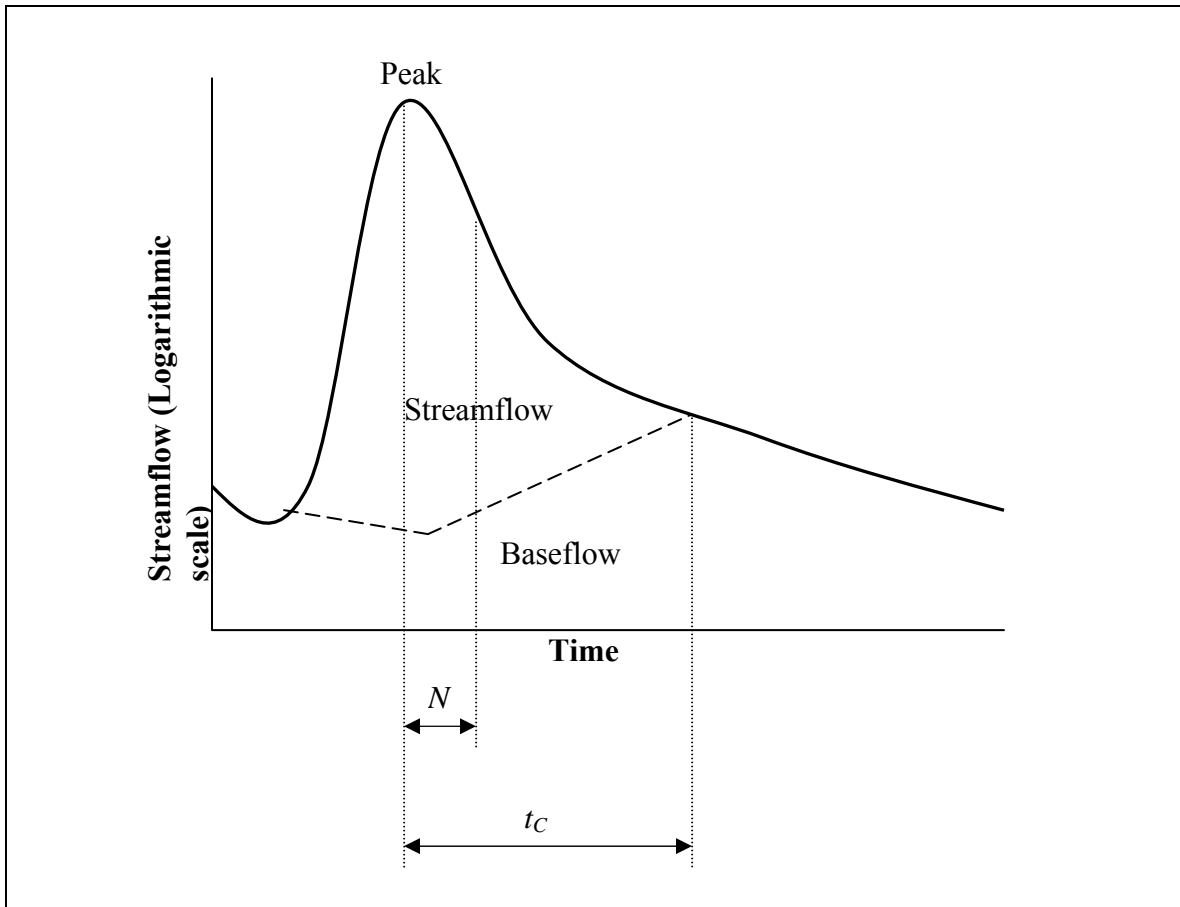


Figure 2-4: Separation of a streamflow hydrograph into surfaceflow and baseflow components using Equation 4 as well as showing the critical time

Similar to locating the start of recession, the recession length, represented by the number of timesteps in the flow sequence can be a constant or variable value (Tallaksen, 1995). A minimum length of recession is usually chosen between 4 and 10 days, depending on the mean duration of dry spells in the region (Tallaksen, 1995).

## 2.4 Master Recessions

According to Tallaksen (1995), a master recession curve (MRC) is defined as an envelope of various individual baseflow-segments representing an averaging of results for a set of storm-events for a given catchment. Climatic factors, such as precipitation and evapotranspiration, highly impact the recession rate resulting in high variability of recession rate.

For example, in a humid climate, the recession period is frequently interrupted by the rainfall events. As a result, a set of rainfall events generates a series of baseflow segments of varying duration causing great variability of the recession constant  $K$  for a given catchment. The MRC is a means to overcome the high variability of recession constant by constructing a mean recession curve. In other words, the MRC is constructed of numerous segments of continuous recession, each of which occurs at various time periods, or storm events; as a result, the MRC represents the groundwater discharge for a time period greater than any such time that could be observed in nature (Rutledge and Mesko, 1996). In addition, evapotranspiration has a great impact on recession rate enhancing the non-linearity of baseflow recession as well as causing high variability of recession index  $K$ . Hence, Rutledge and Mesko (1996) state that for the purpose of minimizing the evapotranspiration effect on recession index, the MRC must be preferably assembled from non-summer streamflow data. Recession characteristics contain valuable information from which aquifer properties along with the sustainability of groundwater discharge can be evaluated.

A number of methods might be applied to construct a master recession curve, such as correlation method, matching strip method, and tabulation method. For complete information on these methods, the reader may refer to the papers of Tallaksen (1995), and Sujono et al. (2004).

#### **2.4.1 The MRC characteristics**

According to Rutledge and Mesko (1996), the important features of MRC adopted to make possible the comparison of hydrological characteristics of various parts of the study area are its inclination (slope of the MRC) and shape. The MRC can be defined by the following equation:



$$t = A(\log Q)^2 + B(\log Q) + C \text{ Equation 7}$$

where  $t$  is time and  $A$ ,  $B$ , and  $C$  are coefficients. The inclination of MRC is represented by the median value of the recession indices of baseflow segments from which the MRC is assembled. On the other hand, shape is a function of the second derivative of the MRC, or  $2A$ . For example, a concave MRC is generated if this variable is positive, while a negative value generates a convex MRC. Finally, if  $2A$  equals zero (or nearly zero), the MRC is linear (or nearly linear).

The authors state that different values of inclination and slope of MRCs seem to be the result of variation in such parameters as transmissivity, storage coefficient, average distance from a stream to hydrological divide, precipitation, relief, and geology (e.g. rock type and structure) in various parts of the study area. It is well known that under the most favorable circumstances, large transmissivity  $T$  and storage coefficients  $S$  favor a sustainable water supply for a given region. However, the opposite effects of these variables on the recession index complicate the interpretation of groundwater recession as they relate to water supply potential (Rutledge and Mesko, 1996). For example, an increase in  $T$  is likely to cause a drop in the recession index, whereas increases in the recession index are often caused by increases in  $S$ . Nevertheless, as a rule of thumb, a large recession index often indicates the best conditions for water supply. That is, storage coefficient outweighs transmissivity from the point of view of the water-supply potential (Rutledge and Mesko, 1996). In addition, basins with higher relief seem to exhibit larger recession indices.

On the other hand, when the water table declines or the zone of saturation of an aquifer becomes thinner, the transmissivity decreases, theoretically resulting in concavity of the MRC. Furthermore, if the storage coefficient increases as the water table declines, then the MRC tends to be concave. In contrast, a convex MRC may be as a result of a gradual decrease in storage

coefficient as the water table drops. In addition, concavity of the MRC can be caused by a rise in a basin relief. A gradual increase in relief also may result in a rise in recession index. For example, in a basin with high relief when the water table declines, some of the stream segments located near the divide line may go dry thus causing aquifer half-width  $a$  to increase. An increase in aquifer half-width  $a$  is likely to cause a rise in the recession index (as revealed by Equation 1). According to Rutledge and Mesko (1996), in some circumstances (e.g. in Blue Ridge Physiographic Province), the concavity of the MRC that results from the high relief seems to outweigh the convexity caused by the reduction of storage coefficient as the water table drops. Rutledge and Mesko (1996) also state that if more than two aquifers discharge their water into a stream, the MRC may become concave. This is because the combination of two near-linear lines with different slopes each of which represent the baseflow rates of each of the two aquifers may result in the concavity of the MRC. In contrast, if a stream downwardly leaks into a deeper groundwater-flow system, the MRC may become convex. Table 2-4 summarizes the relationships between the MRC characteristics and the aquifer properties as well as basin relief.

**Table 2-4: Relationships between the MRC characteristics and the aquifer properties as well as the basin relief**

<i>Concavity of the MRC can occur:</i>	<ul style="list-style-type: none"> <li>• when the aquifer transmissivity decreases as the water table drops</li> <li>• when the storage coefficient of an aquifer increases as the water table declines</li> <li>• When more than two aquifers feed a stream.</li> <li>• for basins with high relief</li> </ul>
<i>Convexity can occur:</i>	<ul style="list-style-type: none"> <li>• when the storage coefficient decreases as the water table declines</li> <li>• when a stream downwardly leaks into deeper groundwater-flow-systems</li> </ul>
<i>A large recession index may be as a result of:</i>	<ul style="list-style-type: none"> <li>• large storage coefficient of an aquifer feeding a stream</li> <li>• large relief of a basin</li> </ul>

## 2.5 Mathematical Basis for Determining Coefficient of Transmissivity

### *T and Storage Coefficient S*

Rorabaugh (1960) derived a formulation for the recession of groundwater level  $h$  based on simple aquifer geometry and boundary conditions similar to Figure 2-5. The aquifer is considered to be thick relative to  $h$ , wide relative to its thickness, and underlain by impermeable material. Its side boundaries are vertical and fully penetrating, and it is uniform, isotropic, and homogeneous (Rutledge, 2006). The aquifer is assumed to have parallel boundaries. The formulation is based on one-dimensional flow, and transmissivity is considered to be constant. The recession curve will appear linear on a semilog plot after a time period has elapsed since the last recharge event. During this critical time  $t_c$  the ground water head profile is unstable. Critical time  $t_c$  is defined as follows:

$$t_c = \frac{0.15a^2 S}{T} \quad \text{Equation 8}$$

where  $a$  is the distance from the stream to the hydrologic divide,  $S$  is the storage coefficient, and  $T$  is transmissivity. After the critical time, the aquifer hydraulic diffusivity  $\frac{T}{S}$  can be calculated from the slope of the straight-line segment, using the following equation (Rorabaugh 1960):

$$\frac{T}{S} = \frac{0.933a^2 \log(h_1/h_2)}{t_2 - t_1} \quad \text{Equation 9}$$

where  $h_1$  is the water level at time  $t_1$  during the period of recession and  $h_2$  is the water level at time  $t_2$  during the same period of recession. Equation 9 can be rearranged by replacing

$(\frac{\log(h_1/h_2)}{t_2 - t_1})$  with  $\frac{1}{K}$  to give Equation 1.

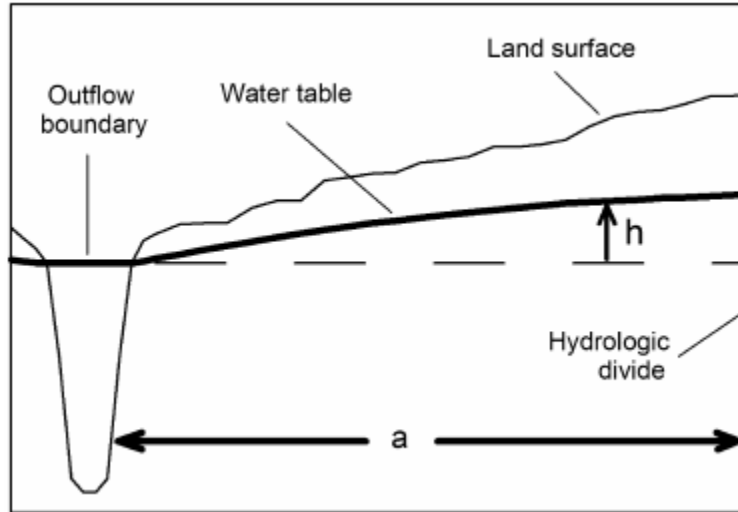


Figure 2-5: General aquifer geometry and boundary conditions (not to scale) (from Rutledge, 2006)

Coefficient of transmissivity is ordinarily measured by means of a pumping test made with a pumped well and one or more observation wells. The methods have become standardized. The standard test procedures, although yielding reasonably accurate results, may be interpreted only in terms of the relatively small sample of material in the vicinity of the pumped well. Moreover, in many areas pumping-test data are not available. However, Olmsted and Hely (1962) believe that a rough estimate of the average coefficient of transmissivity may be made, using the total groundwater recharge (assumed equal to baseflow plus evapotranspiration from groundwater), the average hydraulic gradient (water table slope) adjacent to the discharge areas, and the length of discharge areas. The following equation from Olmsted and Hely (1962) describes the coefficient of transmissivity:

$$T = \frac{R_g + ET_g}{I(2L)} \quad \text{Equation 10}$$

where  $T$  is coefficient of transmissivity (ft<sup>2</sup>/day),  $R_g$  is effective recharge (or the baseflow) (ft<sup>3</sup>/day),  $ET_g$  is riparian evapotranspiration (ft<sup>3</sup>/day),  $I$  is the average groundwater gradient from divides to streams (ft/mi), and  $L$  is the length of discharge areas or total stream length (mi).

## 2.6 Previous Investigations

Olmsted and Hely (1962) were first to estimate transmissivity and storage coefficient using the Equation 10 and the following equation on a regional basis:

$$Y_g = \frac{\Delta S_g}{\Delta H_g} \quad \text{Equation 11}$$

where  $Y_g$  is gravity yield (a dimensionless ratio),  $\Delta S_g$  is the increase in groundwater storage in a specific time period (expressed in inches of water over the area), and  $\Delta H_g$  is the corresponding increase in ground water stage (expressed in inches). Olmsted and Hely (1962) also stated that the change in stage was measured in observations wells, but the change in storage was determined indirectly from baseflow data. In addition, the authors utilized the observation wells for the estimation of hydraulic gradient. The estimated transmissivity and gravity yield for the rocks of Brandywine Creek basin, Pennsylvania were respectively on the order of 1,000 (gal/day) (ft) and 0.075-0.1.

Rutledge and Mesko (1996) exhibited how streamflow records in the Appalachian highland region of the United States can provide estimates of hydraulic properties of the shallow aquifer system. The authors stated that their analyses were categorized into two groups: streamflow recession and baseflow. Their study showed that streamflow records can reveal information about the hydraulic diffusivity of aquifers and their capability to provide sustainable water supply when the recharge is low during the time of sparse rainfall. The authors also stated that baseflow hydrograph techniques were utilized to estimate the groundwater budget for the Appalachian highland region. In addition, Equation 1 was used for the estimation of hydraulic diffusivity and transitivity. Moreover the authors reported that the average values of hydraulic diffusivity and transmissivity were around 20,000 (ft<sup>2</sup>/d) and 900 (ft<sup>2</sup>/d), respectively. The

authors also reported that based on the 89 basins with continuous records (from 1964 to 1990), the median recharge, discharge, and evapotranspiration were respectively around 13, 12, and 15 (in/yr). The authors used a simplistic approach wherein the few results available for storage coefficient  $S$  in the literature were put into Equation 1 to estimate the areal transmissivity. The Rutledge and Mesko (1996) method, however, could be more accurate if the GIS tools were applied for estimation of aquifer half-width  $a$  and areal transmissivity  $T$ .

Nelms et al (1997) estimated areal transmissivity and storage coefficient using Equation 1 and Equation 10 for shallow aquifer system in Virginia. The authors stated that groundwater gradients were determined using a procedure based on median basin relief and the assumption that water table profile mimics the topography of land surface. Furthermore, the median water table elevation beneath the divides was assumed to be half the median elevation of the divides. Moreover, Nelms et al (1997) used the recession index values determined by Rutledge and Mesko (1996) to calculate the hydraulic diffusivities for their study area.

No published study is identified where hydraulic properties of the shallow aquifer system in the southeastern U.S. (Georgia and North Carolina) are estimated on a regional basis. The methodology of this study is an organized synthesis of previous methods such as computerized hydrograph analysis methods (Rutledge and Mesko, 1996), GIS analysis procedure (Nelms et al, 1996), and Rorabaugh and Olmsted equations (Rorabaugh, 1962; and Olmsted and Hely 1962). In other words, this study will extend the use of previous methods (Olmsted and Hely, 1962; Hoos, 1990; Rutledge and Mesko, 1996; Nelms et al., 1997) and apply them to the selected basins in the southeastern United States.

## 3 METHODS

### 3.1 Introduction

The research methodology of this study (summarized in Figure 3-1) is a synthesis of previous methods applied by several investigators (Olmsted and Hely, 1962; Hoos, 1990; Rutledge and Mesko, 1996; and Nelms et al, 1997) to quantify aquifer properties, recession index, and groundwater budget. For example in this study, RECESS program (Rutledge, 1993) is used to determine median recession index and to define the master recession curve (MRC) for each basin. Rutledge and Mesko (1996) applied RECESS to the Appalachian Highland Region streamflow data for the period from about 1960 to 1991. However, in my study the RECESS was applied to 1970-2007 streamflow data allowing the comparison of estimated recession indices with those from Rutledge and Mesko (1996). Furthermore, no published study was identified where RECESS have been applied to quantify the effect of evapotranspiration on recession index on a regional basis. Moreover, in this study, the estimates of  $T$  and  $S$  for basins are based on the method that first was introduced by Olmsted and Hely (1962). However, there is no published study that has applied Olmsted method to the southeastern U.S. streamflow data.

GIS was utilized to estimate the total stream length, drainage density, and groundwater slope for the basins. The GIS was also applied to relate, in various combinations, the following available data layers: (1) soil data from the Soil Conservation Service, (2) land use data from United State Geological Survey, (3) curve number map created in this study, and (4) other relevant data layers. The GIS is also useful in determining factors that have control on recession index and aquifer properties.

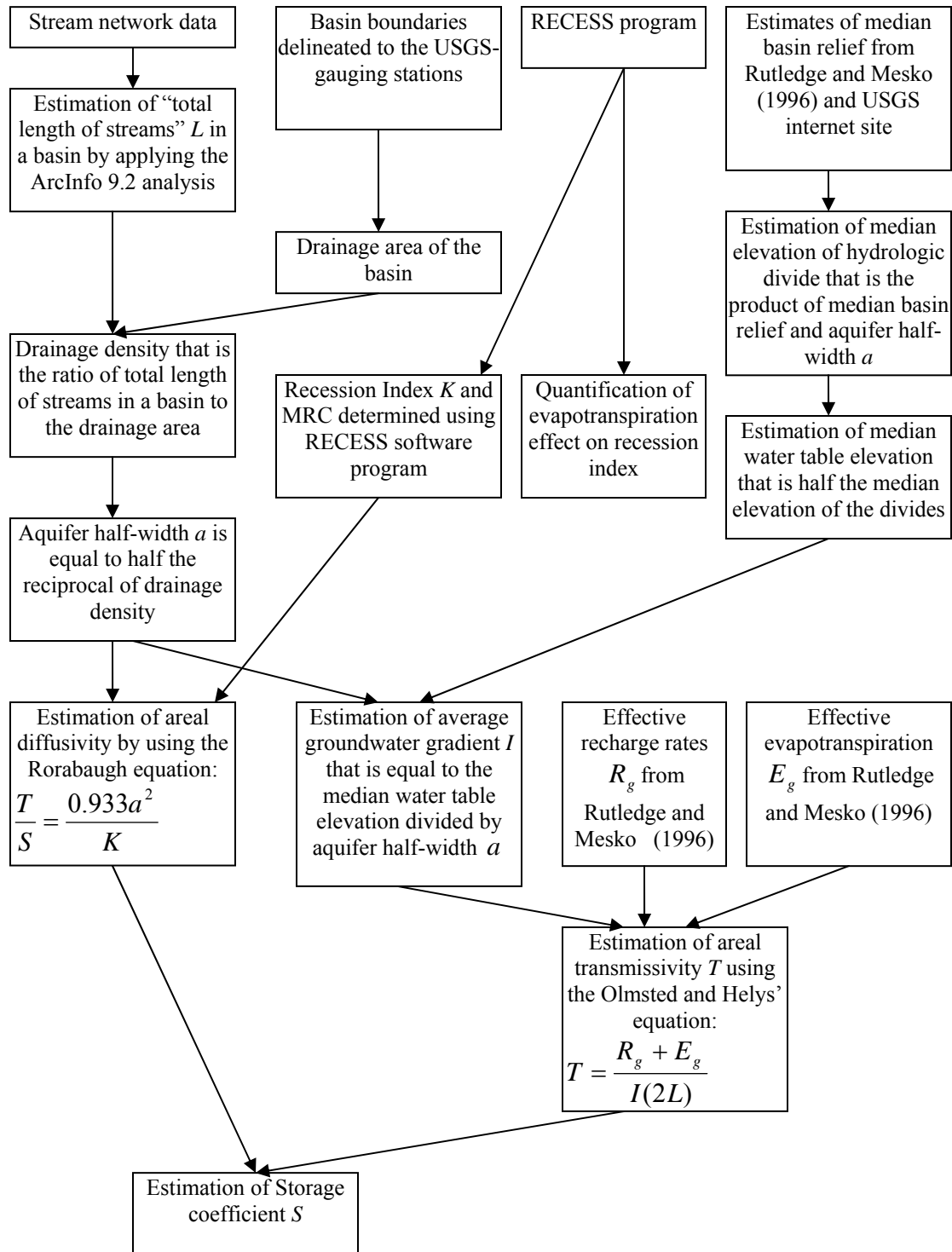


Figure 3-1: A simplified flow chart of the project tasks



## **3.2 Assumptions and Applicability of Methods**

### **3.2.1 Assumptions for use of RECESS program to estimate recession**

#### **index**

The RECESS program calculates the recession of groundwater discharge by selecting several periods of continuous streamflow recession. The selection of these periods must meet the condition that for each pair of subsequent daily values, the first value exceeds or is equal to the second. Furthermore, the period of recession must be long enough so that the profile of groundwater distribution is nearly stable. Moreover, the regulation and diversion of flow for the selected gauging stations should be negligible (Table 3-2). Finally, the designation of baseflow from the straight semi-log segment of the hydrograph is purely “operational” and it cannot be demonstrated directly that the selected baseflow segment is in fact pure groundwater discharge.

