James D. Gosweiler. THRESHOLD DRAINAGE AREAS IN COASTAL PLAIN STREAMS. (Under the direction of Dr. Jonathan D. Phillips) Department of Geography, September, 1994.

Fluvial geomorphologists and watershed hydrologists have long recognized basic differences between small and large drainage basins. However, few have investigated whether there are thresholds that separate small and large drainage basins within a given geomorphic region. While extremes in large and small basins may be a result of differences in fluvial processes, such evidence does not demonstrate a threshold drainage area. This study proposes a model for identifying the threshold between small and large drainage basins based on long-term alluvial sediment storage within the Tar River watershed on the Coastal Plain of North Carolina. Drainage areas associated with the upstream limits of long-term alluvial sediment storage were identified to determine the threshold drainage area. The model was based on Hack's (1957) area-length relationship and tests the effects of local hillslope and upstream drainage area on the location and size of the threshold drainage area.

The model shows that drainage areas associated with the upstream limits of alluvial sediment storage identify a threshold drainage area indicative of the upstream-downstream differences in energy/sediment storage within five Coastal Plain streams. Results suggest that at the threshold there is a characteristic increase of upstream drainage area relative to local hillslope area. The threshold drainage area was < 10 km^2 for all streams with a range of less than two fold between the streams, confirming Hack's (1957) relationship is an important component of the model. The local hillslope/upstream

drainage area ratio was < 0.5 for all streams within the study. This supported the hypothesis that local and upstream hillslopes demonstrate significant control over the location and size of the threshold drainage area.

THRESHOLD DRAINAGE AREAS IN COASTAL PLAIN STREAMS

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CHAPTER I

INTRODUCTION

Drainage basins can be described as open systems that exchange matter and energy with their surroundings (Chorley and Kennedy, 1971). Generally, a basin can be defined for any point along a stream by determining its upstream catchment area. Fluvial geomorphologists and watershed hydrologists have long recognized the basic differences in hydrologic response and geomorphic behavior between small and large drainage basins (Table 1). In general, small basins exhibit rapid response to rainfall and relatively rapid translation of water and sediment downstream. In small basins the channel is coupled with and reacts to the immediately adjacent hillslopes. In contrast, larger basins are characterized by significant lag times in response to runoff events and significant (sometimes extensive) water and sediment storage (Knighton, 1984). In larger basins the channel is influenced by adjacent hillslopes as well as the runoff- and sediment-contributing slopes upstream.

Scientists have established numerous relationships for analyzing the morphometry of drainage basins. Among these, both Horton's (1945) and Strahler's (1952) stream classifications have provided a well-known means of describing the average profile of entire fluvial systems based on stream order (Knighton, 1984). These ideas were advanced by others who described the profiles of rivers based on channel parameters of the drainage basin.

Table 1
Geomorphic Differences Between Small and Large Drainage Basins

Drainage Basin	Geomorphic Characteristics Controlling Sediment Transport Efficiency	Sediment Yield	Hydrologic Response and Geomorphic Behavior	Evidence
Small Basin	Hillslope Morphology and surficial roughness	Sediment Supplied from the Adjacent Hillslopes; Sediment washed into the stream	Rapid Response to Rainfall; Rapid Translation Downstream; Bank Erosion Bed Erosion; Temporary storage	Local spatial variation in hillslope morphology, soil properties, and vegetation cover; course-grain sediments; temporary deposition on alluvial islands, and bars
Large Basin	Hillslope Morphology and efficiency of principal channels	Sediment supplied from adjacent hillslopes and sediment washed into stream; Sediment supplied from contributing hillslopes upstream	Significant Lag Times; Significant Water and Sediment Storage	Floodplain deposits; wetlands; and deltas

Schumm's (1956) constant of channel maintenance, correlated basin area A_d with mainstream length L,

$$C = A_d/L \qquad (1)$$

implying that a minimum area is required to support a unit length of a channel. Hack (1957) suggested that in drainage basins ranging in size up to 10^4 km^2, mainstream length increases directly as the 0.6 power of the drainage area,

$$L = 1.4\ A_d^{0.6} \qquad (2)$$

This indicates that drainage basins elongate with increasing size. These studies represent important principles in network evolution and have laid the foundation for modern drainage basin analysis (Knighton, 1984).

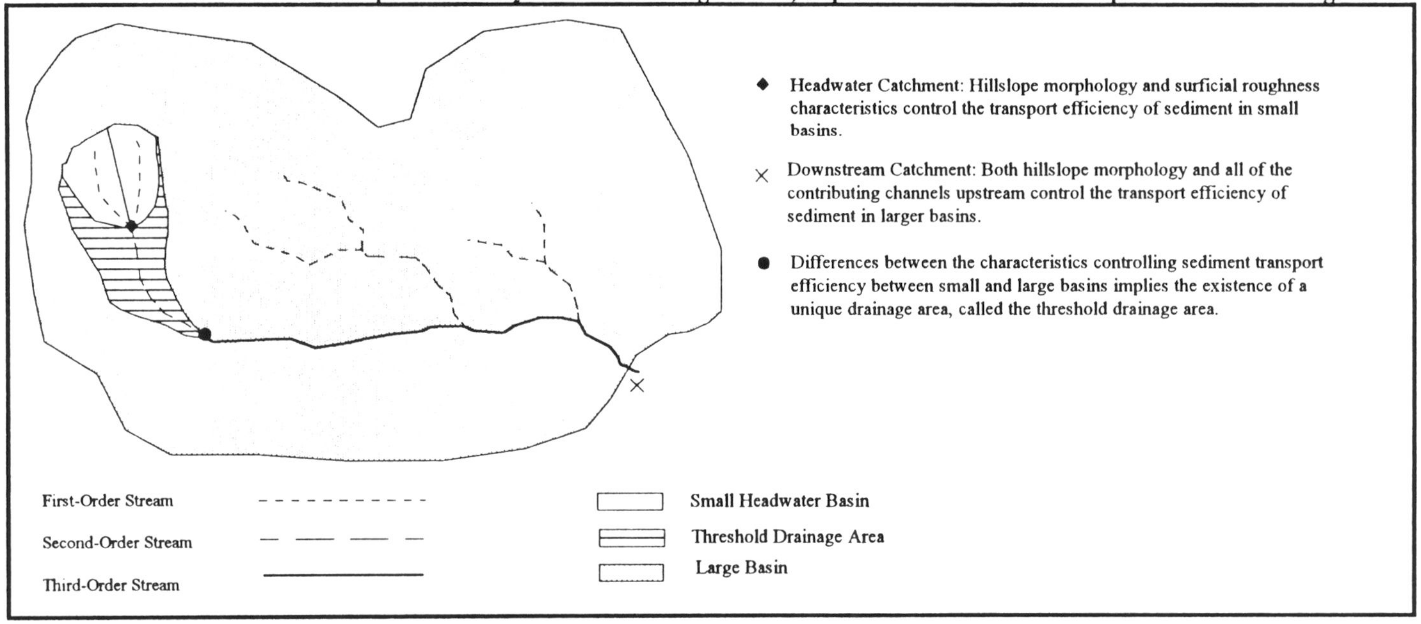
The effect of scale on drainage basin analysis is significant because of the need to relate spatial patterns and form to processes so that valid process/form models can be developed (Penning-Rowsell and Townshend, 1978). Geomorphologists recognize that a wide range of scales may be regarded as appropriate for analysis, with reconciliation of the gap between an understanding of process at the small scale and form development at a larger scale being achieved at some intermediate spatial and temporal scale (Knighton, 1984). Many studies have shown how process-response relationships may vary with spatial scale (Gregory and Gardner, 1975; Penning-Rowsell and Townshend, 1978; Ebisemiju, 1985). However, few have explicitly addressed the question of defining "small", and "large" basins. A recent advance in the attempt to define the spatial extent of basins has been through the concept of threshold drainage areas.