### **3.2.2 Assumptions used in the mathematical model described by**

#### **Rorabaugh**

Assumptions used in the development of the mathematical model described by Rorabaugh are: (1) The aquifer is thick relative to change in the water table elevation caused by the recharge; (2) The aquifer is isotropic and homogenous; and (3) the aquifer is underlain by impermeable materials (Table 3-2).

### **3.2.3 Assumptions used to estimate the median groundwater slope**

Assumptions used for estimating the median groundwater slope include: (1) Water table profile crudely mimic the land surface topography and (2) median water table elevation is half the median elevation of divide line (Table 3-2).

LeGrand (1989) stated 17 key generalizations about the groundwater setting in the Piedmont region based on tens of thousands of well records and observational study of thousands of terrain settings in the Piedmont region of the southeastern U.S. His approach relied primarily on understanding the processes operating and secondarily on statistical analyses. The author stated that the topography of the water table is crudely similar to that of the land surface; thus, it is easy to construct raw synthetic water table maps of a scale of 1:24,000 without water table measurements.

Olmsted and Hely (1962) computed the average water table slope as the average slope of the water table between the observation wells and the discharge outlets in the Brandywine Creek basin, Pennsylvania. The authors stated that the average depths to water were calculated as the arithmetic average of the seasonal extremes for the calendar years 1952-53. In addition, these depths were subtracted from the elevation of the land surface at each of the well sites to obtain the elevation of the water table. The difference in elevation was divided by the horizontal distance from each well to the discharge outlet to obtain the average water table slope in feet per mile. The average slope of groundwater at each of the 16 observation wells and the overall average slope of groundwater for the Brandywine Creek basin are listed in Table 3-1.

Well	Distance to discharge outlet (miles)	Average difference in elevation of water table (feet)	Average slope of water table (feet per mile)
Ch- 2-----	0. 40	17	42
3-----	. 25	19	76
4-----	. 18	59	330
5-----	. 23	46	200
6-----	. 18	5	28
7-----	. 20	13	65
8-----	. 16	12	75
9-----	. 21	44	210
10-----	. 18	5	28
11-----	. 20	38	190
12-----	. 08	11	140
13-----	. 16	70	140
14-----	. 07	20	290
15-----	. 13	39	300
De- 3-----	. 17	70	410
Average-----	-----	-----	190

Table 3-1: Average groundwater slope for the observation wells (electronically copied from Olmsted and Hely, 1962)

Furthermore, another estimate of average water table gradient was made by using stream density, average landslope, and the assumption that average water table elevation beneath the divide line is half the average elevation of divide line. The authors stated that stream density was 2.26 miles per square mile; average inter-stream distance is the reciprocal of stream density; and the average distance from streams to divides ( $a$ ) is half the average inter-stream distance:

$$\frac{1}{2 \times 2.26(mi)(mi)^{-2}} = 0.22mi$$

The authors also stated that the average landslope is 450 feet per mile, and the average elevation of divides above the streams (the average relief) is

$$450 \frac{ft}{mile} \times 0.22mile = 100 feet$$

Moreover, the average depth to the water table beneath divides was assumed to be half the average elevation of divide lines. Therefore, the average slope of the water table from divides to streams is:

$$(100 - 50) \text{ feet} \div 0.22 \text{ mile} \approx 225 \frac{\text{feet}}{\text{mile}}$$

This result is more nearly comparable with the groundwater slope obtained from the observation wells (190 from the observation wells and 225 from the second method).

**Table 3-2: Applicability of assumptions used in the methodology of this study**

<b>Assumptions</b>	<b>Applicability of assumptions</b>
Estimation of recession index during periods when no recharge is occurring.	In the RECESS, these continuous periods must meet the condition that for each subsequent daily values, the first value exceeds or is equal to the second.
The profile of groundwater distribution must be nearly stable	The number of days required to detect a recession period was set to 15. Most departures from linearity occur within the first 6 days of each period of recession.
The regulation and diversion of flow for selected gauging stations should be negligible	The gauging stations with negligible diversion are classified as “good” in the USGS data book
The aquifer is homogenous and isotropic	Drainage areas of many square miles allow for local variability of basin and aquifer parameters to be averaged.
Water table profile crudely mimics the land surface topography	This assumption is supported based on statistical analysis on thousands of well data.
Median water table elevation is half the median elevation of divide line	This assumption was validated by Olmsted and Hely (1962).

### **3.3 Site Selection**

A total of 44 of streamflow gauging stations were selected in the study area for analyses. The criteria for selection of these stations include negligible regulation of diversion of flow, a record that was classified as “good” in USGS data book, and 100 percent of the drainage area

being within the study area. The drainage area of all stations are larger than 10 (mi<sup>2</sup>) and less than 270 (mi<sup>2</sup>). The recession index calculation was performed using streamflow data for the period from 1970 through 2007.

### **3.4 Data**

To accomplish a spatial-hydrologic project several types of datasets are required. In ArcGIS, these data layers can be overlain on each other to obtain new information such as spatial distribution pattern of a given hydrologic variable. As a result, it is appropriate to describe these data and the internet sites from where they can be obtained. Below is a list of data that will be used in this study:

- Basin boundary
- Stream data (RF3 dataset)
- Land use
- Soil data

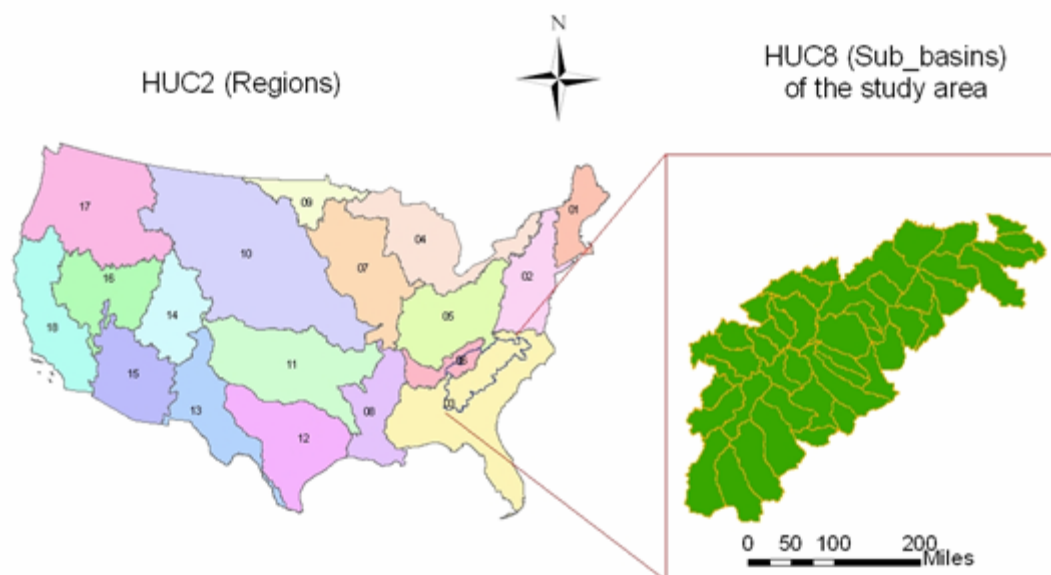
These datasets are discussed in the following sections.

#### **3.4.1 Basin boundary**

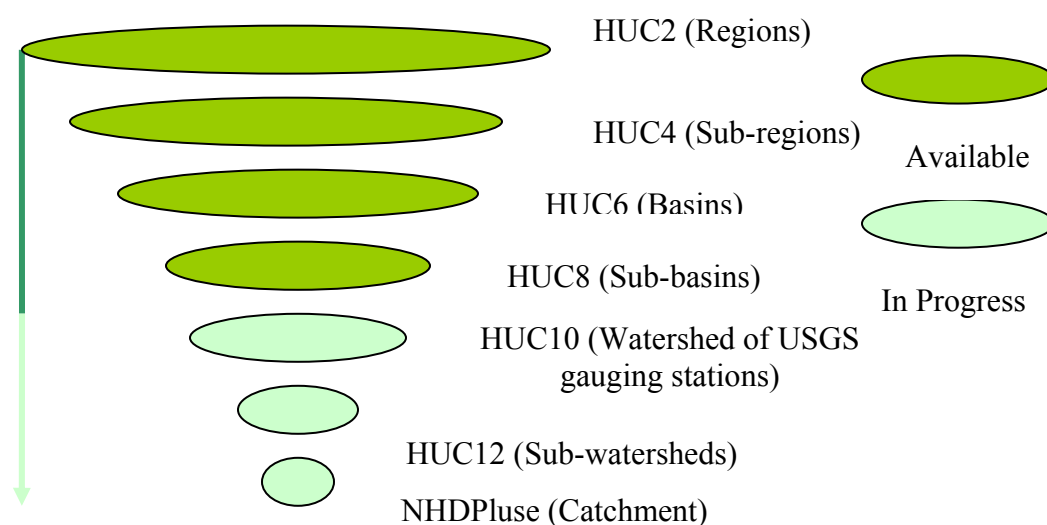
The United State Geological Survey (USGS) has developed a hierarchy of seven levels in which the United States is successively subdivided into the smaller watersheds. In this classification watershed datasets are defined by Hydrologic Unit Codes (HUC) (Seaber et al, 1987). The highest hierarchical level is called region and assigned a two-digit code. Each region consists of several sub-regions designated by a 4-digit code. Likewise, each sub-region is made of several basins with 6-digit code. Furthermore, each basin constitutes several sub-basins with 8-digit code. In the same way, each sub-basin consists of several watersheds with 10-digit code.

Finally, each watershed makes up several sub-watersheds with 12-digit hydrologic codes. The first four levels of hydrologic unit boundaries (HUC2, HUC4, HUC6, and HUC8) are available at scales of 1:2,000,000 and 1:250,000. The 1:250,000-scale hydrologic units for the conterminous United States can be downloaded at no charge at the following USGS webpage:

<http://water.usgs.gov/GIS/metadata/usgswrd/XML/huc250k.xml>

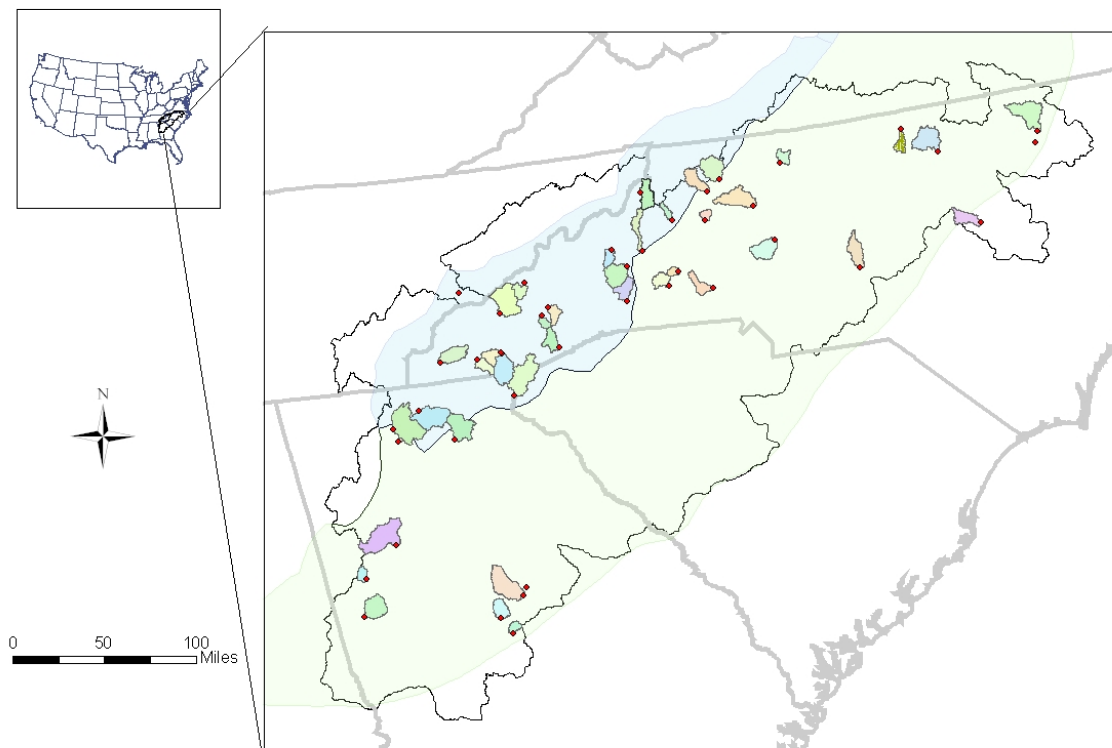


**Figure 3-2: Hierarchical classification of watersheds**



**Figure 3-3: Hierarchical structure and availability of water resource boundaries in the United States**

In this study, 33 drainage basins in North Carolina and 12 basins in Georgia were examined. All of these basins are delineated to the USGS gauging stations. The basins in North Carolina were delineated using a 20 foot Lidar DEM and obtained from North Carolina Water Science Center. The basins in Georgia were obtained from Water Resource Data-Georgia, 2003.



**Figure 3-4: Location of the streamflow gauging stations in the Blue Ridge and Piedmont Physiographic Provinces of the study area**

### 3.4.2 Stream data

The rivers of the United States have been cataloged in a set of river Reach Files. The first version, Reach File 1 (RF1) was completed by the U.S. Environmental Protection Agency during the 1970s. The river lines are mapped at approximately 1:500,000 scale. These lines are merged into a network; they are joined continuously through the landscape and are stored using the HUC8 drainage boundaries as the spatial coverage units. Following the development of Reach

File 1, efforts continued at the EPA to improve the system. The result of these efforts was Reach File 3 mapped at 1:100,000 scale using a similar scheme to Reach File 1. These reaches are for main rivers and streams, which show up in 1:24,000 scale maps as fairly broad lines. When the vector lines from Reach File 3 are overlaid on 1:24,000 scale Digital Raster Graphic maps, they match. During the 1990s, the EPA and the USGS cooperated in a new effort, called the National Hydrography Dataset (NHD). NHD is mapped at 1:100,000 scale, as is RF3, but the data quality of the NHD is better. Also, the NHD has an improved data structure allowing to be used with maps of different scales. The RF3 data for this study was downloaded from the EPA basin site: <http://www.epa.gov/waterscience/BASINS/>

### **3.4.3 Land use and land cover data**

Land cover and land use data describe physical and cultural features on the earth surface, respectively. Examples of natural physical features described by land cover data are type of vegetation, the percentage of impervious cover, and the percentage of tree canopy. As its name suggests the land use features are referred to the type of use that a land cover feature represents. For example, agricultural land use is the type of use that a vegetation land cover may represent. Other examples of land use types are urban, forestland, wetland, and barren land.

Land use and land cover (LULC) data files are provided by the USGS as a part of its National Mapping Program. These files are obtained from Landsat TM (Thematic Mapper) 1960's and 1970's imagery. In addition, 1:250,000-scale topographic maps are used for compilation and organization of these files (Maidment, 2005). LULC1970 is the standard land use and land cover coverage for the conterminous U.S. There are several online sources where these land use data can be obtained. For example, multizone downloads of LULC1970 are available at the following USGS website:



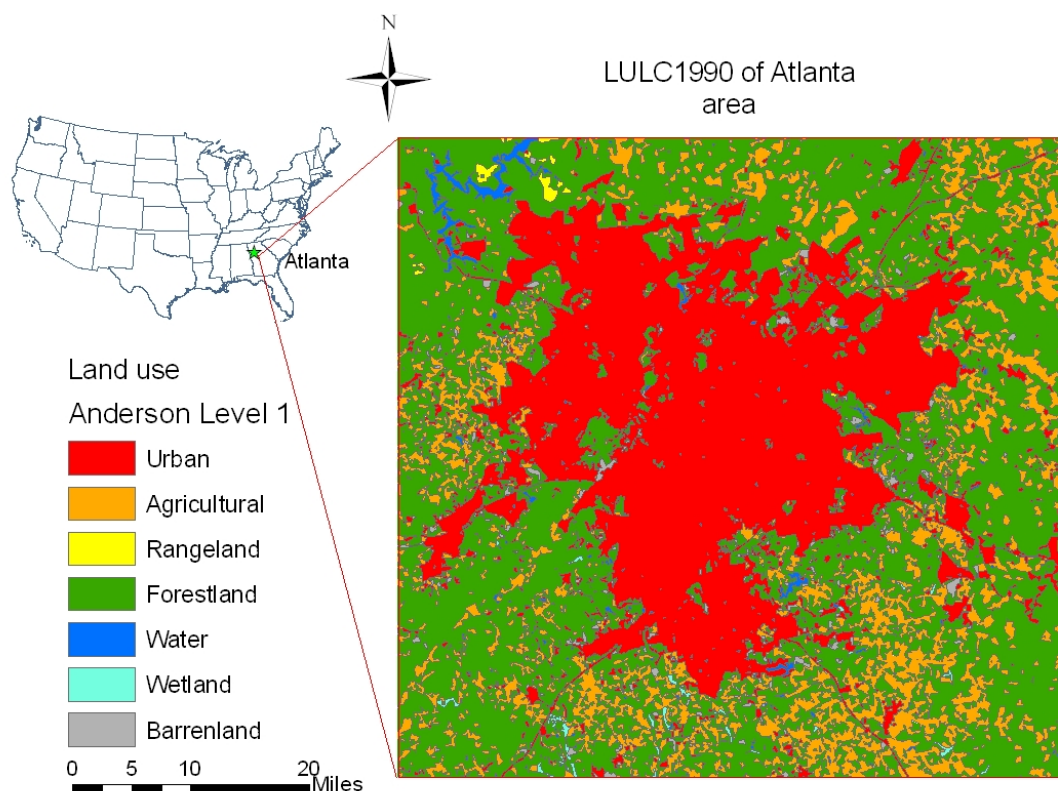
<http://eros.usgs.gov/geodata/>.

The LULC data at this site are provided in the Geographic Information Retrieval Analysis System (GIRAS) format. In addition, a single refined LULC1970 file for the conterminous U.S. can be obtained from the USGS site:

<http://water.usgs.gov/GIS/metadata/usgswrd/XML/newlu90g.xml>

The land use data for this study was downloaded from this site. The format of these datasets are .e00 (Standard Interchange File) and can be easily transformed to the “Coverage ArcInfo” format in the ArcGIS environment. In Addition, these data are projected to USGS national Albers Projection system (Figure 3-5).

The most popular classification systems for the land use and land cover data are Anderson level 1 and Anderson level 2. In Anderson lever 1, the features on the Earth’ surface fall into 7 categories, each designated by a 1-digit number. Figure 3-5 shows the Anderson lever 1 classification of the LULC data for the Atlanta area. A subdivision of Anderson level 1 is Anderson lever 2 where each feature is represented by a 2-digit code. In other words, Anderson level 2 provides more details of land use and land cover data for a given area. For examples, codes from 11 through 17 represent different urban areas. Table 3-3 shows the type of land use and land cove features represented by 2-digit codes.



**Figure 3-5:** This is the USGS LULC1970 of Atlanta area that has been refined with 1990 population density. For example, any area with a population density of at least 1000 people per square mile is assigned as the urban land.

**Table 3-3: Display of Anderson level 1 & 2 (from Anderson et al, 1976)**

Description of Land Use	Anderson level 1	Anderson level 2
Residential	1 Urban	11
Commercial and Services		12
Industrial		13
Transportation, Communications, and Utilities		14
Industrial and Commercial Complexes		15
Mixed Urban or Built-up Land		16
Other Urban or Built-up Land		17
Cropland and Pasture	2 Agricultural	21
Orchards, Groves, Vineyards, Nurseries, and Ornamental Horticultural Areas		22
Confined Feeding Operations		23
Other Agricultural Land	3 Rangeland	24
Herbaceous Rangeland		31
Shrub and Brush Rangeland		32
Mixed Rangeland		33
Deciduous Forest Land	4 Forestland	41
Evergreen Forest Land		42
Mixed Forest Land		43
Streams and Canals	5 Water	51
Lakes		52
Reservoirs		53

Bays and Estuaries		54
Forested Wetland	6 wetland	61
Non-forested Wetland		62
Dry Salt Flats		71
Beaches		72
Sandy Areas other than Beaches	7 Barren land	73
Bare Exposed Rock		74
Strip Mines, Quarries, and Gravel Pits		75
Transitional Areas		76
Mixed Barren Land		77
Shrub and Brush Tundra		81
Herbaceous Tundra	8 Tundra	82
Bare Ground Tundra		83
Wet Tundra		84
Mixed Tundra		85
Perennial Snowfields	9 Perennial	91
Glaciers	snow & ice	92

The LULC1970 is now dated, and therefore a new national land use mapping effort is in progress to obtain a standard land use and land cover map for the conterminous U.S. from the Landsat TM1992 imagery. The outcome of this effort is the National Land Cover Data (NLCD1992) with 21 categories of land cover types. The multizone downloads of these data are available at:

<http://www.mrlc.gov/index.asp>.