Threshold Drainage Areas

While a theoretical basis for a threshold drainage area has not yet been established some scientists have suggested that such a threshold exists. Graf's (1987) analysis of basins in the Colorado Plateau suggests that a threshold is encountered along the drainage divide-to-downstream gradient and related it to the presence of floodplains and other fluvial landforms. Clarke and Waldo's (1986) study of a single basin in southern Illinois suggests that the threshold drainage area is based on geomorphic characteristics that influence sediment transport efficiency. Hillslope morphology and surficial roughness characteristics control the transport efficiencies of sediment in small basins. In contrast, both hillslope morphology and the principal channels control the transport efficiency of sediment in larger basins. Thus, Clarke and Waldo (1986) suggest that the differences between the characteristics controlling sediment transport efficiency differ between smaller and larger drainage basins. This implies the existence of a unique drainage area called the threshold drainage area (Figure 1). Their methods for identifying the threshold drainage area included following selected channels downstream until encountering a recognizable sediment sink or trap (i.e. upper ends of floodplains, alluvial fans, or deltas) indicative of long-term sediment storage. The drainage area above this point is then calculated, where several determinations define the mean and range of values of the threshold drainage area (Clarke and Waldo, 1986).

Despite the general recognition of differences between "small" and "large" drainage basins, few have investigated the threshold separating them. Moreover, little is known about the size of threshold drainage areas. Conceptually, both Clarke and Waldo (1986) and Graf (1987) implicitly recognize that the threshold drainage area separates different modes of operation within the hillslope subsystems (i.e. transport efficiency and sediment storage) between smaller and larger basins. What is lacking is a formal statement in terms of the physical principles showing that the threshold drainage area does exist.

Figure 1. The Threshold Drainage Area (after Clarke and Waldo, 1986). Differences between characteristics that control sediment transport efficiency in small and large basins, implies the existence of a unique threshold drainage area.



- ◆ Headwater Catchment: Hillslope morphology and surficial roughness characteristics control the transport efficiency of sediment in small basins.
- × Downstream Catchment: Both hillslope morphology and all of the contributing channels upstream control the transport efficiency of sediment in larger basins.
- ● Differences between the characteristics controlling sediment transport efficiency between small and large basins implies the existence of a unique drainage area, called the threshold drainage area.

There is a simple deductive argument that for any fluvial system, with the hypothesized properties, there exists a critical threshold separating small and large drainage basins. The basic assumptions are as follows:

(1) An upstream reach that has no significant floodplain development and only temporary sediment storage (2) A downstream reach where there is significant sediment storage indicated by well-developed floodplains, or other sediment sinks and traps. There must, then, be some upstream limit of the sediment storage separating these distinctive zones. As drainage area increases in the downstream direction, the upstream limit of the sediment sinks is associated with a unique upstream drainage area. Thus, the presence or absence of sediment sinks (or a floodplain) is representative of different energy dissipation regimes, so that the drainage area associated with the upstream limit of sediment storage is a threshold drainage area indicative of different upstream and downstream geomorphic behaviors.

Principles of energy dissipation patterns may provide theoretical support that can facilitate the development of a formal statement that a threshold drainage area exists. The area-length relationship provides insight into the factors that determine the size of the threshold because of the relationships established with respect to channel parameters of the drainage basin. Hack (1957) suggested that length and drainage area are interdependent quantities such that one changes with respect to the other at a rate that appears to be uniform over large areas. Furthermore, the length of a stream at any locality is, on the average, proportional to some power of its drainage area at the locality

(Hack, 1957). This relationship has proven to hold true for a variety of environments (Leopold et al., 1964; Gray, 1961; Mueller, 1973; Shimano, 1975; Newson, 1978). If the area-length relationship can be used as a means for calculating the size of threshold drainage areas then a threshold separating small and large drainage basins can perhaps be determined.

Purposes and Objectives

The purpose of this study is to develop a model for identifying the threshold between "small" and "large" drainage basins based on long-term alluvial sediment storage and to test it on streams within the Tar River basin on the Coastal Plain of North Carolina. The primary goal is to determine if drainage areas associated with the upstream limit of alluvial sediment storage reflect differences in the geomorphic behavior of "small" and "large" drainage basins. The upstream limits of long-term alluvial sediment storage were identified as those areas which met the criteria established for significant storage. Cross-sectional area was surveyed and bankfull discharge and stream power were determined for each location identified as a significant storage site.

The model proposed for this study was based on hypotheses concerning the local hillslope/upstream drainage area ratio. Analyses based on Hack's (1957) area-length relationship will be important considerations for developing the model. The threshold drainage area concept suggests that a unique drainage area is manifested in the differences observed in the characteristics between small and large drainage basins. However, few studies have investigated the threshold that separates small and large drainage basins. This study has implications for understanding sediment storage in fluvial systems.

CHAPTER II

LITERATURE REVIEW

Spatial Scale and Basin Behavior

Drainage basin analysis was largely influenced by the work of Horton (1945). The Hortonian framework provided a methodology for classifying segments of stream channels according to their hierarchial position in a network by means of stream order. The results of the former, combined with the work of Schumm (1956), led to the "laws" of drainage network composition. The "laws" of drainage basin composition show a strong loglinear relationship between basin order and the number of streams, mean stream length, and mean drainage area (Knighton, 1984). Correlation of size and shape has been a common theme in geomorphic analysis (Woldenberg, 1966; Mosley and Parker, 1972; Bull, 1975; Church and Mark, 1980).

From a biological and hydrological viewpoint, the attributes of a stream are dependent on the downstream transfer of water, sediment, nutrients and organic debris (Petts and Foster, 1985). Progressive changes in temperature, stream width and depth, channel pattern, velocity, and sediment load contribute to the differences in energy and matter exchanged within the system (Gordon et al., 1992). Graf (1982) maintains that spatial variations in fluvial processes imposed by external considerations, (i.e. surficial materials, structural considerations such as the hierarchical drainage network and distance to points of disruption) are expressed by the distribution of energy in the system. Thus, it has been recognized that there are fundamental size and scale differences associated with different positions in the drainage network.