The LULC1970 data are still adequate for some applications. Generally, land use and land cover data can be used in estimating runoff, modeling nutrient and pesticide runoff, and assessing ecosystem status (Maidment, 2005).

#### 3.4.4 Soil data

The Natural Resources Conservation Service (NRCS) has developed a 1:250,000-scale soil dataset for the conterminous U.S. This dataset is called the State Soil Geographic (STATSGO) and can be obtained at:

<http://www.ncgc.nrcs.usda.gov/products/datasets/statsgo/>.

The soil dataset used in this study is a modified version of STATSGO downloaded from the USGS site:

<http://water.usgs.gov/GIS/metadata/usgswrd/XML/ussoils.xml>.

A new effort is underway by the NRCS to develop a 1:24,000-scale soil map called the Soil Survey Geographic (SSURGO) database. SSURGO contains much more detailed soil information as compared with the STATSGO that is a generalized soil dataset. The preparation of SSURGO is still in progress and not yet completed for the conterminous U.S. This dataset is already available for much of the United States. For obtaining the SSUGO dataset the user is required to generate an online request. Then the NRCS will send the requested dataset to the customer's email account. The SSURGO data can be requested at the following site:

<http://www.ncgc.nrcs.usda.gov/products/datasets/ssurgo/>.

### **3.4.5 Surficial geology data**

This digital dataset describes surficial geology of the conterminous United States. The dataset was generated from a U.S. Geological Survey 1:7,500,000-scale map of surficial geology published as part of the U.S. Geological Survey National Atlas map series and can be obtained from the following USGS site:

[http://water.usgs.gov/GIS/metadata/usgswrd/XML/ofr99-77\\_geol75m.xml](http://water.usgs.gov/GIS/metadata/usgswrd/XML/ofr99-77_geol75m.xml)

This dataset can be used to illustrate the effect of regolith type on the hydraulic parameters of basins in the study area.

### **3.4.6 Curve number map**

The curve number (CN) is a hydrologic parameter used to describe the storm water runoff potential for drainage area. The curve number is a function of land use, soil type, and soil

moisture (e.g. if  $CN = 0$ , no runoff; if  $CN = 100$ , 100% runoff). The spatial information on land use, soil type, and rainfall data were incorporated. This digital information was kept in different layers. The generated layers were used for overlay analysis to drive the average curve number for the basins.

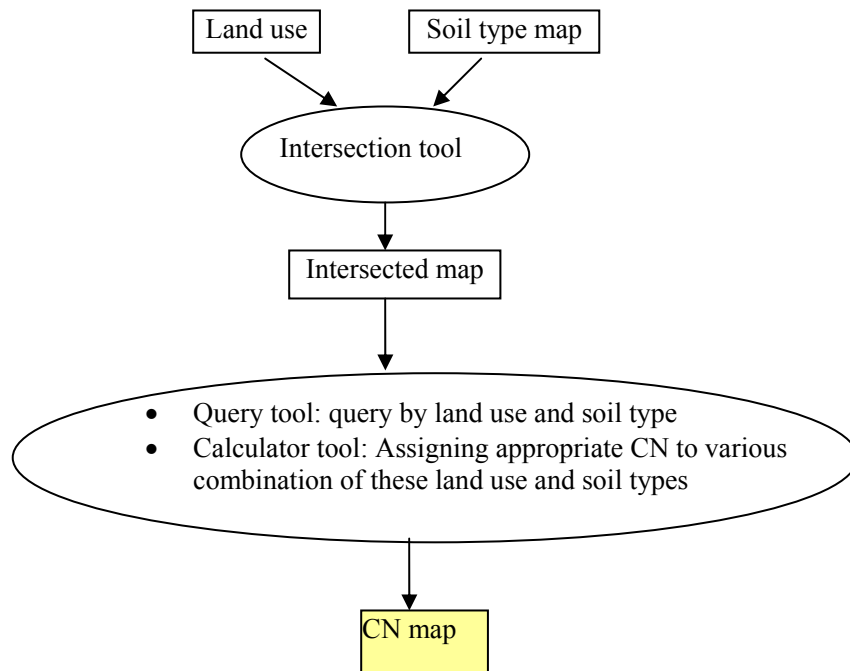
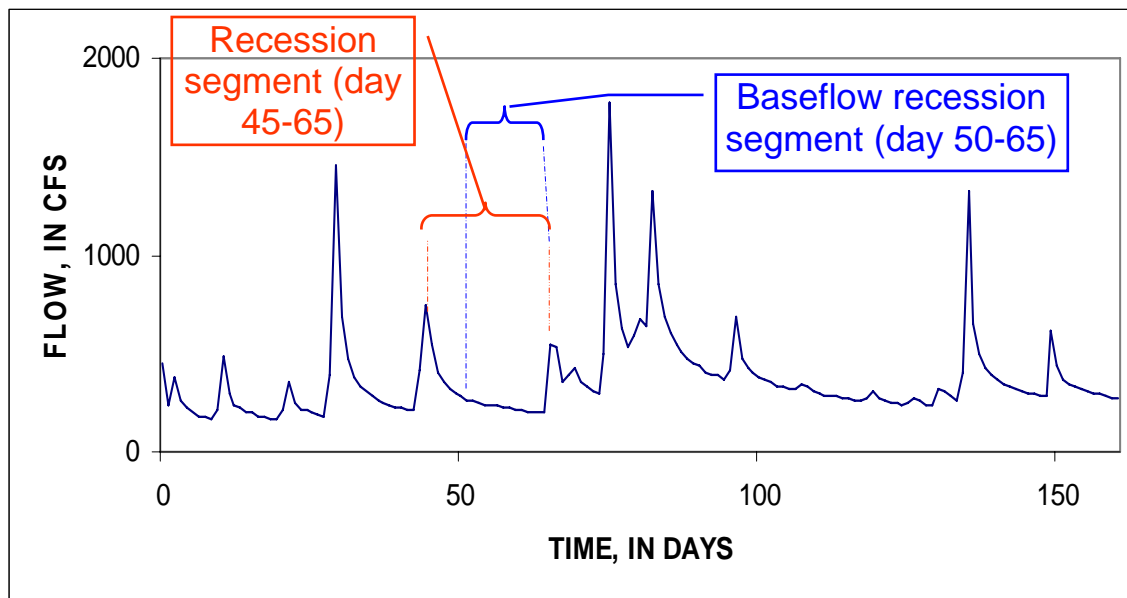


Figure 3-6: Flow chart for creating the CN map

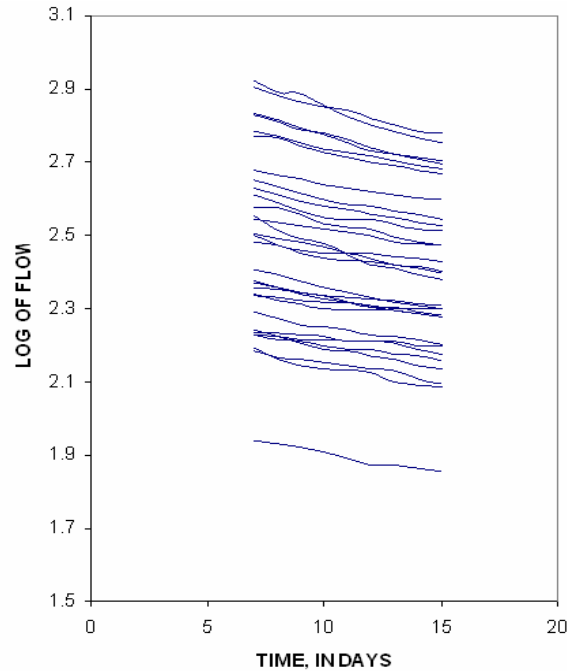
### 3.5 Illustration of use of RECESS in Estimation of $K_r$ and MRC for the Basins

The RECESS program is an empirical method for describing recession characteristics. The program scans the streamflow dataset, finding periods of continuous recession; such as the period from day 45 to day 65 in Figure 3-7. When a segment is found, the program user can decide whether it should be analyzed and, if so, which parts represent near-linear conditions (on the semi-log plot) that can be used to quantify the recession index. In this case (Figure 3-7), the selection might be the period from day 50 to day 65, as prior data indicate the effects of direct-

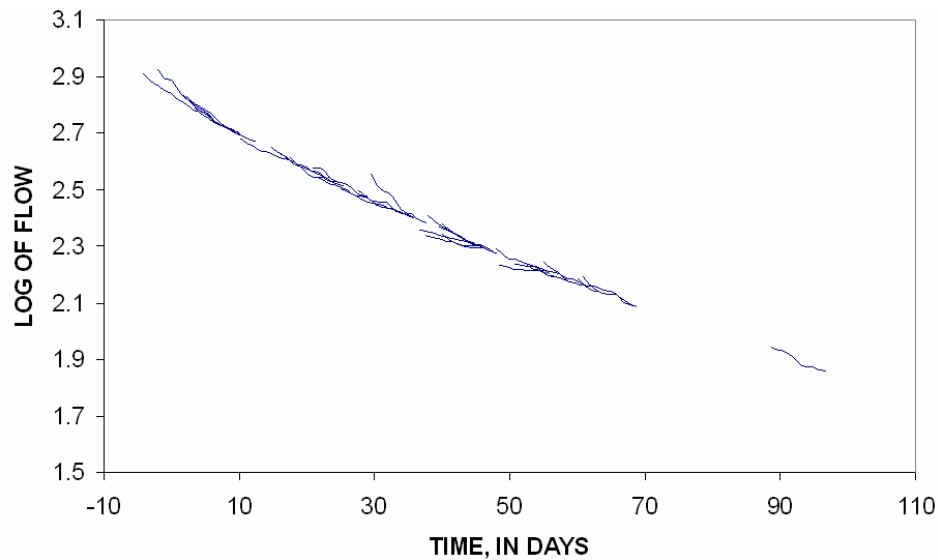
surface runoff or instability of the ground-water-head profile, or both. After the user has selected a segment, the program will calculate Equation 3. This equation is used to calculate the recession index. After this solution is obtained, the program proceeds to the next period of continuous recession, and repeats this process until a number of segments have been analyzed. After the program has found all recession segments in the period of interest, it determines the best-fit linear equation for  $K$  as a function of  $\log Q$ . Coefficients of this equation are used to adjust time on the x axis; and thus, obtaining MRC, which is a polynomial expression of time as a function of  $\log Q$ .



**Figure 3-7: Example of period of continuous recession (day 45 to day 65 or 11-1-06 to 3-10-07) that RECESS detect (The gauging station is Little Tennessee River near Prentiss, NC)**



**Figure 3-8: Baseflow recession segments generated from RECESS program (The x-axis is the time since the last peak) (Little Tennessee River near Prentiss, NC, 1970-2000)**



**Figure 3-9: MRC of the Little Tennessee River near Prentiss, NC (time period of 1970-2000) (The x-axis is the time that is adjusted)**

The RECESS allows user to select certain months for analysis. Furthermore, RECESS can be used to generate multiple MRCs. For example, the user can generate MRCs for summer and winter for the same streamflow datafile. The RECESS applied to the 44 drainage basins in the study area to determine the median recession index and define the MRCs for these basins.

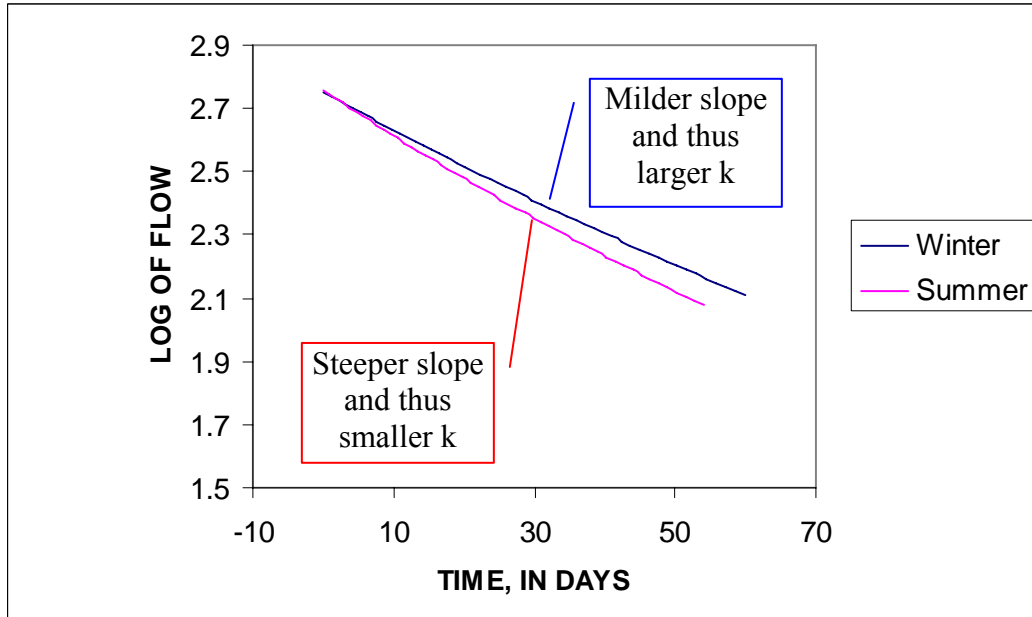


Figure 3-10: MRC of summer shows smaller averaged recession index than winter (summer < winter) (Litter Tennessee gauging station near Prentiss, NC)

### 3.6 Evaluation of Drainage Density and Aquifer Half-Width of Basins

Drainage density  $D$  is a fundamental property in geomorphology because it specifies the scale where there is a transition from hillslope to channel processes. Drainage density of basins was calculated as total length of stream segments  $L$  divided by the drainage area  $A$ :

$$D = \frac{L}{A} \text{ Equation 12}$$

where  $D$  is drainage density in (mi /mi<sup>2</sup>),  $L$  is total stream length in (mi), and  $A$  is drainage area in (mi<sup>2</sup>). Summation of all stream-segment lengths upstream of each streamflow gauging station was accomplished by applying the “Statistics” command to the Reach File 3 dataset. Likewise, the drainage area of basins was estimated using the “Statistics” command. Aquifer half-width  $a$  is the average distance from the stream to the hydrologic divide. The distance  $a$  for each gauging station is equal to half the reciprocal of drainage density

$$a = \frac{1}{2} \times \frac{1}{D} \text{ Equation 13}$$



where  $a$  is aquifer half-width in (ft). Aquifer half-width was estimated for all 44 basins within the study area.

### **3.7 Evaluation of Areal Diffusivity of the Basins**

Rorabaugh (1964) developed an equation that relates the slope of master recession curve to the transmissivity and storage coefficient of the groundwater reservoir. This equation was applied to 44 basins within the study area to calculate areal transmissivity, which is the ratio of transmissivity to storage coefficient (Equation 1:  $\frac{T}{S} = \frac{0.933a^2}{K}$ ). The recession index  $K$  values for basins were determined using a computerized method (RECESS software program) that calculates a mathematical expression of the master recession curve. In addition, aquifer half-width  $a$  for basins was estimated as illustrated in preceding section.

### **3.8 Evaluation of Median Basin Relief and Groundwater Gradients of Basins**

The median elevation and relief for basins were obtained from the USGS website and the previous literature (Rutledge and Mesko, 1996, Nelms and other, 1997). Median basin relief  $S_{median}$  is simply the median of all relief values calculated within a basin. Median basin relief values were multiplied by the values for aquifer half-width to obtain estimates for the median elevation  $E_{median}$  of hydraulic divides above the streams within each basin:

$$E_{median} = a \times S_{median} \quad \text{Equation 14}$$

Rough estimates for hydraulic gradient  $I$  were determined based on the assumption that water table profile mimics the topography of land surface. In addition, the median water table elevation beneath the divides was assumed to be half the median elevation of the divides

( $MedianWaterTableElevation = \frac{E_{median}}{2}$ ) (Olmsted and Hely, 1962, Nelms and others, 1997);

thus, the average groundwater gradient from divides to streams  $I$  was equal to the median water table elevation divided by aquifer half-width:

$$I = \frac{MedianWaterTableElevation}{a} \quad \text{Equation 15}$$

where  $I$  is groundwater gradient, and  $a$  is the aquifer half-width.

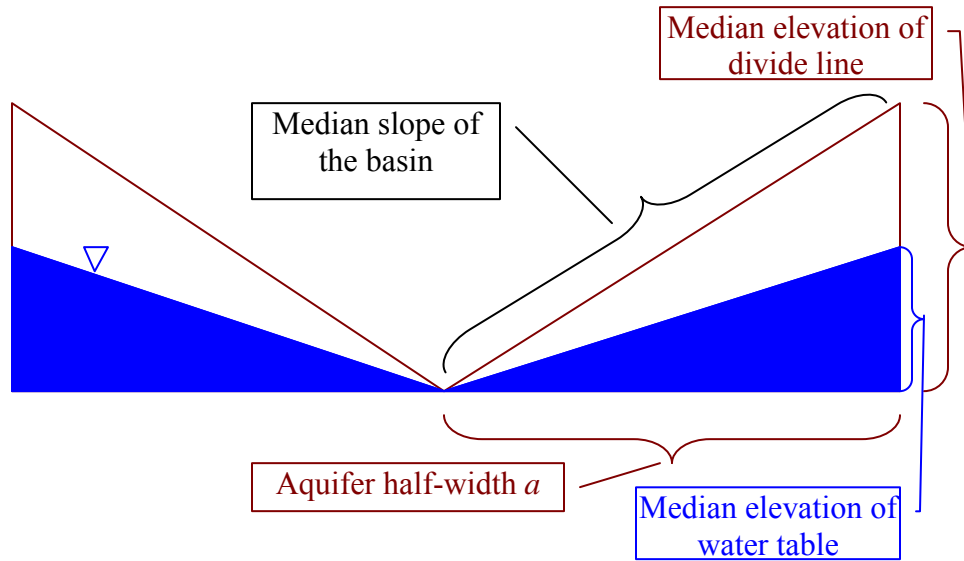


Figure 3-11: Estimation of Groundwater slope using Right Triangle Geometry

### 3.9 Estimation of Areal Transmissivity

Transmissivity is the rate at which water is transmitted through a unit width for the saturated thickness of the aquifer under a unit hydraulic gradient, expressed in (ft<sup>2</sup>/d). Rough estimates of areal transmissivity for all 44 basins were calculated using Equation 10

( $T = \frac{R_g + ET_g}{I(2L)}$ ) from Olmsted and Hely (1962). Values for effective recharge  $R_g$  and riparian

evapotranspiration  $ET_g$  determined by Rutledge and Mesko (1996) were used in Equation 10. In

addition, values for total stream length  $L$  and average groundwater gradient from divides to stream  $I$  were determined as explained in preceding section.

### ***3.10 Estimation of Areal Storage Coefficient***

Storage coefficient is defined as the volume of water released from storage per unit surface area per unit change in head. In an unconfined aquifer, the storage coefficient may be approximated by specific yield, which is defined as the volume of water that an unconfined aquifer yields by gravity per unit surface area per unit decline in water level. Basin-specific estimates of storage coefficient were computed by dividing estimates of areal transmissivity by the hydraulic diffusivity (defined as the ratio  $T/S$ ) for each basin.

### ***3.11 Estimation of Average Saturated Thickness of Regolith in the Piedmont***

A total of 120 wells were selected in the Piedmont Physiographic Province of Georgia to estimate average saturated thickness of the regolith. Well casing typically is installed through the regolith to the top of the unweathered bedrock to prevent collapse of unconsolidated regolith into open boreholes during the well-drilling process. As a result, a sufficiently good approximation for the regolith thickness can be obtained from the casing depth in a bored well. Estimates of the saturated thickness of regolith were determined by subtracting the depth to water from the depth of casing. Because similarities exist between the Piedmont of Georgia and the Piedmont of North Carolina, this information can be used with reasonable limit of confidence.

## 4 RESULTS AND DISCUSSION

### 4.1 *Distribution of Recession Indices in the Study Area*

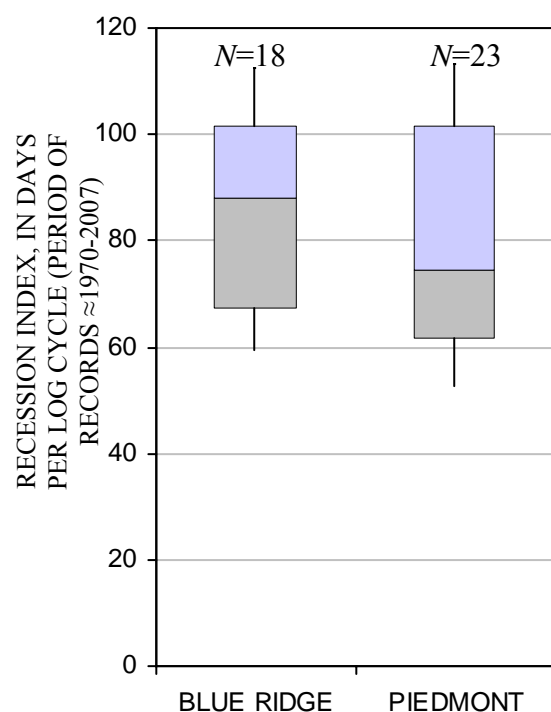
Recession index is the time (number of days) which is required for groundwater discharge to drop through one log cycle after the baseflow recession started. Recession indices for the basins in the Blue Ridge and Piedmont Physiographic Provinces were determined using the RECESS program.

Figure 4-1 shows that most recession indices for the basins in the Blue Ridge range from 56.1 to 122.1 days per log cycle. The median recession index for the basins in the Blue Ridge is 87.8 days per log cycle. Furthermore, this figure shows that the values of recession indices for the basins in the Piedmont range from 40.1 to 138.6 days per log cycle. The median recession index for the Piedmont is 74.5 (days/log cycle).

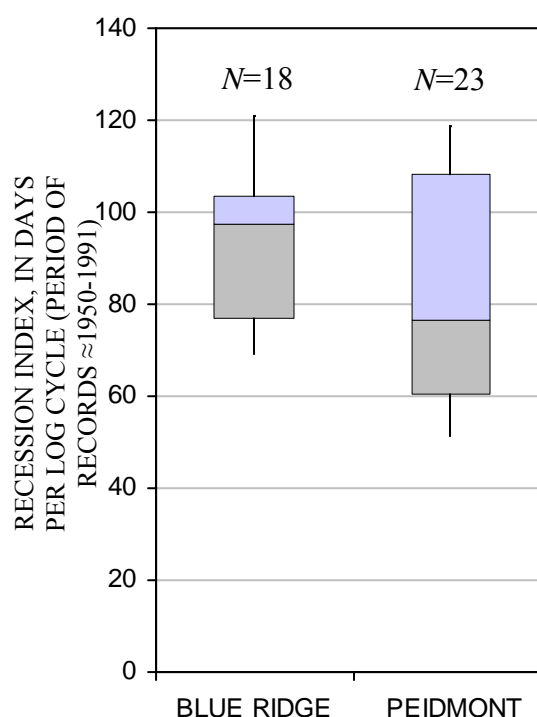
As can be seen by a comparison of Figure 4-1 and Figure 4-2, there is a similarity between distribution of median basin recession index estimated in the present study and that of the Rutledge and Mesko (1996). In the present study, the recession indices were estimated for the period of about 1970-2007, whereas those of Rutledge and Mesko were estimated from the period of about 1950-1991. These figures reveal that the median basin recession index is slightly larger in the Blue Ridge than the Piedmont. The larger median recession index in the Blue Ridge could be due to the large values of basin-relief and smaller median transmissivity in this province. Finally, these figures demonstrate a larger variation in recession indices in the Piedmont than that of the Blue Ridge Province.

It has been said that large  $T$  and  $S$  are favorable for potentially high groundwater yields. However, direct opposition of the effects of  $T$  and  $S$  on recession index (Equation 1) may

complicate the interpretation of groundwater potential from the recession indices. For example, an unconfined aquifer with a large storage coefficient and a good degree of reliability of water supply usually exhibits a large recession index. Nonetheless, aquifers with small transmissivities and small well yields also generally exhibit large recession indices.



**Figure 4-1: Distribution of median recession index for basins in the Blue Ridge and Piedmont Provinces (period of records: ≈1970-2007) (estimated in the present study)**

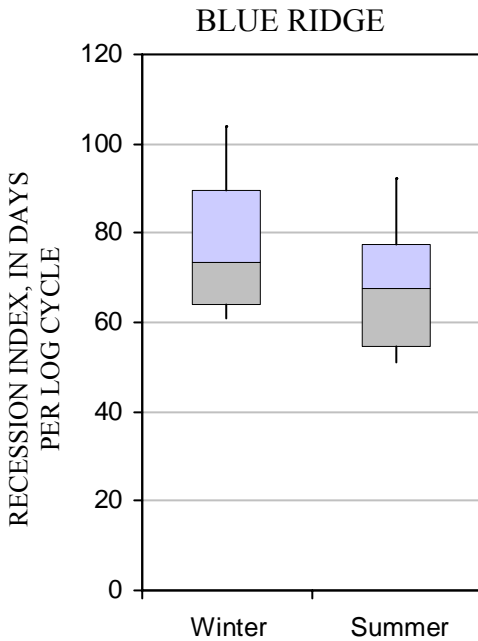


**Figure 4-2: Distribution of median recession index for the same area (period of records: ≈1950-1991) (estimated by Rutledge and Mesko, 1996)**

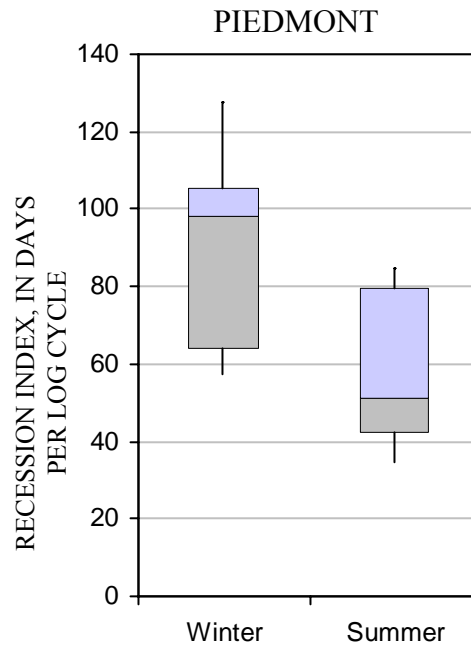
## **4.2 Distribution of Recession Indices for the Summer and Winter**

### ***Months in the Study Area***

The recession indices for the summer (May, June, and July) and winter months (December, January, and February) were determined to examine the effect of evapotranspiration on the recession index.



**Figure 4-3: Distribution of median recession index for summer and winter for basins in the Blue Ridge (1950-2007)**

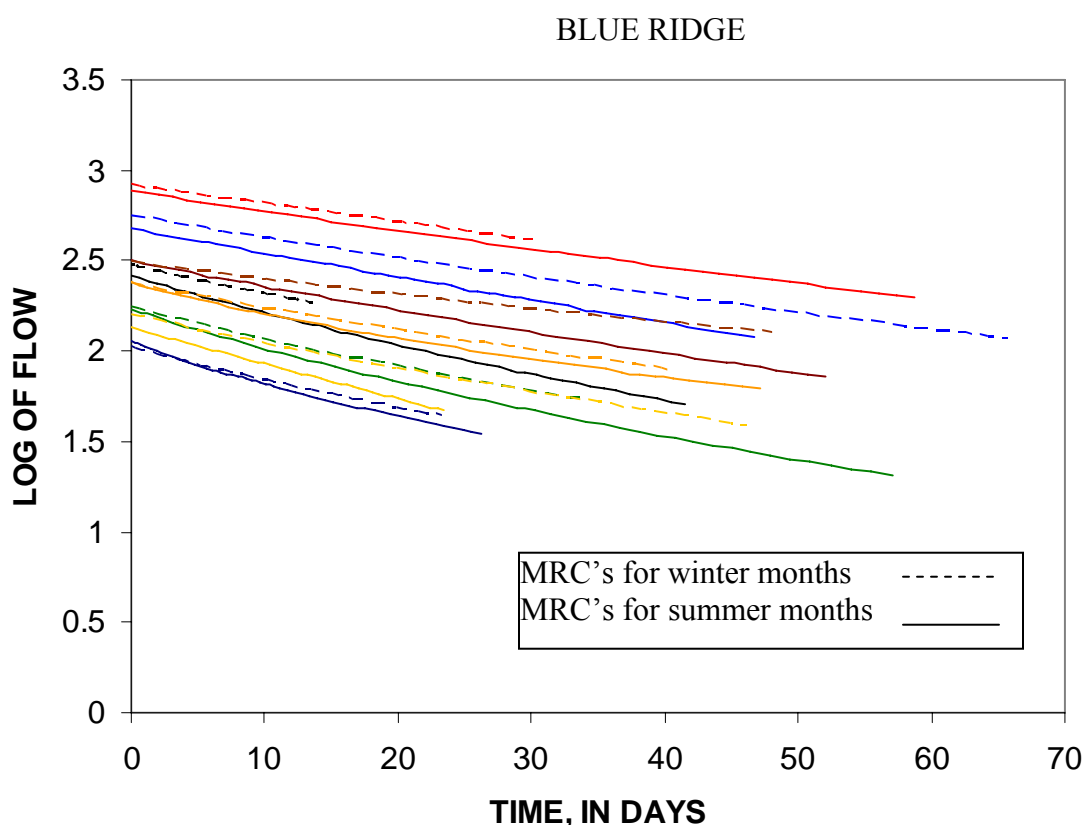


**Figure 4-4: Distribution of median recession index for summer and winter for basins in the Piedmont (1950-2007)**

As shown in Figure 4-3 and Figure 4-4, summer months show smaller median recession index than winter months for basin in the Blue Ridge and Piedmont Provinces. This is because evapotranspiration was higher in summer than it was in winter. In other words, the recession indices in winter were less affected by evapotranspiration. Furthermore, as can be seen in these figures, there is a larger difference between the median recession indices for summer and winter for basins in the Piedmont than that of the Blue Ridge. This larger variability in recession indices for the Piedmont can be attributed to the spatial heterogeneity of regolith. Based on available data, the regolith in the Piedmont exhibits larger spatial heterogeneity than that of the Blue Ridge.

### 4.3 MRCs for Summer and Winter for Basins in the Study Area

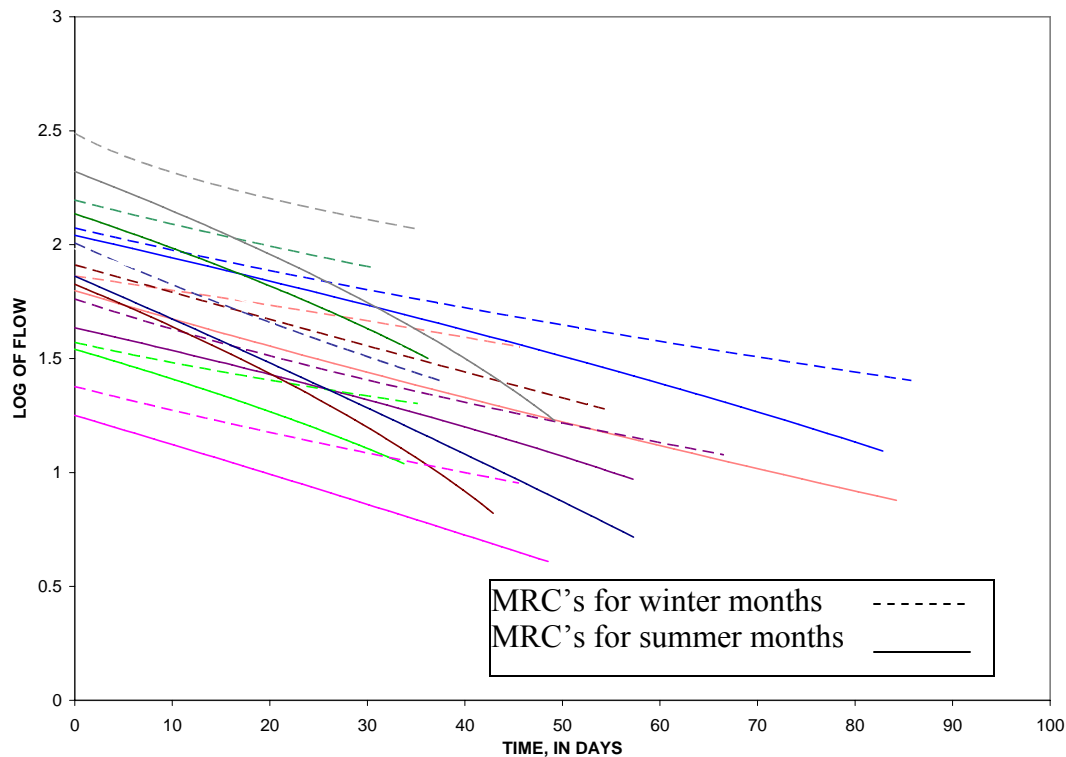
MRCs for summer and winter for basins in the study area were generated using the RECESS program. Additionally, the CURV was used to display multiple MRCs on the same graph.



**Figure 4-5: Typical master recession curves of groundwater discharge in the Blue Ridge for the summer and winter months**

Figure 4-5 shows the winter and summer MRCs for basins in the Blue Ridge. Each pair of MRCs corresponds to a gauging station. MRCs for summer months are represented by solid lines whereas MRCs for the winter months are represented by dash lines. According to our results, MRCs for summer months show steeper slopes compared to those from winter months. This is because evapotranspiration is higher in summer than it is in winter. Furthermore, the MRCs for the Blue Ridge are concave. According to Rutledge and Mesko (1996), there are two

reasons for the concavity of MRCs in the Blue Ridge. One is that a decrease in aquifer transmissivity occurs as the zone of saturation becomes thinner; and the second is high relief of basins in the Blue Ridge. For basins with high relief, as groundwater levels decline, some upland tributaries go dry, thus increasing the average distance from the stream to hydraulic divide. In addition, according to Rorabaugh equation (1966) (Equation 1), there is a positive relation between aquifer half-width and the recession index. Therefore, this parallel increase in aquifer half-width and recession index contributes to the concavity of MRCs in the Blue Ridge.



**Figure 4-6: Typical master recession curves of groundwater discharge in the Piedmont for the summer and winter months**

Figure 4-6 shows the winter and summer MRCs for basins in the Piedmont. According to our results, MRCs for summer months show steeper slopes compared to those from winter months. This is because evapotranspiration is higher in summer than it is in winter. Furthermore, the MRCs for the Piedmont are convex. In the view of some experts (Singh, 1969; Daniel, 1976;



and Rutledge and Mesko, 1996), the convexity of MRCs in the Piedmont may be caused by downward leakage to deeper groundwater flow systems. For example, Nelms and Ahlin (1993) dated groundwater in the Piedmont and Coastal Plain aquifers of Virginia, and they found that the water present in the Piedmont aquifer is younger than that of Coastal Plain. Additionally, the authors identified the presence of young waters (post-1945) at depths greater than 200 (ft) in fractured-rock aquifer of the Piedmont. Moreover, according to Swain (1993), zones of high well yield exist at depths between 350 and 650, which suggest the potential for deep groundwater flow.

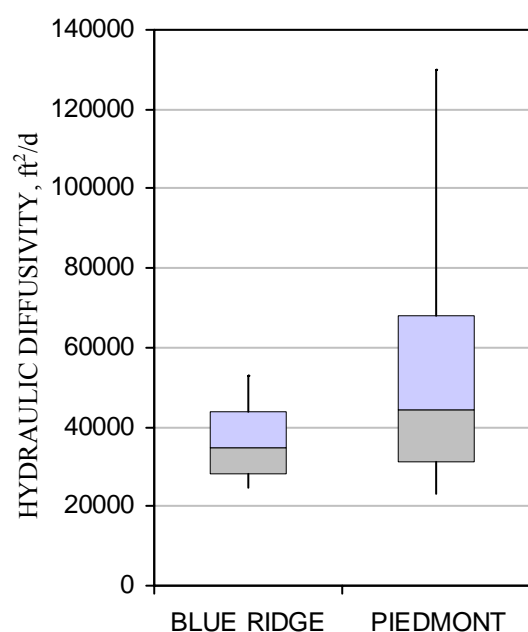
Convexity also may be caused by the decrease in storage coefficient as the water table declines. According to several investigators (Olmsted and Hely, 1962; Stewart, 1962; and Barksdale and others, 1943), the storage coefficient of rocks in the Piedmont decreases with depth. In addition, according to (Equation 1), there is a positive relation between storage coefficient and the recession index. Therefore, this parallel decrease in  $S$  and  $K$  contributes in the convexity of MRCs in the Blue Ridge.

#### ***4.4 Diffusivity and Drainage Density***

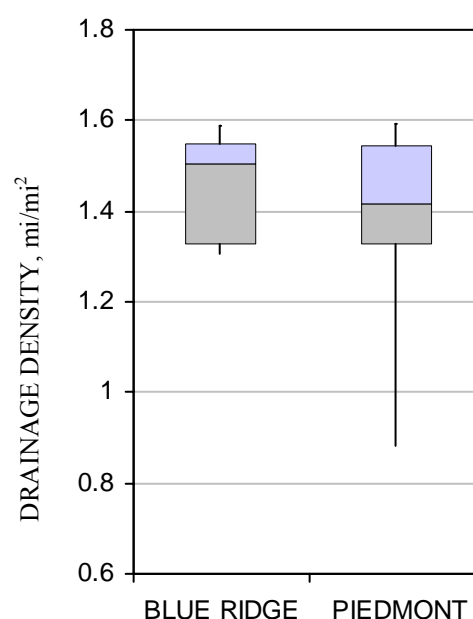
Hydraulic diffusivity is referred to as the ratio of transmissivity to the storage coefficient. Rorabaugh (1964) developed Equation 1 that relates the slope of master recession curve to the aquifer-half width, transmissivity, and storage coefficient of the groundwater reservoir. Values of drainage density were estimated using Equation 12.

Table 4-1 and Table 4-2 display the statistical summary of areal diffusivity and drainage density for the 42 study basins organized by the physiographic province. The areal hydraulic diffusivity in the study area ranges from 23,200 to 74,100 ( $\text{ft}^2/\text{d}$ ), with median values of 35,000, 44,200, and 39,000 ( $\text{ft}^2/\text{d}$ ) for basins in the Blue Ridge, Piedmont, and study area respectively.

These values are in agreement with values reported in the literature (Olmsted and Hely, 1962; Hely and Olmsted, 1963; Trainer and Watkins, 1974; and Nelms and others, 1997). As shown in Table 4-1, median diffusivity of the Piedmont aquifer is greater than that of the Blue Ridge due to the larger areal transmissivity in the Piedmont aquifer.



**Figure 4-7: Distribution of median areal diffusivity for basins in the Blue Ridge and Piedmont Provinces (period of records: ≈1970-2007)**



**Figure 4-8: Distribution of median drainage density for basins in the Blue Ridge and Piedmont Provinces**

**Table 4-1: Statistical summary of areal diffusivity for basins in the Blue Ridge and Piedmont Provinces**

Region	Number of basins	Median	25 <sup>th</sup> Percentile	75 <sup>th</sup> Percentile	10 <sup>th</sup> Percentile (Min)	90 <sup>th</sup> Percentile (Max)
Areal Diffusivity (ft <sup>2</sup> /day)						
Blue Ridge	19	35,000	28,000	43,600	24,800	53,000
Piedmont	23	44,200	31,500	68,000	23,300	129,700
Study area	42	39,000	27,800	52,700	23,200	74,100

**Table 4-2: Statistical summary of drainage density for basins in the Blue Ridge and Piedmont Provinces**

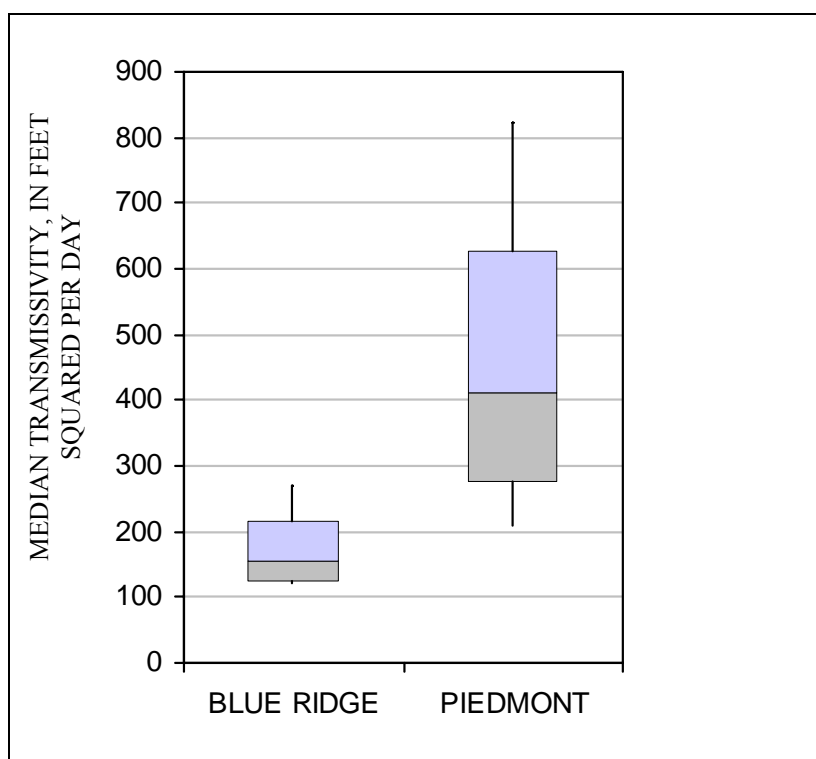
<b>Region</b>	<b>Number of basins</b>	<b>Median</b>	<b>25<sup>th</sup> Percentile</b>	<b>75<sup>th</sup> Percentile</b>	<b>10<sup>th</sup> Percentile (Min)</b>	<b>90<sup>th</sup> Percentile (Max)</b>
<b>Drainage Density (mi/mi<sup>2</sup>)</b>						
Blue Ridge	19	1.5	1.4	1.5	1.3	1.6
Piedmont	23	1.4	1.3	1.5	0.9	1.6
Study area	42	1.4	1.3	1.5	1.1	1.6

As shown in Table 4-2, the median basin, drainage density is slightly higher in the Blue Ridge Physiographic Province. The higher basin relief in the Blue Ridge is probably a major cause of higher median drainage density in this physiographic province.