Penning-Rowsell and Townshend (1978) studied the effects of scale on the factors affecting stream channel slope and showed that local variations in stream slope are related to stream bed material size while broader variations in reach slope are more closely related to stream discharge. Gregory and Gardner (1973) and Morgan (1973) revealed that the relative importance of climate as a factor controlling drainage density varies with the scale of analysis.

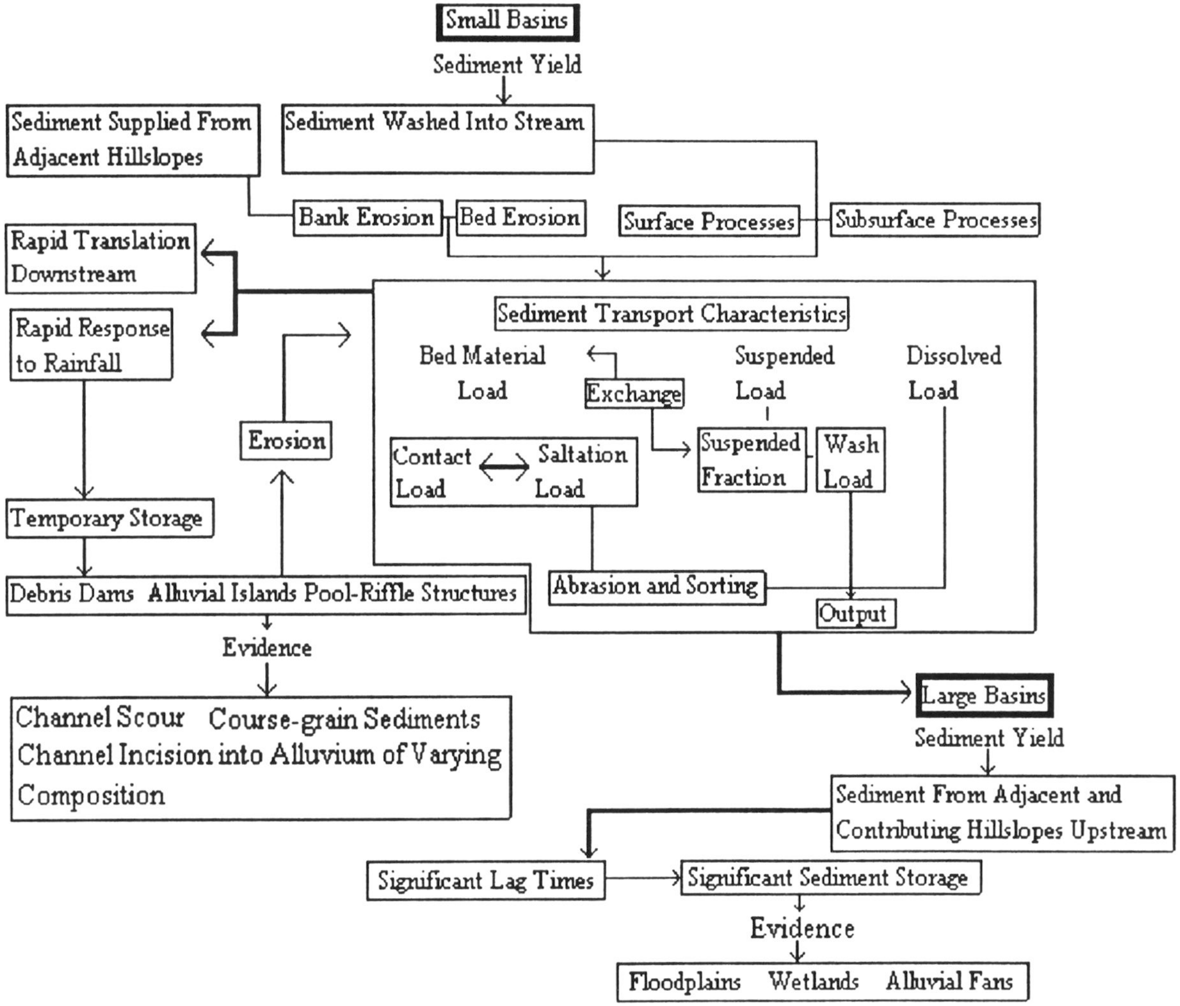
The recognition that process-response relationships vary with scale implies a corresponding acknowledgement that there are fundamental differences between large and small drainage basins. Statistical analyses have shown that most of the variation in the morphometric parameters of drainage basins can be accounted for by variations in basin scale (Doornkamp and King, 1971). Onseti and Miller (1974) investigated the manner in which morphologic variables interacted in a downstream direction within a drainage basin and found that the interrelationship between drainage basin size and stream channel characteristics improves with increasing stream order. The effect of spatial scale changes on the interaction among morphometric properties of drainage basins was also examined by Ebisemiju (1985), through order-by-order principal components analysis. The results show how the strength of the interaction among the variables in each correlation set varied with basin order; the intensity of interaction among the planimetric elements increases while that of slope attributes decreases with an increase in basin size

(Ebisemiju, 1985). Low-order basins are usually characterized by steeper gradients of channels and adjacent hillslopes, with an overall lack of floodplain development. Hillslope morphology and surficial roughness characteristics control the efficiency of low-order basins to deliver eroded sediments to their mouths. Sediment yield from a high-ordered basin depends both on hillslope morphology and the efficiency of the principal channels to transport sediment . Since a higher-order channel integrates the influence of many slopes, it is relatively insensitive to hydrologic or geomorphic changes in any single slope (Petts and Foster, 1985). Thus the fluvial, upstream factors, in addition to local hillslope factors, will be significant in identifying the threshold drainage area between small and large basins.

The spatial and temporal pattern of sediment transfer and storage in drainage basins is a theme central to the study of fluvial geomorphology. Studies of sediment transfer may involve the transport of material in terms of process regimes (Statham, 1977), sediment yield from small watersheds (Clarke and Waldo, 1986) episodic inputs of sediments to headwaters from hillslopes (Macklin and Lewin 1989), and the response of stream channels to short-term flow variability (Rhoads and Miller, 1991). All material entering a stream system must cross the boundary between the slope and fluvial regimes provided by the channel banks and channel head. Figure 2 depicts the movement of sediment in and through the fluvial system.

The most common depositional feature in a fluvial system is the floodplain, formed from a combination of within-channel and overbank deposition. The relative

Figure 2. Sediment movement in and through the fluvial system. Adapted from Knighton (1984), illustrating the differences in sediment transport and storage between small and large drainage basins.



importance of lateral and vertical accretion is not constant and overbank deposition may assume greater significance where flooding is more frequent and fine-grained material more readily available for transport (Knighton, 1984). Floodplains provide storage space for sediment as it moves through a drainage basin, the potential for which increases as they become wider with distance downstream (Knighton, 1984). Alluvial floodplains are typically absent in headwater basins. Therefore, floodplains can be thought of as an indicator, and direct result of, the differences between small and large basins.