Figure 4-7 and Figure 4-8, display a wider range in areal diffusivity and drainage density for basins in the Piedmont than that of the Blue Ridge Physiographic Province. The large range in areal diffusivity and drainage density for basins in the Piedmont is attributed to spatial variability of rock and regolith type, regolith thickness, geologic structures, and climate.

#### **4.5 Areal Transmissivity**

Transmissivity is the volume of water flowing through a unit width of the saturated thickness of the aquifer under a unit of hydraulic gradient, measured in (ft<sup>2</sup>/d). The areal transmissivity for the basins in the study area was estimated using Equation 10 (Olmsted and Hely, 1962, p. A18).



**Figure 4-9: Distribution of median transmissivities for the basins in the Piedmont and Blue Ridge**

**Table 4-3: Statistical summary of areal transmissivity for basins in Blue Ridge and Piedmont Physiographic Provinces**

Region	Number of basins	Median	25 <sup>th</sup> Percentile	75 <sup>th</sup> Percentile	10 <sup>th</sup> Percentile (Min)	90 <sup>th</sup> Percentile (Max)
Areal Transmissivity (ft <sup>2</sup> /d)						
Blue Ridge	19	150	120	210	120	270
Piedmont	23	410	270	630	210	820
Study area	42	260	160	420	120	660

As can be seen in Table 4-3, the median areal (basin-specific) transmissivity for basins in the Blue Ridge Physiographic Province ranges from 120 to 270 (ft<sup>2</sup>/d), with a median value of 150 (ft<sup>2</sup>/d). Table 4-4 and 4-5 show that this estimated median value is consistent with the values of 180, 120, and 100 (ft<sup>2</sup>/d) for the Blue Ridge regolith, obtained by Stewart (1964), Hoos (1990), and Seaton and Burbey (2005, p. 308), respectively. Furthermore, Table 4-3 and 4-6 show that median areal transmissivity for basins in the Piedmont ranges from 210 to 820 (ft<sup>2</sup>/d), with a median value of 410 (ft<sup>2</sup>/d), which is consistent with the values reported by Kasper

(1989). In general, the median basin transmissivity in the Piedmont is roughly twice as much as it is in the Blue Ridge. The student's t-test indicates that the mean values of areal transmissivity in the Piedmont are significantly greater than that of the Blue Ridge at the 95% confidence interval. Likewise, F-test indicates that the variance values of areal transmissivity in the Piedmont are greater than that of the Blue Ridge at the 95% confidence interval. The large values of transmissivity obtained for the Piedmont regolith may be attributed to the thick regolith, low values of basin relief, and voids that develop as a result of fracturing, foliation, weathering, and fractured quartz veins in the saprolite.

As can be seen in Figure 4-9, the values of areal transmissivity obtained for basins in the Piedmont show extremely high variability. This is because the regolith in the Piedmont is more heterogeneous than it is in the Blue Ridge Physiographic Province. Composition, thickness, foliation, fracture density, and the extent of saprolite differ from one rock type to another. Therefore, in relatively short distances, transmissivity of regolith material vary widely.

**Table 4-4: Statistical summary of areal (basin-specific) transmissivity for selected basins in the Blue Ridge and Piedmont Physiographic Provinces**

<b>Source citation</b>	<b>Number of basins</b>	<b>Location of basins</b>	<b>Method</b>	<b>Representative value of <math>T</math> (ft<sup>2</sup>/day)</b>
Hoos (1990)	8	Blue Ridge aquifer in Tennessee	Stream hydrographs analysis and $T/S$ ratio	120
Present study	19	Blue Ridge aquifer in North Carolina and Georgia	Using Rorabaugh, Olmsted and Helys' equations and hydrologic analysis of stream hydrographs	150
Present study	23	Piedmont aquifer in North Carolina and Georgia	Using Rorabaugh, Olmsted and Helys' equations and hydrologic analysis of stream hydrographs	410

**Table 4-5: Statistical summary of site-specific estimates of transmissivity for regolith wells in the Blue Ridge Province**

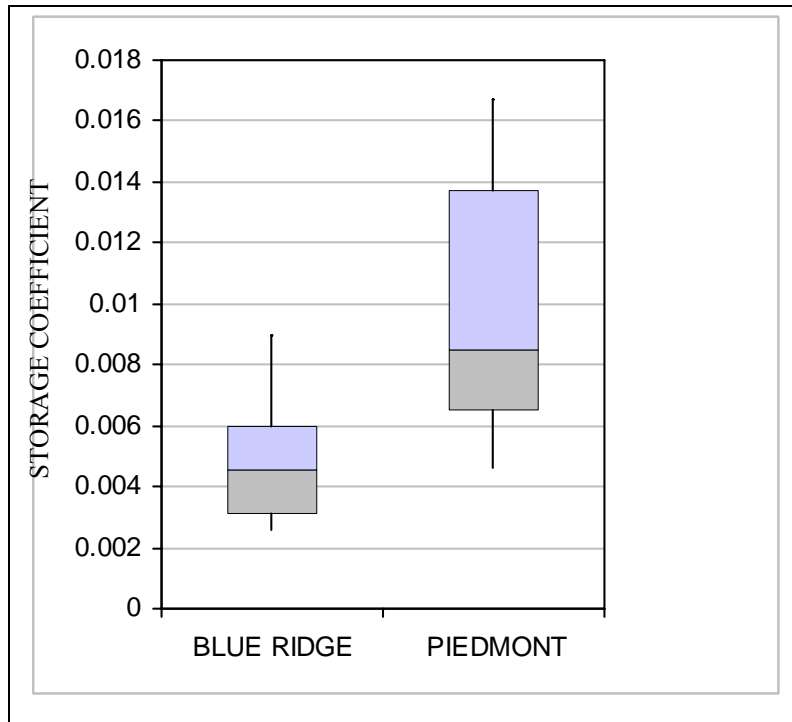
Source citation	Number of site-specific estimates	Location of site	Method	Mean transmissivity (ft <sup>2</sup> /day)
Seaton and Burbey (2005, p. 308)	4	Floyd County, southwest Virginia	Hvorslev	100
Campbell (2005)	About 7	Wells at Bent Creek Experimental Forest, North Carolina	Cooper Jacob Straight Line	485
Stewart (1964)	About 11	Wells in Georgia Nuclear Laboratory, Dawson County, Georgia	Aquifer test	180

**Table 4-6: Statistical summary of site-specific estimates of transmissivity for regolith well in the Piedmont Province**

Source citation	Number of site-specific estimates	Location of site	Method	Transmissivity
Kasper (1989)	About 10	Piedmont aquifer in South Carolina	Aquifer test	Ranging from 270 to 320

## 4.6 Storage Coefficient

The storage coefficient is defined as the volume of water released from an aquifer per unit surface area per unit change in head. In an unconfined aquifer the water yield is from gravity; whereas, in a confined aquifer the release of water is relative to compaction of the aquifer. Basin-specific estimates of storage coefficient were computed by dividing estimates of areal transmissivity by the hydraulic diffusivity (defined as the ratio  $T/S$ ) for each basin.



**Figure 4-10: Distribution of median storage coefficients for basins in the Blue Ridge and Piedmont Physiographic Provinces (period of records: ≈1970-2007)**

**Table 4-7: Statistical summary of median storage coefficient for basins in the Blue Ridge and Piedmont Physiographic Provinces**

Region	Number of basins	Median	25 <sup>th</sup> Percentile	75 <sup>th</sup> Percentile	10 <sup>th</sup> Percentile (Min)	90 <sup>th</sup> Percentile (Max)
Storage Coefficient						
Blue Ridge	19	0.005	0.003	0.006	0.002	0.009
Piedmont	23	0.009	0.006	0.014	0.005	0.017
Study area	42	0.007	0.004	0.010	0.003	0.015

Table 4-7 shows that the median storage coefficient for basins in the Blue Ridge ranges from 0.002 to 0.009, with a median value of 0.005. This estimated median value is in agreement with the values of 0.0045, 0.005, and 0.005 for the Blue Ridge regolith, obtained by Stewart (1964), Trainer and Watkins (1975, p. 40), and Campbell (2005) respectively (Tables 4-8 and 4-10). Moreover, Table 4-7 shows that the median storage coefficient for basins in the Piedmont ranges from 0.005 to 0.017, with a median value of 0.009. The estimated median value in the present study is in agreement with representative values of 0.01, 0.012-0.023, and 0.01 obtained

by Trainer and Watkins (1975, p. 40), Kasper (1989), and CH2M HILL (2005), respectively (Tables 4-9 and 4-11). The student's t-test indicates that the mean values of areal storage coefficient in the Piedmont are significantly greater than that of the Blue Ridge at 90% confidence interval. Furthermore, F-test indicates that the variance values of areal storage coefficient in the Piedmont are greater than that of the Blue Ridge at 95% confidence interval.

Figure 4-10 displays great differences in basin-specific storage coefficient of the regolith material within the Piedmont Physiographic Province. In general, the regolith in the Piedmont is made of silt and clay zones whose permeability varies widely, depending upon the thickness and lateral extend of the zones and type of material form which the regolith is derived. In addition, fracture density, foliation, and the presence of other geologic structures vary widely from one basin to another.

**Table 4-8: Statistical summary of site-specific storage coefficient for regolith wells in the Blue Ridge Physiographic Province of the study area**

<b>Source citation</b>	<b>Number of site-specific estimates</b>	<b>Location of basin or station</b>	<b>Method name</b>	<b>Mean Storage coefficient</b>
Campbell (2005)	3	Wells at Bent Creek Experimental Forest, North Carolina	Theis	0.005
Stewart (1964)	8	Wells in Georgia Nuclear Laboratory, Dawson County, Georgia	Aquifer test	0.0045

**Table 4-9: Statistical summary of site-specific storage coefficient for regolith wells in the Piedmont Physiographic Province of the study area**

<b>Source citation</b>	<b>Number of Site-specific estimates</b>	<b>Location of basin or site</b>	<b>Method name</b>	<b>Mean storage coefficient</b>
CH2M HILL	About 25	Marietta, Georgia	Aquifer Test	0.01
Kasper (1989)	About 10	Piedmont aquifer in South Carolina	Aquifer test	Ranged between 0.012 to 0.023



**Table 4-10: Statistical summary of basin-specific storage coefficient for selected basins in the Blue Ridge**

<b>Source citation</b>	<b>Number of basins</b>	<b>Location of basins</b>	<b>Method</b>	<b>Representative value of storage coefficient</b>
Hoos (1990)	2	Blue Ridge aquifer in Tennessee	Hydrologic analysis of concurrent water table and stream hydrographs of the basins	0.01
Trainer and Watkins (1975, p. 40)	1	Blue Ridge aquifer in upper Potomac river basin, Virginia	Base-runoff recession curves	0.005
Present study	19	Blue Ridge aquifer in North Carolina and Georgia	Using Rorabaugh, Olmsted and Helys' equations and hydrologic analysis of stream hydrographs	0.005

**Table 4-11: Statistical summary of areal storage coefficient for selected basins in the Piedmont**

<b>Source citation</b>	<b>Number of basins</b>	<b>Location of basins</b>	<b>Method</b>	<b>Representative value of storage coefficient</b>
Trainer and Watkins (1975, p. 40)	1	Piedmont aquifer in Potomac river basin, Virginia	Base-runoff recession curves	0.01
Present study	23	Piedmont aquifer in North Carolina and Georgia	Using Rorabaugh, Olmsted and Helys' equations and hydrologic analysis of stream hydrographs	0.009

#### ***4.7 Relation between Areal Transmissivity and Storage Coefficient for Basins in the Study Area***

Figure 4-11 and Figure 4-12 demonstrate a positive relation between areal transmissivity and storage coefficient and the coefficient of determination  $R^2$  is 0.295. As can be seen in Figure

4-11, the areal transmissivity and storage coefficient for basins in the Blue Ridge are clustered together; whereas, those of the Piedmont are scattered. It appears regolith in the Piedmont is more heterogeneous than it is in the Blue Ridge, due to the variable grain size, mineralogy, thickness, lateral extent of saprolite, land use and the presence of fractures and other geologic structures. .

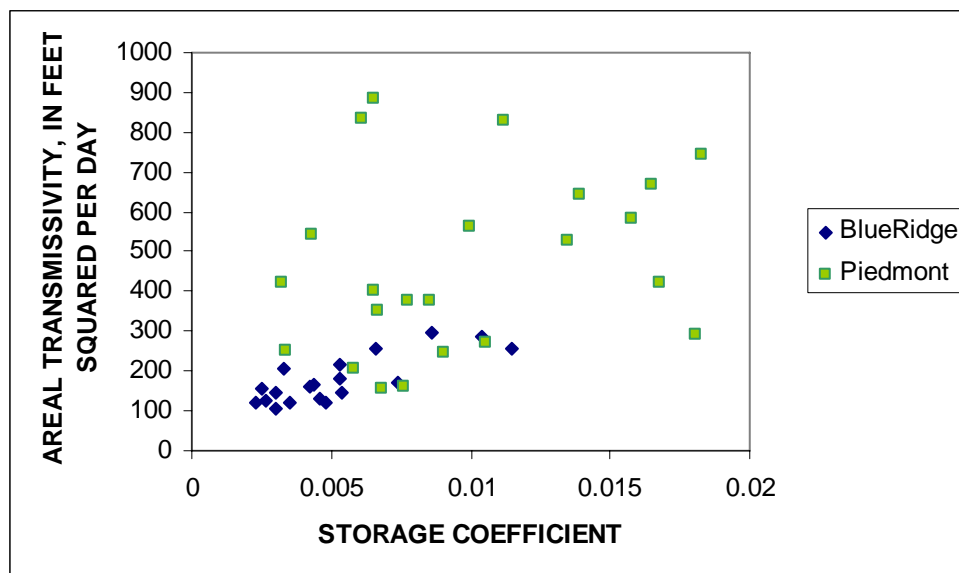


Figure 4-11: Relation between areal transmissivity and storage coefficient in the study area

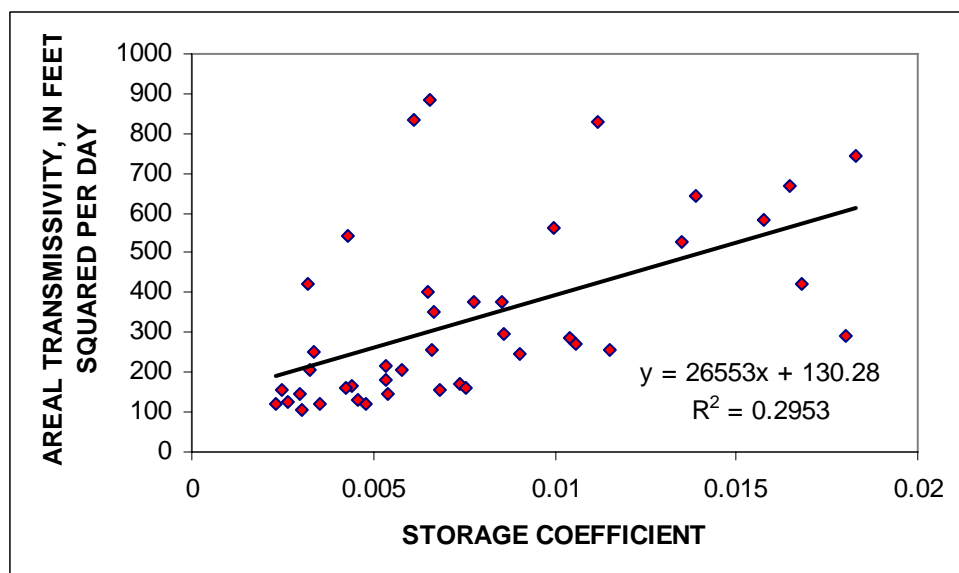


Figure 4-12: Regression equation and plot of transmissivity and storage coefficient in the study area

## 4.8 Relation between Areal Storage Coefficient and Basin Relief

Figure 4-13 reveals an inverse relation between median storage coefficient and basin relief for basins in the Blue Ridge. Figure 4-14 shows that this relation for basins in the Piedmont is slight and can be obscured somewhat by scatter. The inverse relation between storage coefficient and basin relief is shown in Equation 10 and Equation 1.

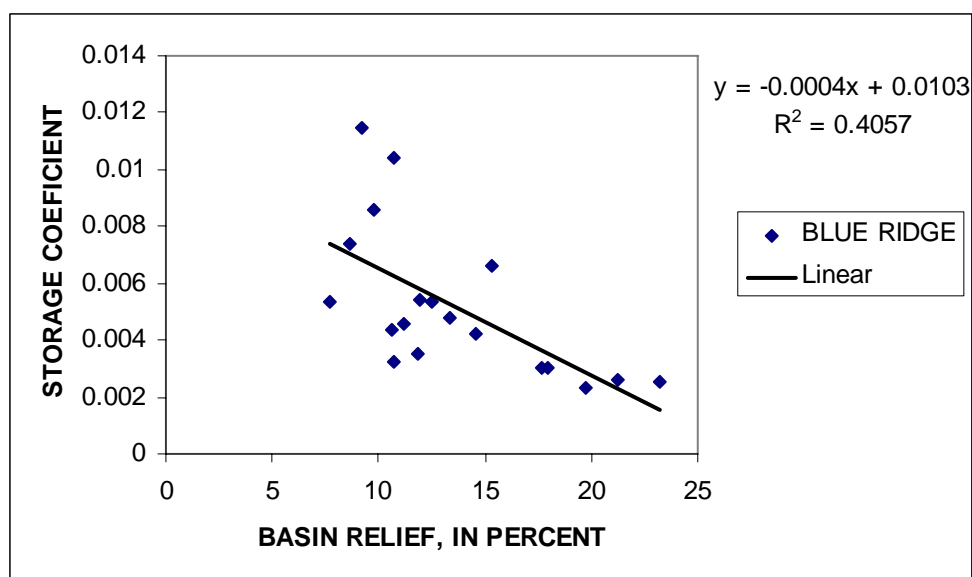


Figure 4-13: Relation between basin relief and median storage coefficient for basins in the Blue Ridge

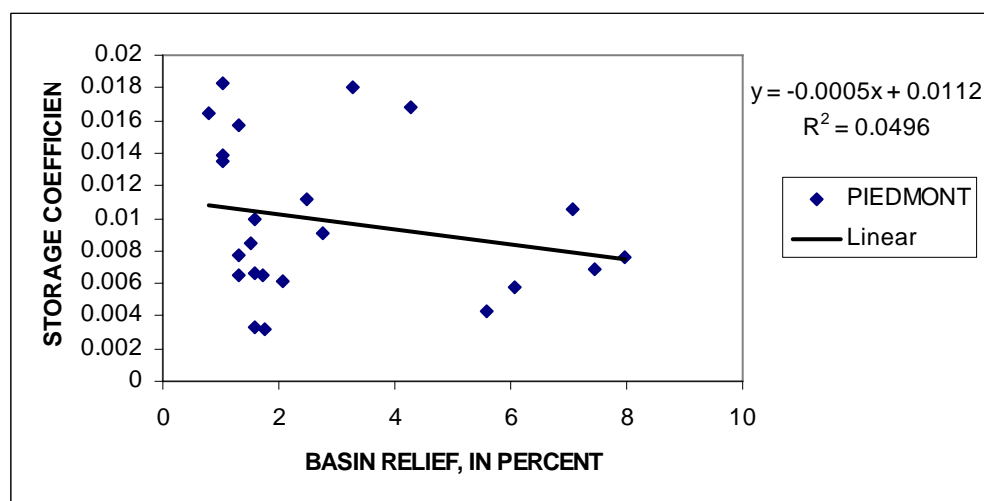


Figure 4-14: Relation between basin relief and median storage coefficient for basins in the Piedmont

## **4.9 Relation between Areal Storage Coefficient and Median Basin, Recession Index**

Figure 4-15 demonstrates that there is a positive relation between median recession index and median storage coefficient in the Blue Ridge. Furthermore, this positive relation is shown in the Rorabaugh equation. As can be seen in Figure 4-16, in the Piedmont the relation between  $K$  and  $S$  is slight and can be obscured somewhat by scatter. Furthermore, the result of this study reveals no apparent relation between the areal transmissivity and recession index for basins in the study area. It is well known that under the most favorable circumstances, large  $T$  and  $S$  favor a sustainable water supply for a given region. However, the opposite effects of these variables on the recession index complicate the interpretation of groundwater recession as they relate to water-supply potential (Rutledge and Mesko, 1996). For example, an increase in  $T$  is likely to cause a drop in recession index, whereas increases in recession index are often caused by increases in  $S$ . Nevertheless, as a rule of thumb, a large recession index often indicates the best conditions for water supply. That is, storage coefficient outweighs transmissivity from the point of view of the water-supply potential (Rutledge and Mesko, 1996). In addition, basins with higher relief seem to exhibit larger recession indices.

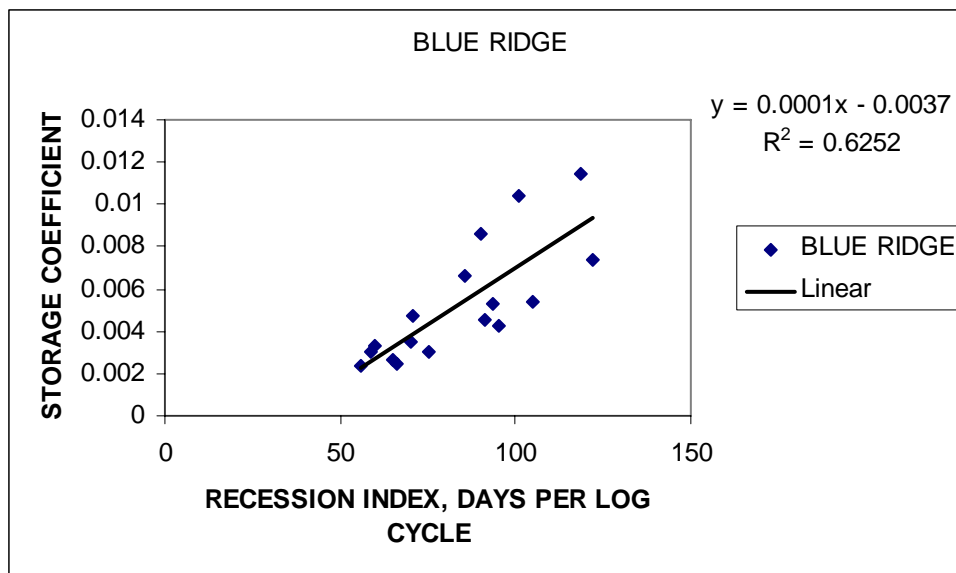


Figure 4-15: Relation between recession index and the storage coefficient for the basins in the Blue Ridge.

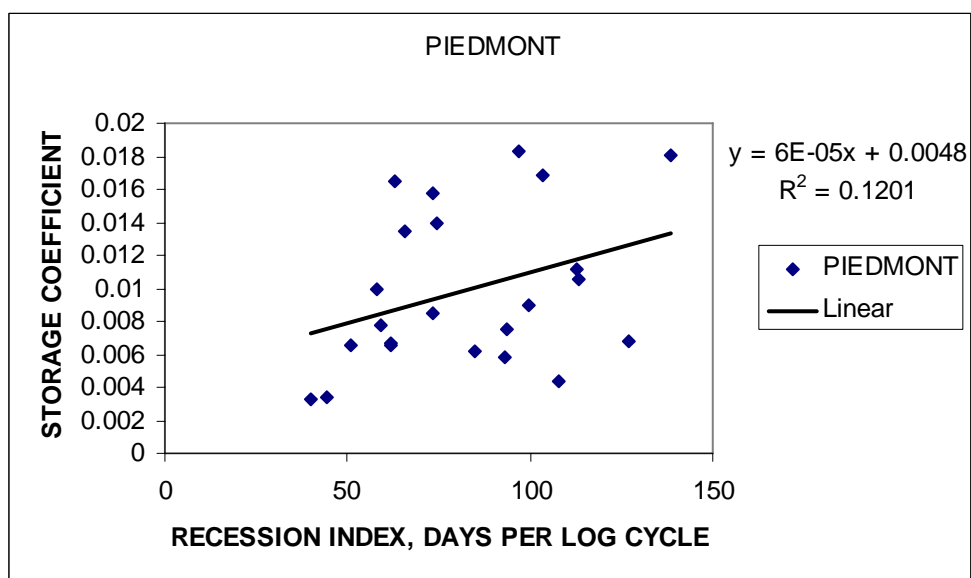
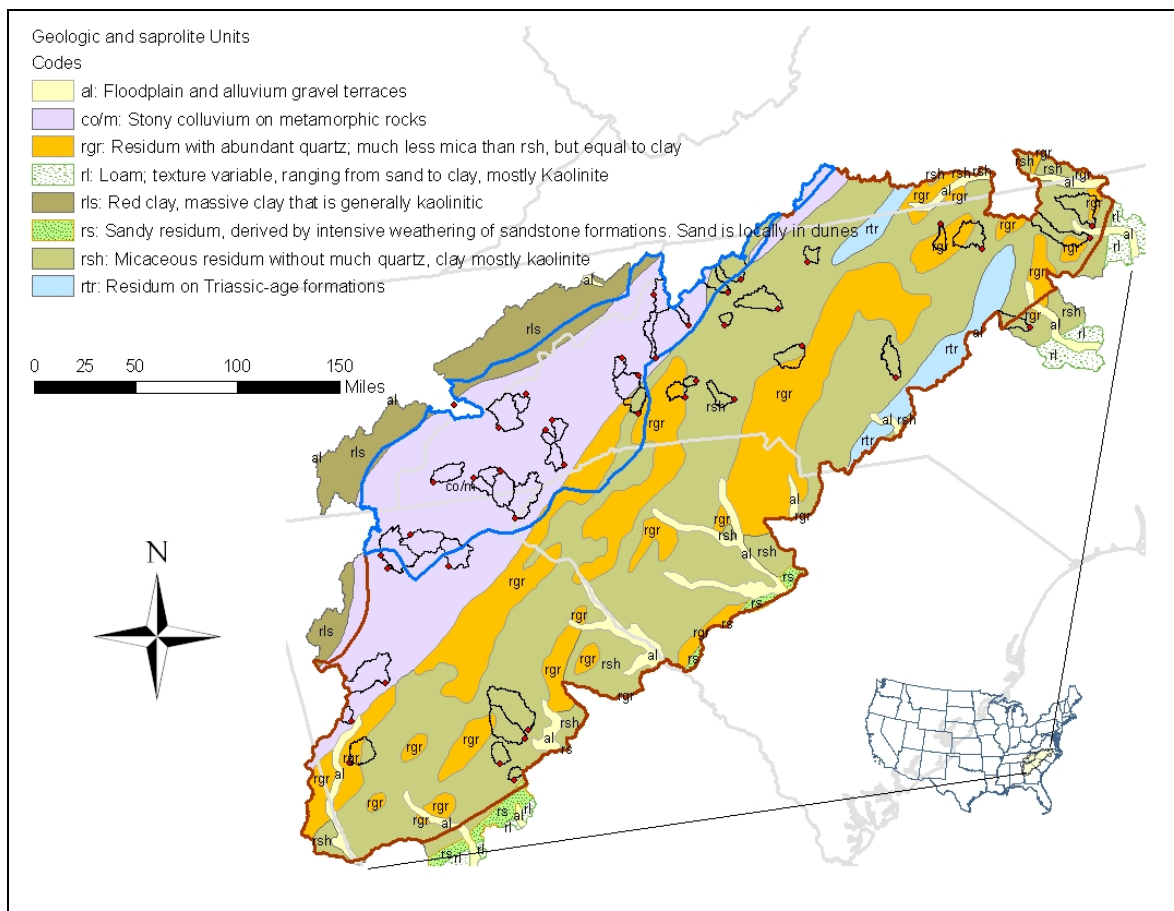


Figure 4-16: Relation between recession index and the median storage coefficient of the basins in the Piedmont

#### 4.10 Two Contrasting Types of Regolith within the Piedmont

In general, two predominant rock types in the Piedmont Physiographic province are gneiss and schist. Regolith, is the product of chemical and mechanical weathering of the underlying bedrock, and thus may reflect the texture of the rock from which it was formed.

Therefore, as can be seen in Figure 4-17, the weathering product of granite-derived metamorphic rocks may be quartz-rich and sandy-textured; whereas, rocks poor in quartz and rich in feldspar and other soluble minerals form a more clayed saprolite. In other words, during weathering, the amphiboles and feldspars are dissolved leaving behind clay size particle of iron oxides and aluminum silicates.

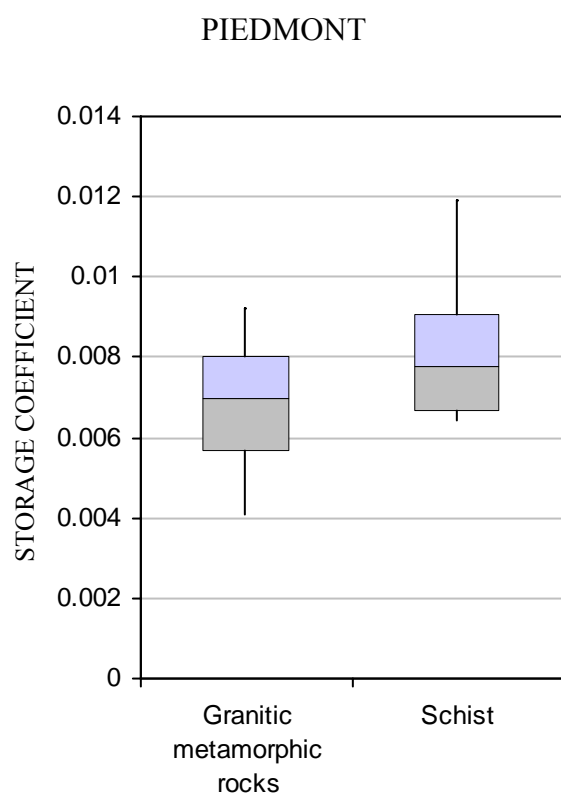


**Figure 4-17: Surficial geology of the study area. The original data set was generated from a U.S. Geological Survey 1:7,500,000-scale map of surficial geology published as part of the U.S. Geological Survey National Atlas map series.**

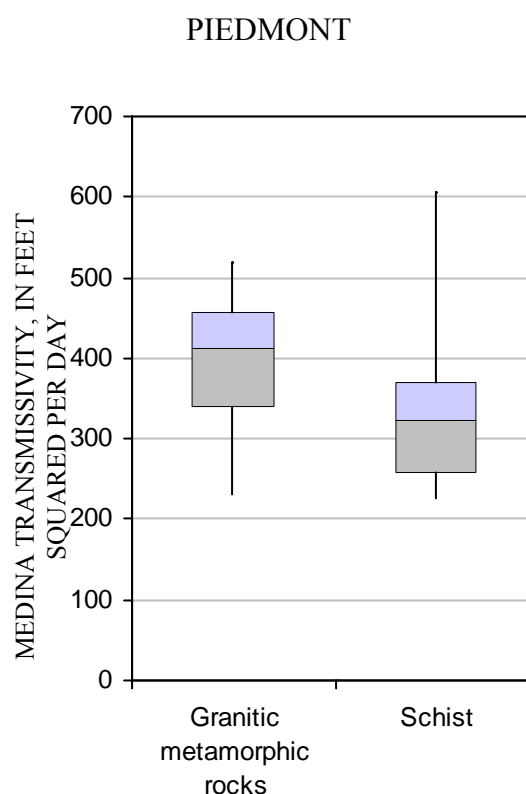
Figure 4-18 and Figure 4-19 shows the distribution of median basin  $T$  and  $S$  within two contrasting media: (1) granular saprolite and (2) clayey saprolite. The clay-rich saprolite is capable of storing water readily, but transmitting it slowly; in contrast, the granitic granular saprolite has a relatively low storage capacity, but it is capable of transmitting water readily.

Furthermore, these figures suggest greater variability in the hydraulic parameters obtained for the clay-rich regolith within the Piedmont Physiographic Province.

According to our results, basins in the Piedmont exhibit a larger variability in their hydraulic characteristics. In general, this larger variability in basin hydraulic characteristics for the Piedmont can be attributed to the spatial heterogeneity and the anisotropy (foliation) of the regolith.



**Figure 4-18: Relation between the lithologic characteristics of regolith and storage coefficient in basins for the Piedmont**



**Figure 4-19: Relation between the lithologic characteristics of regolith and transmissivity in basins for the Blue Ridge**

#### 4.10.1 Soil map of the study area

An important factor is the infiltration capacity of the soil, which depends not only on soil parameters derived from weathering of the bedrock, but on land use and land cover. Figure 4-20

shows that there is a greater spatial variability of infiltration rates for soils in the Piedmont than there is in the Blue Ridge. In other words, basins in the Piedmont are considerably more likely than basins in the Blue Ridge to be underlain by more than one soil or geologic type. This characteristic may cause a larger spread among hydraulic values for basins in the Piedmont than that of the Blue Ridge.

With the exception of a few locations, it seems that the dominant soil type in the study area is type B (moderate infiltration rate). However, soil type C with a slow infiltration rate is probably prominent in such areas as Atlanta Metropolitan Area, north central Georgia, and eastern sections of Piedmont provinces of South and North Carolina. Furthermore, this figure reveals that the areas around the boundary between the Blue Ridge and the Valley and Ridge provinces are made mainly of soil type D (very slow infiltration rates).

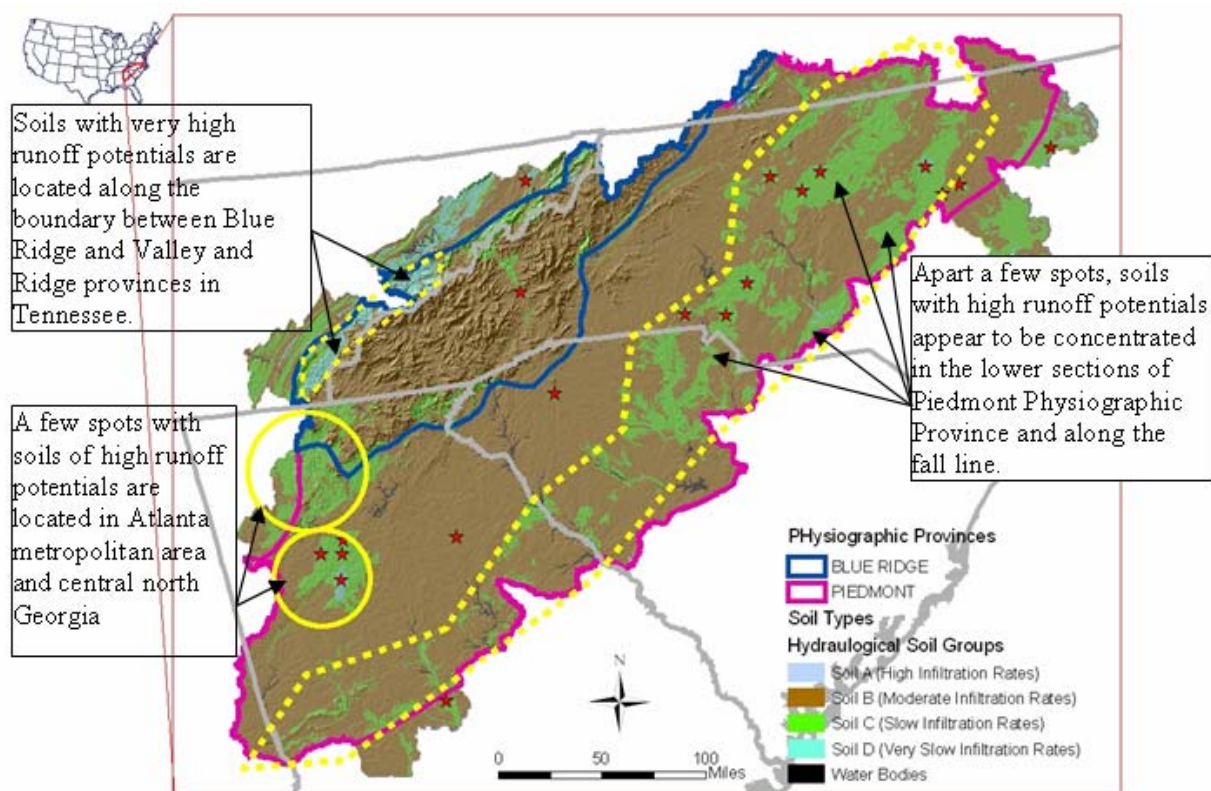


Figure 4-20: Soil map of the study area



#### 4.10.2 Land use map of the study area

Figure 4-21 suggests a more heterogeneous land use for basins in the Piedmont than that of the Blue Ridge. In other words, basins in the Piedmont are considerably more likely than basins in the Blue Ridge to be underlain by more than one land use type. This characteristic may cause great differences in hydraulic parameters of regolith material within short distances. Moreover, this figure shows a high concentration of agricultural and urban areas in the mid-section of Piedmont as well as a cluster of agricultural and urban areas around the Asheville, North Carolina region in the Blue Ridge Physiographic Province.

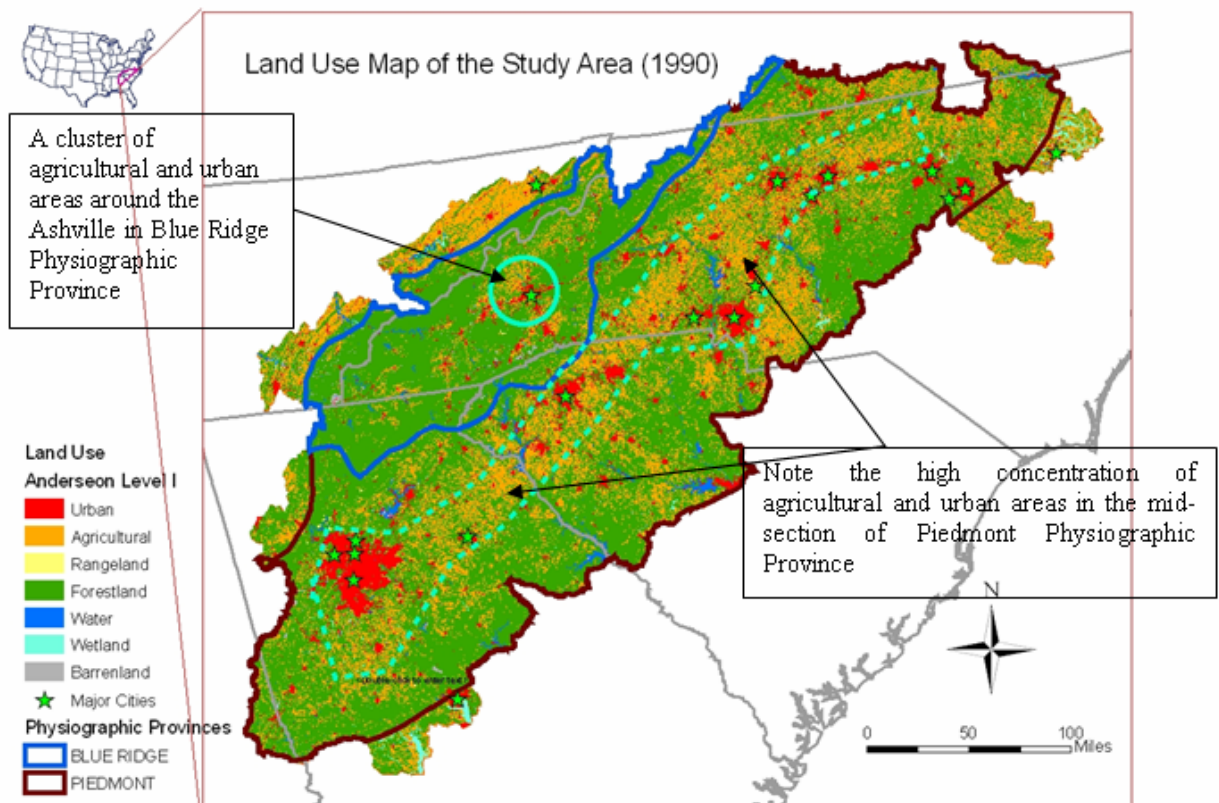


Figure 4-21: Land use map of the study area

#### **4.10.3 Curve number map of the study area**

The curve number (CN) is a hydrologic parameter used to describe the storm water runoff potential for drainage area. The curve number is a function of land use, soil type, and soil moisture (e.g. if  $CN = 0$ , no runoff; if  $CN = 100$ , 100% runoff). Figure 4-22 indicates a wide range in curve number within the Piedmont than in the Blue Ridge, depending upon the land use and infiltration capacity of soil in the area.

Atlanta Metropolitan area appears to be a high-runoff-potential area due to the dominant soil type C and urban land use type. Furthermore, north central Georgia exhibits high potentials for runoff due to the dominant soil type C. Moreover, mid- and eastern-sections of Piedmont provinces of South and North Carolina appear to have high runoff potential due to the dominant soil type C and agricultural land use. This figure suggests that there are more potential for runoff in the Piedmont province than that of the Blue Ridge province.