Downstream Variations in Channel Slope and Gradient

Whether position is defined as a topologic, geometric or flow-related expression, Knighton (1982) contends the most striking element of fluvial change is in the longitudinal direction. Downstream trends of specific fluvial processes may explain, individually or in combination, the geomorphic and hydrologic differences between small and large basins. The overland transport and deposition of sediment is highly dependent on both the hillslope and channel characteristics which influence the velocity of surface runoff. The slope or gradient of a channel is one of several factors involved in mutual adjustment to imposed loads of water and sediment. (Knighton, 1984).

In general, channel slope adjusts more slowly than other hydraulic geometric variables. Headwater tributaries and the upper reaches of streams are generally steeper than lower reaches. The longitudinal profile shows a characteristic concave shape, with slope decreasing from the upper "eroding" reaches to the lower "depositional" ones

(Gordon et al., 1992). Increased stream gradient usually increases the stream power with concomitant increases in channel width, decreases in channel depth, and increases in channel roughness. Thus, stream gradient has a significant influence on the factors that determine discharge (channel width and depth). Graf (1982) related the control of slope on stream power by discussing how the change in values corresponds to a change in gradient within watersheds. Hupp (1982) investigated the effects of increased stream gradient on fluvial landforms and vegetation patterns. The study concluded that the effects of floods can be expected to be more damaging in areas of steeper gradients rather than gentle ones (Hupp, 1982). It is therefore, inevitable that fundamental differences will exist between the upper reaches and lower portions of a stream. Furthermore, it implies that a threshold area may exist somewhere between a small, headwater basin and a larger, higher-ordered basin.

Stream Runoff and the Nature of Overbank Flow

Differences between low and high-ordered basins are further demonstrated by analyzing stream hydrographs. Headwater catchments produce narrow, peaked hydrographs (defined as flashy) which rise and fall quickly. Conversely, high-order basins typically produce wide, rounded hydrographs. The hydrograph shape is also a function of drainage density, slope and length of the runoff surface, the microtopography, subsurface geology, channel characteristics and storm patterns (Gordon et al., 1992).

The influence of floodplains on channel hydraulics can be determined by considering the width-depth ratio. The width-depth ratio generally increases in the downstream direction in most streams (Gordon et al., 1992). However, it is strongly dependent on the composition of the stream banks. If the channel has low-relief floodplains, the width-depth ratio decreases with depth of flow until it reaches a minimum at bankfull flow. As water spreads over the floodplains, the ratio increases again. Under these conditions there are essentially two separate flow regimes, due to dramatically different depths of flow and roughness in the channel and floodplain components. In a v-shaped valley, typical of a headwater stream, however, there is likely to be a generally monotonic trend in width-depth ratio as water level rises, and a single flow regime.

The factors above provide logical physical explanations for variations between small, headwater basins and large high-ordered basins. However, it has not been established that the transitions between "large", high-ordered basins and "small" low-ordered basins within watersheds takes the form of a threshold drainage area.

Geomorphic Thresholds

Traditionally, geomorphologists have focused their efforts on the development of an understanding of the erosional and depositional evolution of landforms through geologic time (Schumm, 1979). However, the numerous deviations from an orderly progression of the erosion cycle have led many to discount erosion cycles (Schumm, 1979). For example, Schumm (1973) concluded that the alluvial and morphological details of drainage basins are much too complex to be explained by progressive erosion alone. His research into the aspects of fluvial landform development indicated that it is critical to acknowledge geomorphic thresholds and the complex response of drainage systems for analyzing drainage system evolution (Schumm, 1973). Schumm (1973) summarized the concept of a geomorphic threshold as:

> "one that is inherent in the manner of landform change; it is a threshold that is developed within the geomorphic system by changes in the system itself through time. It is the change in the geomorphic system itself that is most important because until the system has evolved to a critical situation, adjustment or failure will not occur. It may not always be clear whether the system is responding to geomorphic thresholds or to an external influence, but when a change of slope is involved, the control is geomorphic, and the changes whereby the threshold is achieved is intrinsic to the system."

The significance of the former concept is that abrupt erosional and depositional changes can be inherent in the normal development of a landscape and that a change in an external variable is not always required for a geomorphic threshold to be exceeded (Schumm, 1979). Thus, Schumm (1979) redefines geomorphic thresholds in broader terms as, "a threshold of landform stability that is exceeded either by intrinsic change of the landform itself or by a progressive change of an external variable."

Many studies have focused on geomorphic thresholds to explain the interrelations between process and form in fluvial systems. Schumm and Khan (1972) designed a study to investigate the influence of slope and sediment loads on channel patterns. It was found that if a straight channel was cut in alluvial material at a very low slope, the channel would remain straight, whereas in steeper slopes the channel meandered. The experiments revealed a discontinuous change in stream patterns with increasing slope (from straight through meandering to braided) with fluctuations primarily occurring at two threshold values of stream power. In the following, Schumm (1979) summarizes the importance of the preceding study and the relevance of geomorphic thresholds.

> "Although the increase of valley slope by deposition to a condition of instability is obviously an intrinsic control, the variation of valley-floor gradient that produce river-pattern changes are an extrinsic control, as would be any change in sediment load or stream power that forces a channel across a pattern threshold...When the valley slope is near a threshold, a major flood will significantly alter the stream pattern. This conclusion has bearing on the work of Wolman and Miller (1960), concerning the geomorphic importance of events of high magnitude. They concluded that, although a major amount of work is done by events of moderate magnitude and relatively frequent occurrence, nevertheless, the large storm or flood may have a major role in landscape modification. However, evidence on the influence of rare and large events on the landscape is equivocal. Major floods have destroyed the floodplain of the Cimarron River (Schumm and Lichty, 1963), but they did not significantly affect the Connecticut River (Wolman and Eiler, 1958). These and other observations indicate that a major event may be of either major or minor importance in landscape modification, and an explanation of the conflicting evidence requires further consideration of the threshold concept...The recognition of geomorphic thresholds within a given region will be a significant contribution to the understanding of the details of regional morphology as well as providing criteria for identification of incipiently unstable land forms."

Contemporary research on the presence and prevalence of geomorphic thresholds in fluvial geomorphology has concentrated on recognizing such a contribution. While a theoretical basis for a threshold concerning sediment storage in large and small basins has not yet been established, some work suggests, indirectly, that there is an excellent theoretical foundation for such a threshold. Bull (1979) analyzed the critical-power threshold, which separates the modes of net deposition and net erosion in fluvial systems and found that it is important in understanding complex interactions between the hillslope and stream subsystems. Field studies have also shown that the upslope extent of a channel network may be well defined by an inverse relationship between drainage area and slope (Montgomery and Dietrich 1988, 1989, 1992). Further investigations of the former have led to conclusions that typically, at the channel head, there is a process change, upslope of which mass wasting and diffusive processes predominate and downslope of which runoff-driven incision occurs (Dietrich and Dunne 1993). Dietrich et al. (1993) investigated the connection between erosion process and channel network extent and proposed two threshold of erosion models to explain channel initiation.