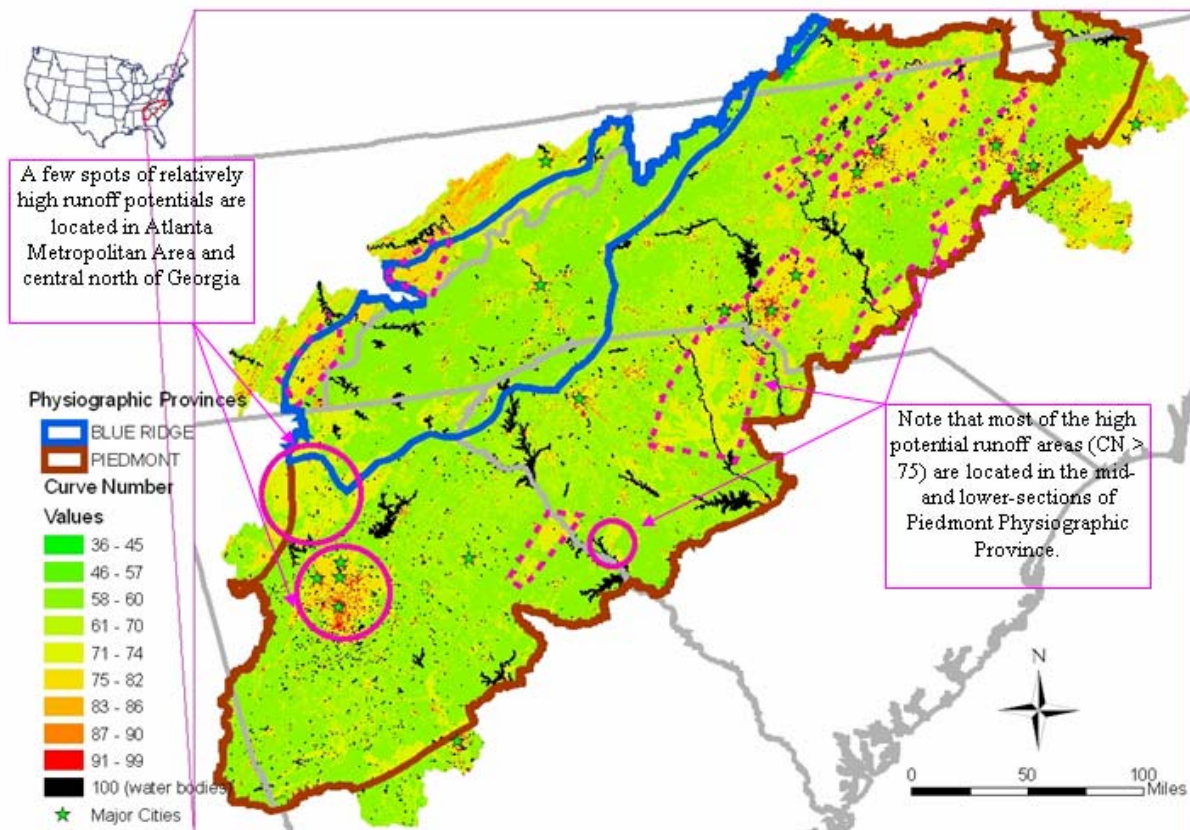
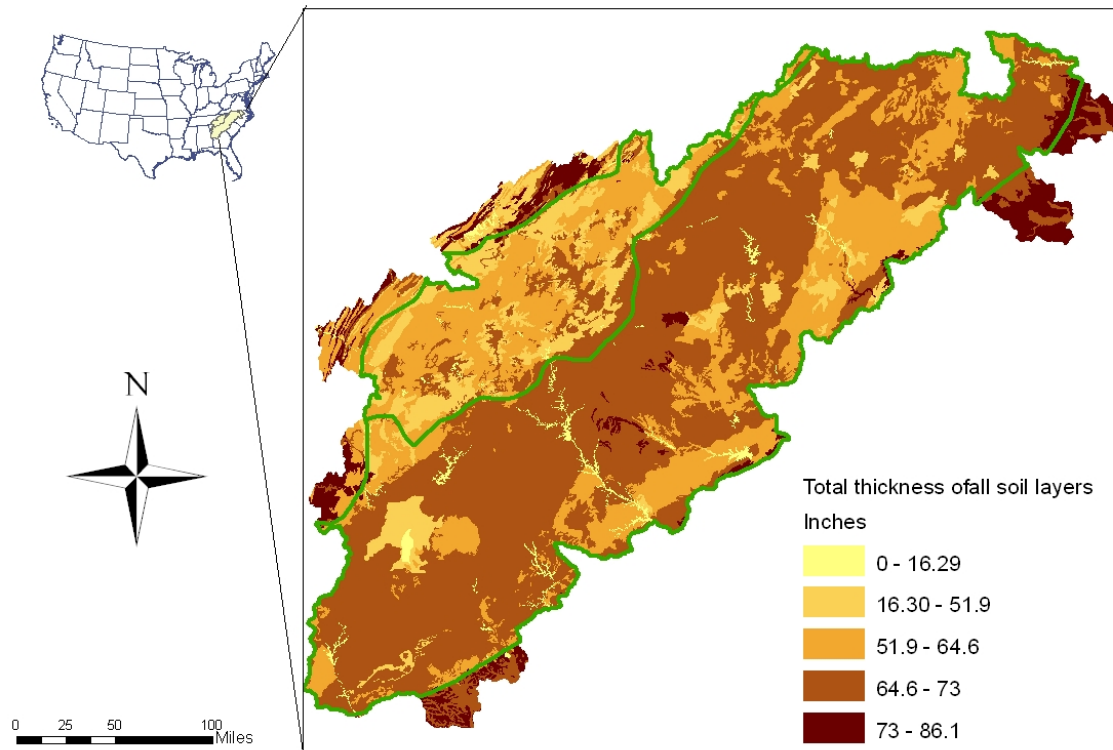


Figure 4-22: Curve number map of the study area

#### 4.10.4 Distribution of soil thickness in the study area

Figure 4-23 indicates that thickness of soil-saprolite zone throughout of the study area is highly variable. Furthermore, in the Piedmont, massive granitic-type metamorphic rocks tend to have thin soils; whereas, schist has thicker soil. The thickness of soil-saprolite zone varies according to the type of parent rock, topography, climate, and geologic history. In the Piedmont, soil-saprolite zone is usually thicker beneath broad upland area than in valleys. In the Blue Ridge, the soil-saprolite zone is thin beneath the ridges, mountains, and steep hillsides due to the higher rate of weathering at the regolith-bedrock boundary. As a result, soil tends to be thin or absent, and bedrock outcrops can be found at land surface.



**Figure 4-23: Spatial distribution of soil thickness in the study area. This map shows that the total thickness of soil layers is greater in the Piedmont than the Blue Ridge**

#### ***4.11 Average Saturated Thickness of Regolith Associated with Each Physiographic Province***

Table 4-12 shows statistical summary of data on total regolith thickness (well casing), water level from land surface and estimated saturated thickness of regolith for wells in the Piedmont of “Greater Atlanta Region”. The average saturated thickness of regolith for 120 wells in this area is 46 (ft).

**Table 4-12: Data on depth of well casing, depth to water, and estimated saturated thickness for wells in the "Greater Atlanta Region" (Piedmont Physiographic Province) (original data are from Cressler and other, 1983)**

<b>Well number</b>	<b>Total regolith thickness (ft)</b>	<b>Water level below land surface (static head) (ft)</b>	<b>Regolith saturated thickness (ft)</b>
3CC4	42	27	15
4CC4	75	20	55
4CC14	48	30	18
4DD6	63	34	29
10CC13	88	25	63
10DD36	82	38	44
11CC3	200	8	192
11CC10	38	10	28
11CC11	54	30	24
8EE5	81	25	56
9EE2	89	15	74
10FF2	70	16	54
6AA1	64	20	44
6BB1	35	5	30
6BB2	69	30	39
6BB3	116	40	76
6BB5	40	8	32
7AA3	66	40	26
7AA4	113	30	83
7AA12	98	10	88
7AA13	77	40	37
7AA15	69	30	39
7AA17	95	22	73
7AA19	78	35	43
7BB10	107	40	67
7BB15	72	30	42
7BB16	71	50	21
7BB31	75	20	55
7CC2	57	35	22
8AA5	56	20	36
8AA7	53	40	13
8AA8	41	27	14
8AA9	80	35	45
9Z1	78	10	68
9Z2	109	20	89
9AA1	81	20	61
12DD11	73	20	53
7CC4	40	10	30
7CC6	49	30	19
8DD3	99	45	54
9BB7	47	35	12
9BB15	87	40	47
9CC16	78	50	28
9CC17	82	30	52
10BB5	95	16	79
10BB11	60	20	40
10BB12	41	4	37
10BB15	78	28	50
10BB16	101	32	69
10BB19	33	10	23

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10BB20	116	30	86
12HH5	40	6	34
12HH6	36	12	24
12JJ1	57	22	35
13JJ6	47	12	35
14JJ3	60	38	22
7CC3	48	17	31
8CC2	67	20	47
8CC6	98	20	78
8CC7	56	3.8	52.2
8CC8	78	3.8	74.2
8DD10	40	10	30
9CC21	44	31	13
9DD2	46	15	31
9EE3	133	40	93
10CC17	65	20	45
10DD1	49	25	24
10DD21	106	65	41
10DD23	95	60	35
10DD37	62	6	56
10DD45	113	30	83
10DD59	42	28	14
10DD64	40	15	25
10FF5	61	35	26
10GG11	52	18	34
11GG6	22	10	12
12GG5	67	20	47
12GG3	96	40	56
12GG4	78	20	58
13EE3	44	20	24
13EE6	133	50	83
13FF2	50	35	15
13FF7	89	10	79
13GG12	68	8	60
13GG14	74	40	34
14EE4	76	40	36
14EE5	56	20	36
14FF4	85	22	63
14HH1	79	56	23
14HH7	66	20	46
15JJ1	110	35	75
15JJ16	106	90	16
15JJ17	101	90	11
15JJ19	180	93	87
11BB5	124	7	117
11BB9	30	4	26
11BB11	70	40	30
11BB12	138	20	118
12AA1	70	40	30
12BB9	46	10	36
12DD4	90	18	72
13AA1	102	37	65
13AA2	73	18	55
13AA3	43	21	22
13AA4	60	25	35
14CC13	77	30	47

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12CC31	45	30	15
12CC33	85	15	70
13CC54	74	2	72
13DD56	103	60	43
13DD63	86	45	41
13DD66	23	10	13
13DD67	58	15	43
13DD77	44	15	29
13DD81	72	50	22
13DD85	37	15	22
13DD87	84	20	64
13DD88	40	30	10
14DD45	105	34	71
14DD58	35	10	25
<b>Average</b>			<b>46</b>

Table 4-13 displays the estimated saturated thickness of regolith in the Blue Ridge (presented by William and others, 2005). The average saturated thickness of regolith in this area is 32 (ft).

**Table 4-13: Estimated saturated thickness of regolith in Floyd County, Virginia (the Blue Ridge Physiographic Province) (Modified from William and others, 2005)**

<b>Shallow well number</b>	<b>Saturated Thickness (ft)</b>
W-05	35
W-06	49
W-08	26
P-1	19
<b>Average</b>	<b>32</b>

#### ***4.12 Average Estimates of the Hydraulic Conductivity for the Blue Ridge and Piedmont***

The estimated transmissivities for basins in the study area can be verified by the estimation of hydraulic conductivity values using the following equation:

$$T = bK \quad \text{Equation 16}$$

where  $T$  is transmissivity (ft<sup>2</sup>/d),  $b$  is saturated thickness of the aquifer (ft), and  $K$  is the hydraulic conductivity (ft/d). In the case of a multilayer aquifer, its total transmissivity is estimated by the summation of transmissivity of each of the layers:

$$T = \sum_{i=1}^n T_i \quad \text{Equation 17}$$

As can be seen in Table 4-14, average hydraulic conductivity for shallow aquifer in the Blue Ridge and Piedmont Physiographic Provinces are 4.6, and 8.9 (ft/d), respectively. These values are in agreement with those reported in the literature (Table 4-16 and Table 4-17)

**Table 4-14: Average estimated hydraulic conductivity for shallow aquifer system in the study area**

<b>Physiographic Province</b>	<b>Saturated thickness (ft)</b>	<b>Medina specific-estimate of transmissivity (ft/d)</b>	<b>Estimated hydraulic conductivity (ft/d)</b>
Piedmont	46	410	8.9
Blue Ridge	32	150	4.6

**Table 4-15: Hydraulic conductivities for unconsolidated sediments and rocks**

<b>Material</b>	<b>Hydraulic Conductivity (ft/d)</b>
Igneous and Metamorphic Rocks	$10^{-8} - 200$
Clay	$10^{-7} - 10^{-3}$
Carbonate Rocks	$10^{-3} - 10^4$
Clean Sand	$1 - 10^3$
Gravel	$5 \times 10^2 - 10^4$

**Table 4-16: Statistical summary of hydraulic conductivity for regolith in the Blue Ridge**

<b>Source citation</b>	<b>Number of site-specific estimates</b>	<b>Location of basin or station</b>	<b>Method type</b>	<b>Hydraulic conductivity, in foot per day</b>
Seaton and Burbey (2005, p. 308)	4	Floyd County, southwest Virginia	Hvorslev	3.56
Campbell (2005)	About 3-7	Wells at Bent Creek Experimental Forest, North Carolina	Slug test	3.3



**Table 4-17: Statistical summary of hydraulic conductivity for the regolith in the Piedmont**

<b>Source citation</b>	<b>Number of site-specific estimates</b>	<b>Location of basin or station</b>	<b>Method Name</b>	<b>Representative value of Hydraulic conductivity (ft/d)</b>
Huffman and others (2006, p. 11)	9	Wells at the Lake Wheeler Road research station, North Carolina	Slug test	(mean value) 1.33
Huffman and other (2006, p.37)	5	Wells at the Langtree Peninsula research station, North Carolina	Slug test	(mean value) 2.52
Huffman and other (2006, p. 61)	4	Wells at the Upper Piedmont research station, North Carolina	Slug test	(mean value) 6.25
Huffman and other (2006, p. 95)	9	Wells at the Bent Creek research station, North Carolina	Slug test	(mean value) 7.5
Fleck and White (1989)	Road cut saprolite	Piedmont aquifer in Clemson University Research Watershed, South Carolina	Klute and Dirksen, 1986	(Median value) 9.4
Fleck and White (1989)	Samples from the watershed saprolite	Piedmont aquifer in Clemson University Research Watershed, South Carolina	Klute and Dirksen, 1986	(median value) 3.9

As can be seen by a comparison of Table 4-14 and Table 4-15, average values of hydraulic conductivity for the Blue Ridge and Piedmont Physiographic Provinces fall in the category of sand. Large values of hydraulic conductivity estimated in this study and those reported in the literature could be due to the presence of fractures and foliation within bedrock and regolith of the study area. Furthermore, Table 4-17 shows great differences in hydraulic conductivity of regolith material within the study area due to the anisotropy of shallow aquifer. In other words, the aquifer in the study area is highly anisotropic (the hydraulic conductivity

varies with direction at any location) because of well-developed foliation and sporadic occurrence of fractures.

Results of this study suggest that the regolith within the study area could be modeled as an anisotropic aquifer due to the presence of quartz veins, subsurface tubes, clay lenses, and foliation. These various subsurface structures within the regolith complicate the modeling of the complete hydrologic system of the basins.

## **4.13 Uncertainties**

### **4.13.1 Uncertainty in estimating the total stream length**

Estimation of areal hydraulic diffusivity and areal transmissivity from equations 9 and 10 may be subject to some degree of uncertainty. This uncertainty arises from geometric factors where aquifer half-width  $a$  is determined using the following equation that is a rearrangement of Equation 12 and 13:

$$a = \frac{A}{2L} \quad \text{Equation 18}$$

This equation can be solved by measuring the total length of all perennial streams  $L$  in the area  $A$  on a 1:100,000 Reach File 3 stream network. Had the total stream length been estimated from a larger scale topographic map (e.g. on a 1:24,000 topographic map), the estimated values of diffusivity and transmissivity would be slightly lower. The magnitude of such uncertainty can be illustrated by comparing aquifer parameters derived from the 1:24,000 USGS topographic map with the 1:100,000 RF3 stream network.

#### **4.13.2      Uncertainty in estimating hydraulic conductivity**

Estimation of hydraulic conductivity from Equation 16 may be subject to uncertainty. This uncertainty is due to the fact that the exact depth of groundwater flow is difficult to define. The aquifer in the study area is a two-part system consisting of regolith and fractured crystalline bedrock. The fractured crystalline part was excluded in estimation of the saturated thickness; as a result, the estimated values of average saturated thickness are only the minimum values. Had the fracture crystalline part been included in the estimation of the saturated thickness, the estimated values of hydraulic conductivities would be lower.

## **5 CONCLUSIONS AND RECOMMENDATIONS**

Quantification of the hydraulic characteristics of shallow aquifer system in the Blue Ridge and Piedmont Physiographic Provinces of the southeastern U.S. is essential for effective development and management of groundwater and surface water resources, as well as reducing the pollution of aquifers and connected surface water. The purpose of this study was (1) to define the hydraulic characteristics of basins in the study area, (2) identify regional differences in these characteristics, and (3) describe the relations among properties of the aquifers and physical properties of the basins.

### **5.1 Conclusions**

This study provides basin-specific estimates of the recession index, diffusivity, transmissivity, and storage coefficient. All of the methods used are appropriate during times when no groundwater recharge is occurring, when all flow is from groundwater discharge, and when the profile of the groundwater head distribution is nearly stable. Moreover, the aquifer is considered to be thick relative to depth of the water table, wide relative to its thickness, and underlain by impermeable material. Furthermore, its side boundaries are vertical and fully penetrating, and the aquifer is uniform, isotropic, and homogeneous. Because the ideal condition may not apply to the entire shallow aquifer, the estimates of hydraulic properties may be imprecise or inaccurate in some areas. Listed below are the objectives and results of this study:

#### **5.1.1 Distribution of recession indices in the study area**

Recession indices for basins in the Blue Ridge range from 56.1 to 122.1 days per log cycle. The median recession index for basins in the Blue Ridge is 87.8 days per log cycle. Values of recession indices for basins in the Piedmont range from 40.1 to 138.6 days per log cycle. The

central tendency of recession indices for the Piedmont is 74.5 (day/log cycle). The median basin recession index is slightly larger in the Blue Ridge than it is in the Piedmont. The larger median recession index in the Blue Ridge could be due to the lower median transmissivity in this province. Moreover, there is a larger variation in recession indices in the Piedmont than that of the Blue Ridge Province.

### **5.1.2 Distribution of recession indices for summer and winter months in the study area**

According to our results, summer months show smaller median recession index than that of winter months for basins in the Blue Ridge and Piedmont Provinces. This is because evapotranspiration is higher in summer than it is in winter. In other words, the recession indices in winter are less affected by evapotranspiration. Furthermore, there is a larger difference between the median recession indices for summer and winter for basins in the Piedmont than there is in the Blue Ridge. This larger variability in recession indices for the Piedmont can result from spatial heterogeneity of regolith in the basins. Based on the available data, regolith in the Piedmont exhibits larger spatial heterogeneity than that of the Blue Ridge.

### **5.1.3 Master recession curves (MRCs) for summer and winter for basins in the study area**

According to our results, MRCs for summer months show steeper slopes compared to those from winter months. This is because evapotranspiration is higher in summer than it is in winter. Furthermore, the MRCs for the Blue Ridge are concave. According to Rutledge and Mesko (1996), there are two reasons for the concavity of MRCs in the Blue Ridge. First, there is a decrease in aquifer transmissivity that occurs as the zone of saturation becomes thinner; and the

second, large relief of basins in the Blue Ridge. In addition, according to Rorabaugh equation (1966) (Equation 1), there is a positive relation between aquifer half-width and the recession index. Therefore, this parallel increase in aquifer half-width and recession index contributes in the concavity of MRCs in the Blue Ridge. In contrast, the MRCs for the Piedmont are convex. In the view of some experts (Singh, 1969; Daniel, 1976; and Rutledge and Mesko, 1996), the convexity of MRCs in the Piedmont may be caused by downward leakage to deeper groundwater flow systems. Convexity also may be caused by the decrease in storage coefficient as the water table declines. In addition, according to Equation 1, there is a positive relation between storage coefficient and the recession index. Therefore, this parallel decrease in storage coefficient and recession index contributes in the convexity of MRCs in the Blue Ridge.

#### **5.1.4 Diffusivity and drainage density**

The areal hydraulic diffusivity in the study area ranges from 23,200 to 74,100 ( $\text{ft}^2/\text{d}$ ), with median values of 35,000, 44,200, and 39,000 ( $\text{ft}^2/\text{d}$ ) for basins in the Blue Ridge, Piedmont, and study area respectively. These values are in agreement with values reported in the literature (Olmsted and Hely, 1962; Hely and Olmsted, 1963; Trainer and Watkins, 1974; and Nelms and others, 1997). The median diffusivity of the Piedmont aquifer is greater than that of the Blue Ridge due to the larger areal transmissivity in the Piedmont aquifer. The median basin, drainage density is slightly higher in the Blue Ridge physiographic province. The larger basin relief in the Blue Ridge is probably a major cause of higher median drainage density in this physiographic province. Moreover, according to our results, there is a wider range in areal diffusivity and drainage density for basins in the Piedmont than that of the Blue Ridge. The large range in areal diffusivity and drainage density for basins in the Piedmont is attributed to spatial variability of rock and regolith type, regolith thickness, geologic structures, and climate.

### **5.1.5 The median areal transmissivity of shallow aquifer system in the study area**

The median areal (basin-specific) transmissivity for basins in the Blue Ridge Physiographic Province ranges from 120 to 270 (ft<sup>2</sup>/d), with a median value of 150 (ft<sup>2</sup>/d). This estimated median value is consistent with the values of 180, 120, and 100 (ft<sup>2</sup>/d) for the Blue Ridge regolith, obtained by Stewart (1964), Hoos (1990), and Seaton and Burbey (2005, p. 308), respectively. Furthermore, the median areal transmissivity for basins in the Piedmont ranges from 210 to 820 (ft<sup>2</sup>/d), with a median value of 410 (ft<sup>2</sup>/d), which are consistent with the values reported by Kasper (1989). In general, the median basin transmissivity in the Piedmont is roughly twice as much as it is in the Blue Ridge. The large values of transmissivity obtained for the Piedmont regolith may be attributed to the thick regolith, low values of basin relief, and voids that develop as a result of fracturing, foliation, weathering, and fractured quartz veins in the saprolite.

### **5.1.6 The areal storage coefficient of shallow aquifer system in the study area**

The median storage coefficient for basins in the Blue Ridge ranges from 0.002 to 0.009, with a median value of 0.005. This estimated median value is in agreement with the values of 0.0045, 0.005, and 0.005 for the Blue Ridge regolith, obtained by Stewart (1964), Trainer and Watkins (1975, p. 40), and Campbell (2005) respectively. Moreover, the median storage coefficient for basins in the Piedmont ranges from 0.005 to 0.017, with a median value of 0.009. The estimated median value in the present study is in agreement with representative values of

0.01, 0.012-0.023, and 0.01 obtained by Trainer and Watkins (1975, p. 40), Kasper (1989), and CH2M HILL (2005), respectively.

#### **5.1.7 Relation between areal transmissivity and storage coefficient for basins the in the study area**

The relation between areal transmissivity and storage coefficient is noticeably positive. Furthermore, on the plot of  $T$  versus  $S$ , the areal transmissivity and storage coefficient for the basins in the Blue Ridge are clustered together; whereas, those of the Piedmont are scattered. It appears that the regolith in the Piedmont is more heterogeneous than that in the Blue Ridge, due to the variable grain size, mineralogy, thickness, lateral extent of saprolite, land use and the presence of fractures and other geologic structures.

#### **5.1.8 Relation between areal storage coefficient and basin relief**

There is an inverse relation between median storage coefficient and relief for basins in the Blue Ridge. This inverse relation for basins in the Piedmont is slight and can be obscured somewhat by scatter. The inverse relation between storage coefficient and basin relief is shown in Equation 10 and Equation 1.

#### **5.1.9 Relation between areal storage coefficient and median basin, recession index**

There is a strong positive relation between median recession index and the median storage coefficient in the Blue Ridge and the coefficient of determination  $R^2$  is 0.625. In contrast, in the Piedmont the relation between  $K$  and  $S$  is slight and can be obscured somewhat by scatter.



Furthermore, storage coefficient outweighs transmissivity from the point of view of the water-supply potential.

#### **5.1.10 Regional differences in the study area**

According to our results, the values of hydraulic properties obtained for basins in the Piedmont show extremely high variability. This is because the regolith in the Piedmont is more heterogeneous than it is in the Blue Ridge Physiographic Province. Composition, thickness, foliation, fracture density, and extent of saprolite differ from one rock type to another. Therefore, in relatively short distances, hydraulic values of regolith material vary widely. In general, regolith in the Piedmont is made of silt and clay zones whose permeability varies widely, depending upon the thickness and lateral extend of the zones and type of material form which the regolith is derived. In addition, fracture density, foliation, and the presence of other geologic structures vary widely from one basin to another.

#### **5.1.11 Two contrasting types of regolith within the Piedmont and the distribution of $T$ and $S$ in them**

There are two contrasting media in the Piedmont: (1) granular saprolite and (2) clayey saprolite. The clay-rich saprolite is capable of storing water readily, but transmitting it slowly; in contrast, the granitic granular saprolite has a relatively low storage capacity, but capable of transmitting water readily. Furthermore, results of this study suggest greater variability in the hydraulic parameters obtained for the clay-rich regolith within the Piedmont Physiographic Province.

### **5.1.12      Distribution of soil, land use type, and curve numbers in the study area**

The thickness of soil-saprolite zone through of the study area is highly variable. In the Piedmont, massive granitic-type metamorphic rocks tend to have thin soils; whereas, schist has thicker soil. Furthermore, there is a greater spatial variability of infiltration rates for soils in the Piedmont than it is in the Blue Ridge. Moreover, the land use map of the study area suggests a more heterogeneous land use for basins in the Piedmont than that of the Blue Ridge. The curve number map also indicates a wide range in curve number values within the Piedmont than that of the Blue Ridge, depending upon the land use and infiltration capacity of soil in the area. In general, basins in the Piedmont are considerably more likely than basins in the Blue Ridge to be underlain by more than one land use and soil type. This characteristic may cause great differences in hydraulic parameters of regolith material within short distances.

### **5.1.13      Average estimates of hydraulic conductivity for shallow aquifer in the study area**

According to our results, average hydraulic conductivities for the shallow aquifer in the Blue Ridge and Piedmont Physiographic Provinces are 4.6, and 8.9, respectively. These values are in agreement with those reported in the literature (Table 4-16 and Table 4-17). Average values of hydraulic conductivity for the Blue Ridge and Piedmont Physiographic Provinces fall in the category of sand. Large values of hydraulic conductivity estimated in this study and those reported in the literature could be due to the presence of fractures and foliation within bedrock and regolith of the study area. In general, the aquifer in the study area is highly anisotropic (the

hydraulic conductivity varies with direction at any location) because of well-developed foliation and sporadic occurrence of fractures.

Results of this study suggest that the regolith within the study area could be modeled as an anisotropic aquifer due to the presence of fractures, quartz veins, subsurface tubes, clay lenses, and foliation. These various subsurface structures within the regolith complicate the modeling of the complete hydrologic system of the basins.

## **5.2 Recommendations**

Although the aquifer system in the study area is simple in the overall view; it is extremely complex in detail. In the overall view it is a two-part regolith-fractured crystalline rock aquifer system. The regolith provides the bulk of the water storage within the Blue Ridge and Piedmont and slowly feeds water downward into fractures in the bedrock. These fractures form an intricate interconnected network of pipelines through which some groundwater moves. In contrast, another component flows through the regolith parallel to the bedrock surface. In detail, the regolith and bedrock are anisotropic heterogeneous medium whose hydraulic characteristics vary widely from one site to another site.

Site characterization at the local and regional scale often is inadequate due to the complexity of interaction between groundwater and surface water in this region. More studies are needed to use surface and subsurface technologies at a variety of sites throughout the region to delineate fractures. Future work is needed to determine more accurately the direction and rate of groundwater movement in these aquifers. For example, geochemical traces, isotopic studies, and GIS can be used to determine the groundwater flowpaths and to delineate fractures. Furthermore, a more feasible approach is to conduct comprehensive hydrogeologic studies on selected local

areas which are most representative of land use, soil type, geology and hydrology and then transfer the knowledge from them to similar regional hydrogeologic areas.

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## 7 APPENDIXES

**Appendix 1. Characteristics of recession analysis and master recession curve coefficients for selected basins in the Blue Ridge and Piedmont Physiographic Provinces.**

Station number	Period of analysis	Number of recession segments used	Kmin (d/log cycle)	Median recession index (d/log cycle)	Kmax (d/log cycle)	Range of MRC		Coefficients of MRC		
						MinLogQ	MaxLogQ	A	B	C
02337500	1970-2006	36	62.3	112.9	294.5	1.057	1.906	-7.4015	-101.6149	220.6168
02337000	1970-2006	27	24.8	61.8	101.4	1.54	2.528	-12.194	-14.993	115.8031
02333500	1970-2006	33	51.1	103.4	192.4	1.51	2.725	-0.5365	-107.5675	297.1398
03558000	1960-1996	33	60.3	118.6	289.9	2.109	3.074	55.942	-417.5274	754.874
02177000	1970-2006	41	56.4	90.1	201.9	2.085	3.08	11.559	-155.3621	368.868
03550000	1970-2006	31	39.1	70.6	229	1.595	2.474	36.133	-228.5338	344.2557
03504000	1970-2006	40	39.8	85.7	252	1.643	2.419	80.454	-428.0837	564.7739
03500000	1970-2006	34	50.3	93.5	253.9	1.892	2.837	35.261	-271.9527	487.7097
03439000	1970-2006	44	41.7	100.1	349.3	1.719	2.546	70.518	-422.0434	617.4359
03512000	1970-2006	25	39	65.2	227.6	1.966	2.844	68.899	-420.2605	637.9479
03455500	1970-2006	32	34.1	66.1	233.2	1.23	2.11	55.131	-261.3705	306.0457
03456500	1970-2006	46	27.9	56	207.4	1.293	2.323	54.951	-284.649	364.7152
02149000	1970-2006	36	54.9	122.1	361	1.449	2.329	23.072	-224.4919	397.6486
03463300	1970-2006	42	28.5	58.5	220.3	1.254	2.248	47.992	-247.7156	314.3304
02138500	1970-2006	31	38.1	59.9	191.5	1.379	2.225	54.321	-277.9914	349.5919
02111000	1970-2006	36	43.4	91.5	227.7	0.987	1.867	45.792	-240.9507	290.2078
03479000	1970-2006	30	32.9	70.3	229.9	1.343	2.283	47.076	-268.4779	367.5683
02152100	1970-2006	32	52.4	127.1	239.7	1.23	2.073	57.129	-338.9896	457.257
02143500	1970-2006	10	72.6	84.9	308.7	1.327	1.975	70.265	-341.8974	401.1655
02142000	1960-1995	47	51.5	92.9	241.2	0.98	1.762	44.616	-220.8583	250.6303
02118500	1970-2006	20	58.8	138.6	351.4	1.37	2.38	97.723	-554.7438	766.7047
02111500	1970-2006	25	66.4	122	512.6	1.449	2.335	95.266	-527.0767	711.3091
02114450	1970-2006	21	48.1	99.5	273.3	1.026	1.687	72.323	-317.4254	329.6801
02128000	1970-2006	18	33.7	65.6	134.9	1.129	1.911	-25.538	7.3474	79.2656
02085500	1970-2006	34	18.1	51.1	101.7	0.341	2.007	-13.732	-14.5641	84.5211
02082950	1970-2006	18	25.1	74.6	152.5	0.903	2.084	-24.925	1.9069	104.268
02088000	1970-2006	14	16.3	63.2	224.1	0.508	1.857	-7.3756	-49.6187	117.5716

**Appendix 1. Characteristics of recession analysis and master recession curve coefficients for selected basins in the Blue Ridge and Piedmont Physiographic Provinces--Continued**

Station number	Period of analysis	Number of recession segments used	Kmin (d/log cycle)	Median recession index (d/log cycle)	Kmax (d/log cycle)	Range of MRC		Coefficients of MRC		
						MinLogQ	MaxLogQ	A	B	C
						1.015	2.349	-5.1195	-35.7799	112.3137
02213050	1961-1994	21	33.6	59.3	219.7	0.046	1.376	6.8716	-96.1828	119.3167
02212600	1970-2006	23	20.6	44.2	99.3	0.111	1.734	-6.8544	-32.2882	76.6128
02221525	1977-2007	19	44.4	73.5	195.2	1.411	2.338	-7.0051	-59.6325	177.7539
02381600	1974-2006	24	60.9	107.8	222.6	0.605	1.415	31.128	-177.2872	188.5282
02380500	1970-2007	35	61.5	113.3	248.2	1.817	2.93	17.061	-205.2879	455.0744
03500240	1970-2006	31	53.7	95.3	277.4	1.414	2.345	67.928	-372.0135	498.8119
03460000	1970-2007	36	36.7	75.5	254.3	1.336	2.211	84.251	-395.4257	462.4303
02137727	1980-2006	19	45.6	104.9	173.8	1.761	2.49	5.528	-129.6897	288.6744
02143040	1970-2006	30	53.6	93.4	211.5	0.961	1.699	11.514	-139.3743	203.5252
02120780	1979-2006	10	41.3	96.9	236	1.573	2.191	-66.158	135.5907	20.5357
02112120	1970-2006	24	53	109.6	298.4	1.625	2.341	62.038	-380.726	551.2617
02077200	1970-2006	24	19	40.1	146.3	0.482	1.464	-1.1069	-45.2941	68.668
02082770	1963-1994	18	32	73.4	157.8	1.056	2.329	-11.317	-34.8078	142.4762
02220900	1977-2007	24	30.8	61.7	159.9	0.988	2.489	-1.9112	-64.447	172.2389

**Appendix 2. Basin parameters and aquifer properties for selected basins in the Blue Ridge and Piedmont Physiographic Provinces**

<b>Station number</b>	<b>Region</b>	<b>Relief (percent)</b>	<b>Drainage Area (mi<sup>2</sup>)</b>	<b>Total stream length (mi)</b>	<b>Drainage density (mi/mi<sup>2</sup>)</b>	<b>Aquifer half-width (ft)</b>	<b>Recession index (d/log cycle)</b>	<b>Groundwater recharge (in/yr)</b>	<b>Average groundwater gradient (ft/mi)</b>	<b>Areal diffusivity (ft<sup>2</sup>/d)</b>	<b>Aral transmissivity (ft<sup>2</sup>/d)</b>	<b>Storage coefficient</b>
03558000	Blue Ridge	9.25	177	277	1.56	1690	118.6	30.95	244	22466	258	0.011
02177000	Blue Ridge	9.76	207	298	1.44	1831	90.1	34.88	258	34718	299	0.009
03550000	Blue Ridge	13.35	104	198	1.91	1383	70.6	25.51	352	25284	121	0.005
03504000	Blue Ridge	15.28	51.9	73	1.40	1889	85.6	45.53	403	38877	257	0.007
03500000	Blue Ridge	12.49	140	184	1.31	2008	93.5	29.19	330	40238	214	0.005
03439000	Blue Ridge	10.69	67.9	104	1.53	1723	101.1	38.81	282	27410	286	0.010
03512000	Blue Ridge	21.22	184	268	1.46	1813	65.2	31.58	560	47037	123	0.003
03455500	Blue Ridge	23.25	27.6	35	1.26	2089	66.1	37.6	614	61611	154	0.003
03456500	Blue Ridge	19.7	51.5	78	1.51	1749	56.1	29.18	520	50859	118	0.002
03463300	Blue Ridge	17.67	43.3	65	1.50	1755	58.5	32.41	466	49150	147	0.003
02111000	Blue Ridge	11.21	28.8	45	1.58	1675	91.5	19.17	296	28613	131	0.005
03479000	Blue Ridge	11.81	92.1	151	1.63	1615	70.3	19.63	312	34633	123	0.004
02111500	Blue Ridge	10.65	89.2	106	1.19	2222	122	17.4	281	37765	166	0.004
03500240	Blue Ridge	14.57	57.1	76	1.34	1972	95.3	26.02	385	38083	161	0.004
03460000	Blue Ridge	17.95	49.2	77	1.57	1682	75.5	24.67	474	34969	106	0.003
02137727	Blue Ridge	11.89	126	190	1.50	1754	104.9	21.93	314	27371	148	0.005
02112120	Blue Ridge	7.68	128	169	1.32	2004	109.6	15.27	203	34193	182	0.005
02138500	Blue Ridge	10.73	66.7	88	1.32	2002	60	23.93	283	62327	204	0.003
02149000	Blue Ridge	8.68	79	120	1.52	1736	122.1	18.64	229	23021	170	0.007
02337500	Piedmont	2.48	37	33	0.88	2995	112.9	15.06	65	74104	830	0.011
02337000	Piedmont	1.71	246	217	0.88	2994	61.8	11.09	45	135289	886	0.007
02333500	Piedmont	4.26	153	243	1.59	1665	103.4	23.54	112	25002	420	0.017
02152100	Piedmont	7.46	60.5	91	1.50	1763	127.1	14.42	197	22824	156	0.007
02143500	Piedmont	2.07	69.2	52	0.75	3522	84.9	10.75	55	136352	835	0.006
02142000	Piedmont	6.08	28.2	40	1.41	1877	92.9	14.58	161	35385	206	0.006
02118500	Piedmont	3.29	155	265	1.71	1546	138.6	13.52	87	16089	290	0.018

**Appendix 2. Basin parameters and aquifer properties for selected basins in the Blue Ridge and Piedmont Physiographic Provinces--Continued**

<b>Station number</b>	<b>Region</b>	<b>Relief (percent)</b>	<b>Drainage Area (mi<sup>2</sup>)</b>	<b>Total stream length (mi)</b>	<b>Drainage density (mi/mi<sup>2</sup>)</b>	<b>Aquifer half-width (ft)</b>	<b>Recession index (d/log cycle)</b>	<b>Groundwater recharge (in/yr)</b>	<b>Average groundwater gradient (ft/mi)</b>	<b>Areal diffusivity (ft<sup>2</sup>/d)</b>	<b>Aral transmissivity (ft<sup>2</sup>/d)</b>	<b>Storage coefficient</b>
02114450	Piedmont	2.75	42.8	66	1.54	1714	99.5	8.73	73	27537	248	0.009
02128000	Piedmont	1.04	106	169	1.59	1658	65.5	7.25	27	39153	528	0.013
02085500	Piedmont	1.31	149	214	1.43	1842	51.1	6.28	35	61940	403	0.007
02082950	Piedmont	1.05	177	243	1.37	1921	74.5	7.69	28	46210	642	0.014
02338660	Piedmont	1.59	127	179	1.41	1874	58.2	10.44	42	56302	562	0.010
02213050	Piedmont	1.32	29	44	1.51	1752	59.3	6.18	35	48321	375	0.008
02212600	Piedmont	1.57	72.2	102	1.41	1872	44.2	4.57	41	74004	249	0.003
02221525	Piedmont	1.53	190	269	1.41	1866	73.5	6.76	40	44207	376	0.009
02381600	Piedmont	5.59	10	7	0.69	3811	107.7	18.17	154	125851	543	0.004
02380500	Piedmont	7.06	236	354	1.50	1762	113.3	23.71	186	25577	270	0.011
02143040	Piedmont	7.95	25.7	47	1.82	1451	93.4	19.11	210	21042	159	0.008
02120780	Piedmont	1.04	118	151	1.28	2057	96.9	8.24	27	40723	744	0.018
02077200	Piedmont	1.77	45.9	51	1.11	2370	40.1	6.88	47	130635	421	0.003
02082770	Piedmont	1.31	166	257	1.55	1705	73.4	9.79	35	36948	582	0.016
02220900	Piedmont	1.58	262	369	1.41	1877	61.7	6.52	42	53255	354	0.007
02088000	Piedmont	0.79	83.5	133	1.59	1657	63.2	6.97	21	40542	668	0.016



### Appendix 3. List of Equations

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#### Appendix 4. Glossary

$\frac{T}{S_y}$  is **areal diffusivity** (ft<sup>2</sup>/d) - The ratio of transmissivity divided by specific yield. Hydraulic diffusivity is proportional to the speed at which a pressure pulse will propagate.

**$T$  is areal transmissivity** (ft<sup>2</sup>/d) - The volume of water flowing through a cross-sectional area of an aquifer that is 1 ft.  $\times$  the aquifer thickness, under a hydraulic gradient of 1 ft./ 1 ft. in a given amount of time (usually a day).

**$S_y$  is specific yield (dimensionless) in regard to an unconfined aquifer** - The ratio of the volume of water which the porous medium after being saturated, will yield by gravity to the volume of the porous medium

**$a$  is aquifer half-width** (mi)

**$K$  is recession index** (day) - The time (number of days) which is required for groundwater discharge to drop through one log cycle after the recession started

**$Q_t$  is discharge at time  $t$**  (ft<sup>3</sup>/s) - The streamflow at some time after the recession started

**$Q_0$  is the initial discharge** – The flow at the start of recession

**$t$  is the time** (day) – The time since the start of recession

**$\log Q$**  (ft<sup>3</sup>/s)- The logarithm of flow

**$k_1$  and  $k_2$  are coefficients** - The absolute value of  $k_1$  is accounted for by recession constant; that is, the reciprocal negative value of the slope.  $k_1$  is the recession index (days per one log cycle)

**$N$  (days)** - The number of days between the storm peak (peak of stream-hydrograph) and start of the baseflow (end of surfaceflow).

**$A$  (mi<sup>2</sup>)** - The drainage basin area in square miles

**$Tc$  is critical time** - The time period (in day) since the recharge event during which the profile of water table distribution is instable.

**$h_1$  is the water level at time  $t_1$  during the period of recession and  $h_2$  is the water level at time  $t_2$  during the same period of recession**

**$R_g$  is effective recharge (or the baseflow)** (ft<sup>3</sup>/day) - Streamflow coming from groundwater seepage into a stream or river

**$ET_g$  is riparian evapotranspiration** (ft<sup>3</sup>/day) - The loss water from the soil through both evaporation and transpiration from plants.

**$I$  is the groundwater slope** (ft/mi) - The average groundwater gradient from divides to streams

**$L$  total stream length** (mi) - The length of discharge areas or total stream length

**$Y_g$  is gravity yield (a dimensionless ratio)**

**$\Delta S_g$  is the increase in groundwater storage in a specific period (expressed in inches of water over the area)**

**$\Delta H_g$  is the corresponding increase in ground water stage (expressed in inches)**

**$D$  is drainage density in** (mi/mi<sup>2</sup>) - Length of all channels above those of a specified stream order per unit of drainage area.

**$S_{median}$  is median basin slope (percent)** - The median of all slope values calculated within a watershed

**$E_{median}$  is the median elevation of hydraulic divides** above the streams within each watershed (ft)