There has been some investigation of and speculation on the size of the threshold drainage area. Graf (1987) identified four size classes of basins with respect to sediment storage and transport patterns in the Colorado plateau. He concluded that source streams have drainage areas of less than 1 km^2, local streams have drainage areas of 1 to 1,000 km^2, regional streams have drainage areas of 1,000 to 10,000 km^2, and inter-regional streams have drainage areas exceeding 10,000 km^2. In the humid eastern United States, Clarke and Waldo (1986) contend that small watersheds vary up to about 0.26 km^2, medium watersheds range from 0.26 km^2 to 2600 km^2, and large watersheds are greater than 260 km^2. That study used three approaches: (1) hydrological evidence developed by Leopold et al. (1964) on the power function relationship between drainage area and discharge and stream order and drainage area, (2) correlations between stream order and the presence of significant sediment sinks, and (3) field identification of significant sediment sinks. The hydrologic method accounted for friction losses to the floodplain during flood stages and not friction losses to the channel bed and banks or sediment entrained in the flow (Clarke and Waldo, 1986). Stream order analysis identified the lower order tributaries as most efficient at sediment production. However, Clarke and Waldo (1986) concluded that agriculture activities had extensively modified the drainage characteristics of Raccoon Creek watershed and the field identification method provided the best estimate of the threshold drainage area. Because agricultural activities in the North Carolina coastal plain modify drainage characteristics within the Tar River watershed, the field identification method will be used.

Fluvial geomorphologists have commonly recognized size and scale differences associated with different positions in the drainage network. Many workers have discussed how process-response relationships may vary with spatial scale (Gregory and Gardner, 1975; Penning-Rowsell and Townshend, 1978; Ebisemiju, 1985) and the general hydrologic and geomorphic differences between basins. Furthermore, many scientists have discussed the downstream variations in fluvial processes and the physical explanations accounting for the differences between small and large basins. It is not clear whether the differences observed between basin sizes can be attributed to geographic variations in factors that control sediment transport efficiency (Clarke and Waldo, 1986) and/or to observed differences in sediment storage behavior associated with basin area and upstream-downstream position (Graf, 1987). The validity of the concept is further restricted due to the lack of a formal statement concerning the physical principles showing that a threshold drainage area does exist. This thesis attempts to show that a threshold drainage area does exist between "small" headwater basins and "large" higher-ordered ones within the Tar River watershed. Determination of the threshold separating large and small basins requires the identification of alluvial sediment sinks. Identification of alluvial sediment storage thresholds, following Clarke and Waldo (1986) involves following a channel downstream until encountering a recognizable sediment sink or trap.

Energy and Sediment Storage in Fluvial Systems

The first and second laws of thermodynamics require that all energy and matter supplied to a given fluvial system must be either stored or dissipated. Accordingly,

$$E = E_c + E_d \tag{3}$$

where E is the total matter and energy input at time t, E_c is the rate of internal energy storage and E_d is the rate of energy dissipation, including energy performing work and entropy production, for the system at time t. Thermodynamic rules also prescribe that

$$E_d \geq 0 \tag{4}$$

Generally, there are three relevant types of energy in the fluvial system- potential, kinetic and thermal (Knighton 1984). Potential and kinetic energy perform the mechanical work in the system. Knighton (1984) recognizes that mechanical work operates in a variety of ways: (1) work done against viscous shear and turbulence (internal friction); (2) work done against friction at the channel boundary; (3) work done in eroding the channel boundary; and (4) work done in transporting the sediment load.

In an alluvial stream system positive or negative energy storage is represented by net deposition or remobilization of alluvium. The E_c term in this context is sediment storage. For a steady-state, where sediment is transported through the system, $E_c = 0$ and $E = E_d$. A reference steady-state condition (E'_d) for a given energy input can be defined as

$$E = E'_c + E'_d = E'_d \qquad (5)$$

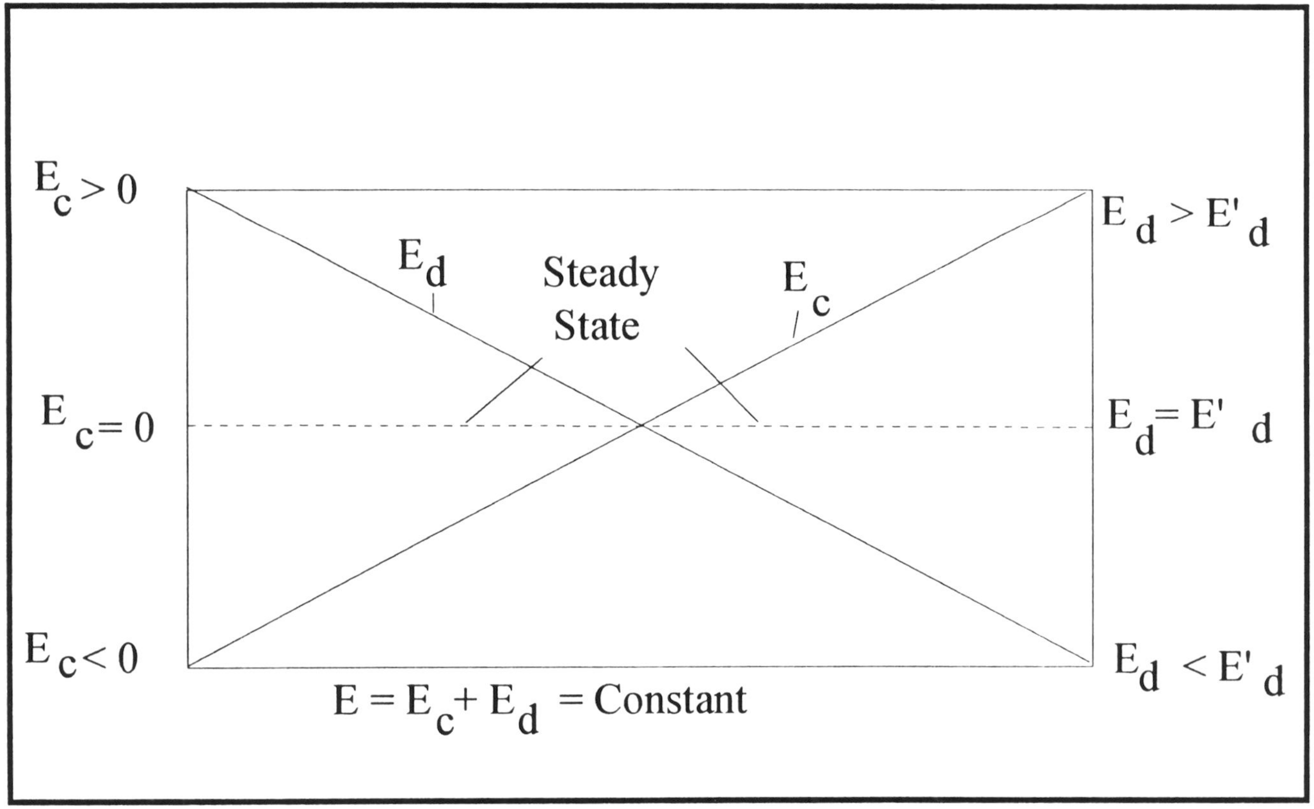
This illustrates how the system must either minimize or maximize energy dissipation relative to E'_d according to the sign of E_c. Thus:

$$\text{If } E_c > 0, \text{ then } E_d < E'_d \text{ and } E_d = \text{min} \qquad (6)$$

$$\text{If } E_c < 0, \text{ then } E_d > E'_d \text{ and } E_d = \text{max} \qquad (7)$$

Figure 3 shows the relationship between energy storage and dissipation rates for constant energy input. Regardless of mutual adjustments and interrelationships between sediment storage/remobilization and energy dissipation mechanisms, the former shows explicitly the role of energy/sediment storage as an indicator of differences in system behavior. Positive or negative values for energy/sediment storage should be reflected in either the minimization or maximization of the rate of energy dissipation. Therefore, by identifying zones of long-term, alluvial storage one can identify zones where the fundamental energy dissipation differs with respect to areas without sediment storage. The above discussion on thermodynamic principles therefore provides a theoretical underpinning that alluvial sediment storage is an indicator of the threshold drainage area.

Figure 3. Relationship showing energy storage (E_c) and dissipation rates (E_d) for constant energy input. E' is the Steady-State dissipation rate.



Empirical Models

A major distinction in describing and explaining the aspects of fluvial systems can be drawn between empirical and theoretical approaches. The former involves the collection and analysis of field data in order to establish relationships between form variables or between a form variable and factors summarizing some aspect of process. The latter is characterized by the formulation and testing of specific hypotheses based on established principles, and involves the construction of models of varying complexity. Most progress is likely to be made with a mixed approach (Knighton, 1984). However, in the absence of definitive theory, drainage network analysis has largely been an empirical and inductive procedure (Knighton, 1984).

Church and Mark (1980) suggest that an important approach in contemporary geomorphology is to search for empirical relations among landform parameters, or between landform parameters and their controlling processes. These relationships have normally been expressed in simple two parameter equations known as power functions (Church and Mark, 1980). One such relationship was established by Hack (1957) between drainage area and mainstream length.

Mainstream Length and Basin Area

Hack's (1957) work suggested that there is a very uniform relationship between stream length and drainage area such that length (measured from any locality on the stream to the source along the longest channel above it), increases directly as the .6 power of the drainage area (Equation 2) for streams in the northeastern United States. However, for two regions in Arizona and South Dakota he found a value of .7 for the exponents and advised that such relations should be considered valid for the region in which they were derived. For basins in the Midwest, in North Carolina, and based on the data from Taylor and Schwarz (1952) in the North and Middle Atlantic States, Gray (1961) found almost identical results,

$$L = 1.40\ A^{0.568} \tag{8}$$

The basins which Hack (1957) and Gray (1961) analyzed ranged in size up to 10^3 - 10^4 km^2.

Mueller (1972) investigated variations of basin shape with size for very large basins ranging from 13,000 to nearly 7.8 million km^2 located around the world in various climatic and geologic environments. Mueller (1972) found that,

$$L = 4.499\ A^{0.466} \tag{9}$$

Mueller (1972) concluded that an exponent below 0.5 indicates that "moderate and large sized drainage basins widen faster than they lengthen" and "the value of 0.466 is so close to 0.5 that the rates at which a basin widens and lengthens are essentially the same throughout much of the basin's development and hence are derived from some equilibrium condition."

A comprehensive model of the area-length relationship over all basin sizes was introduced by Shreve (1974). Shreve's (1974) theory was based on his probabilistic theory of channel networks which postulates that: (1) natural channel networks in the absence of strong geologic controls are very nearly topologically random and (2) interior and exterior link lengths and associated areas in basins with homogeneous climate and geology have separate statistical distributions independent of location. For sub-networks drawn randomly from an infinite topologically random network, Shreve (1974) found that the exponent of the relation declined with increasing network size from 0.6 to 0.5. By simulating network topology and calibrating link lengths and drainage areas empirically, Shreve's (1974) mainstream length/basin area relationship, agreed with the findings of Hack (1957), and Gray (1961).

Since Hack's (1957) original finding, scientists have reported different values of b, generally > 0.5 (Gray, 1961, Leopold et al., 1964; Mosley and Parker, 1973; and Shreve, 1974) and have more recently been concerned with its fractal interpretation (Mandelbrot, 1982; and Robert and Roy, 1990). The area-length relationship therefore suggests that drainage basins must change their overall shape in a downstream direction, becoming longer and narrower as they enlarge regardless of the geologic or structural characteristics of the area.

The Local Hillslope and Upstream Drainage Area Model

The upstream-downstream distribution of sediment storage depends on both the local and upstream hillslopes. Therefore, local hillslope area (A_h) and upstream drainage area (A_d) provide the necessary insight for developing a model to investigate the size and location of the threshold drainage area. Following Schumm's (1956) constant of channel maintenance (C), mean local hillslope area A_h can be represented as the minimum area required to support a unit length of a channel,

$$\bar{A}_h = A_d/\Sigma L = C \tag{10}$$

where A_d is the upstream drainage area and ΣL is total length of the channels. Since upstream drainage area increases relative to local hillslope area downstream and both local hillslope and upstream drainage area determine the location of the upstream limit of sediment storage, the threshold drainage area (A_t) is a function of the ratio of local hillslope to upstream drainage area,

$$(A_t) = f(A_h/A_d) \tag{11}$$

Substituting Hack's (1957) area-length relationship in equation (10),

$$L = Aa_d^b \tag{12}$$

$$\text{then } A_h/A_d = 1/L \tag{13}$$

$$\text{and } A_t = A_h/A_d = 1/L \tag{14}$$

$$\text{thus } A_h/A_d = 1/Aa_d^b \tag{15}$$

Equation (15) suggests there should be some general consistency of (A_t) where equation (11) holds.

Most of the work pertaining to the area-length relationship has been concerned with interpreting the exponent b. With the exception of very large basins (Mueller, 1972), it has been established that typical values for the exponent b fall in the 0.6 to 0.8

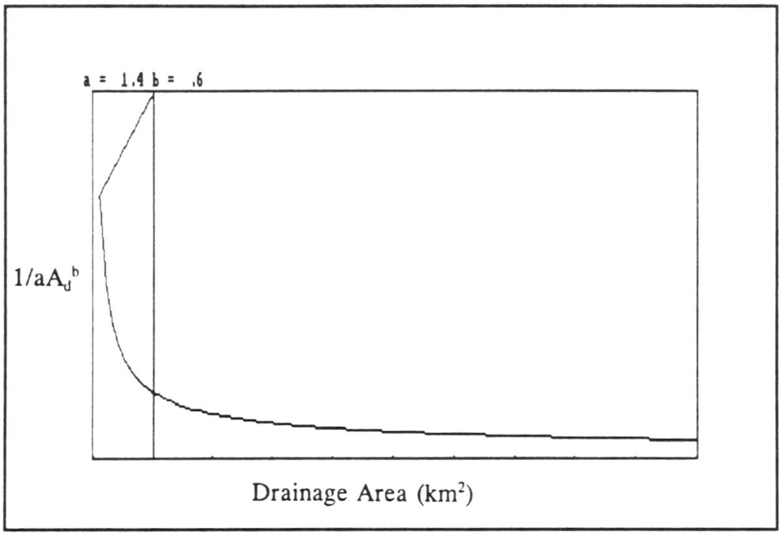
range. This suggests an allometric trend (size- or scale-related change in shape or morphology) toward basin elongation (Hack, 1957; Church and Mark, 1980; Phillips, 1993). However, if a basin's shape is independent of size or length $b = 0.5$. This represents a state of constant proportionality or "self-similarity" and is termed isometry (Church and Mark, 1980). Phillips (1993) showed that a theoretical range for b can be established by considering the role of environmental constraints on drainage density:

> At one extreme, consider the case where drainage density is unconstrained and the network has had sufficient time to attain an equilibrium. In this case, drainage density is constant and $b = 1$. At the opposite extreme, consider a case where drainage density and stream length are constrained by environmental controls (lithology and climate) and L is isometrically related to A. In this case, dimensional analysis gives $b = 0.5$. Any allometry associated with variation of environmental constraints with basin size would be positive, as observed in the trend toward basin elongation, giving $b \geq 0.5$. In this way the theoretical range of b is established: $0.5 < b < 1.0$ (Phillips, 1993 p. 150).

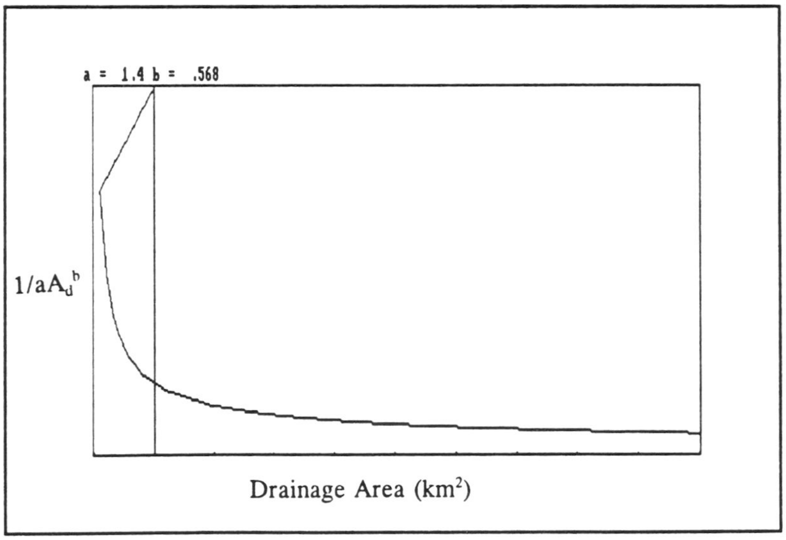
Thus, considering the empirical data introduced by Hack (1957), subsequently investigated by others (Gray, 1961; Leopold et al., 1964; Mosley and Parker, 1973; Shreve, 1974) and using Phillips's (1993) theoretical arguments, the value of the exponent b must fall within a range between 0.5 and 1.0. Hack (1957) also found the coefficient a to average around 1.4 but range between 1 and 2.5. Others have shown various estimates of the coefficient based on samples of fixed sizes (Krumbein and Shreve, 1970) and the reduced major axis technique for various map scales (Robert and Roy, 1990). Hack's (1957) area-length relationship provides insight into the size of the threshold drainage area. Using the empirical and theoretical range of values for the exponent b with any a values, plots of $1/Aa_d^b$ (Figure 4) suggested that the threshold

Figure 4.
Computer plots showing the empirical and theoretical range of coefficients and exponents for the area-length relationship.

Drainage Area is in 10 km^2 intervals and ranges from 0 to 100 km^2 (X Axis). Vertical lines represent 10 Km^2. Plots of $1/aA_d^b$ (Y axis) level off after the first 10 km^2 suggesting that the threshold drainage area $A_t < 10$ km^2 for a wide range of environments for which $L = aA_d^b$ holds.

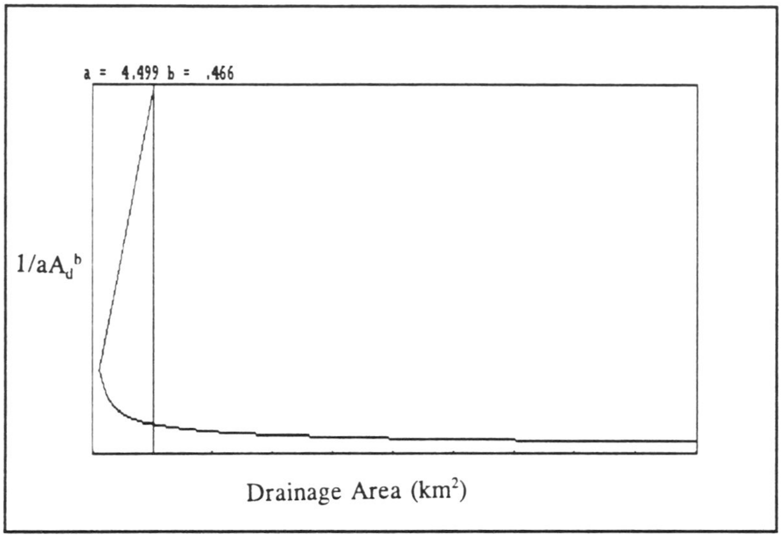


4a. Hack's (1957) coefficient and exponent for the area-length relationship.

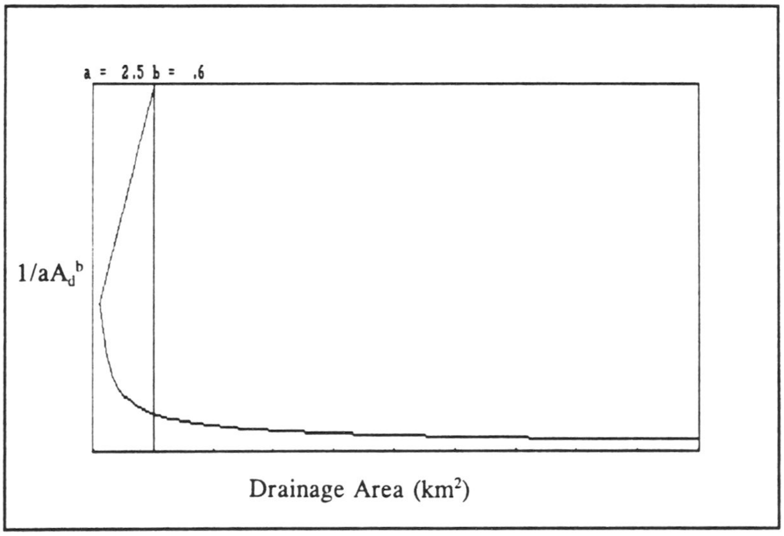


4b. Gray's (1961) coefficient and exponent for the area-length relationship.

Figure 4
Continued

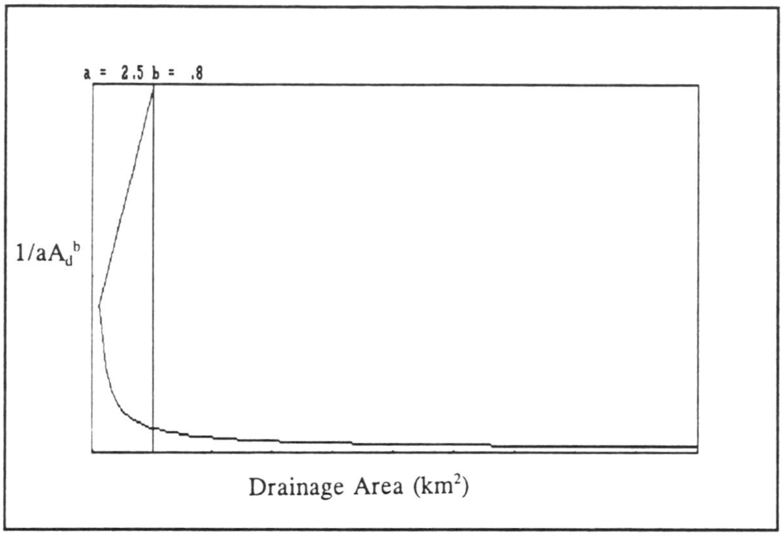


4c. Mueller's (1972) coefficient and exponent for the area-length relationship.

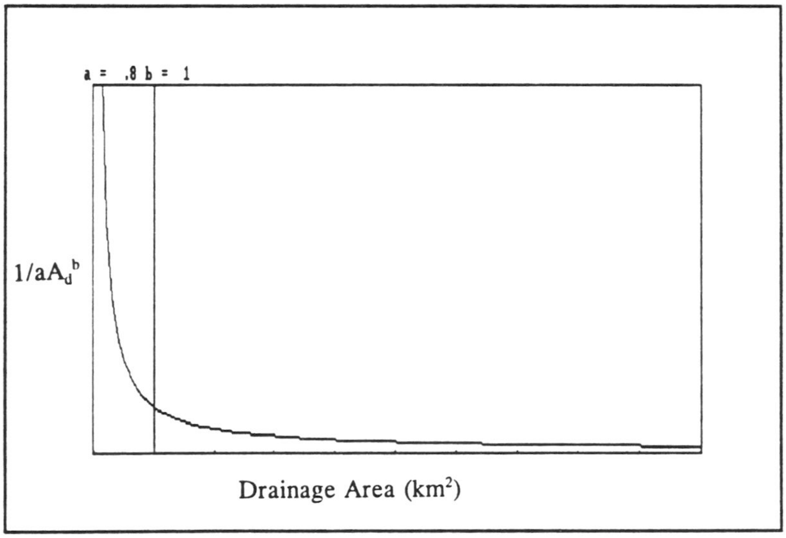


4d. Theoretical values for the area-length relationship, using Hack's (1957) upper and lower limits for the coefficient and exponent.

Figure 4
Continued

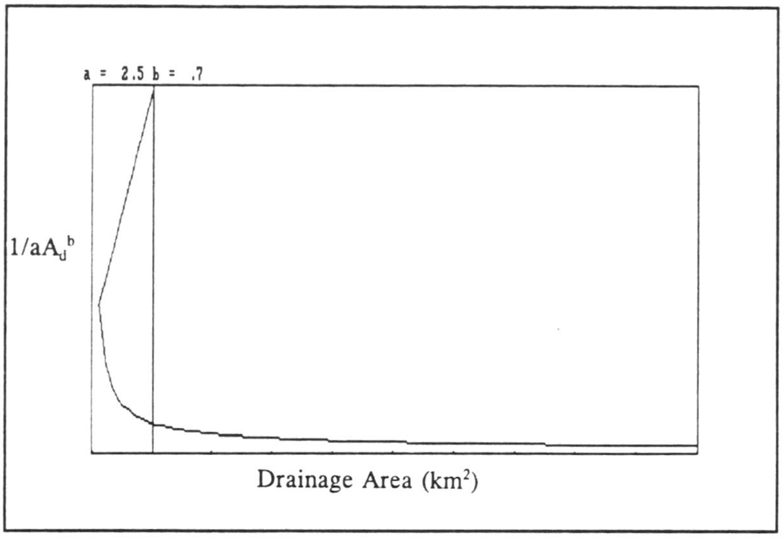


4e. Theoretical values for coefficients and exponents following Phillips (1993)

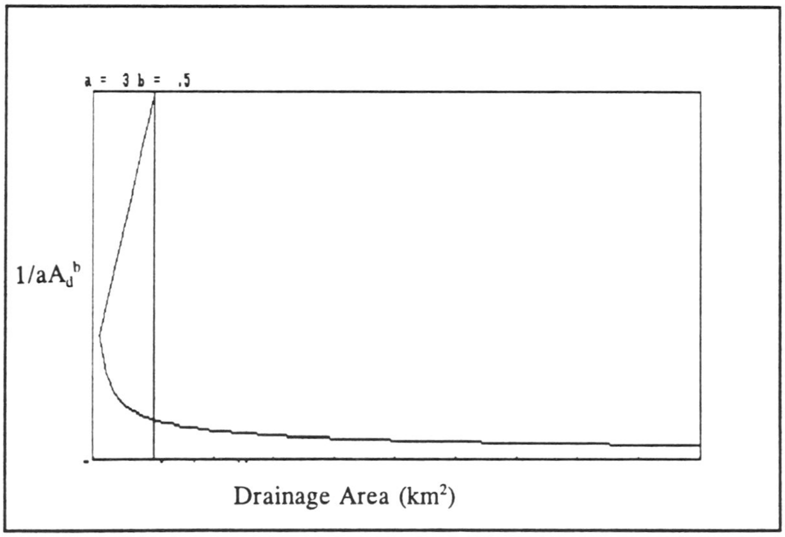


4e. Theoretical values for coefficients and exponents following Phillips (1993)

Figure 4.
Continued



4e. Theoretical values for coefficients and exponents following Phillips (1993)



4e.Theoretical values for coefficients and exponents following Phillips (1993)

4e. Theoretical values for coefficients and exponents of the area-length relationship. Following Phillips (1993), a theoretical range of values for the exponent b can be established ranging from 0.5 to 1.0.

drainage area, $A_t < 10$ km^2 anywhere. Furthermore, because

$$A_d/\Sigma L \approx \overline{A_h} \tag{16}$$

where $\overline{A_h}$ is mean local hillslope area, ΣL is total length of the channels and A_d is drainage area, it followed that,

$$A_h/A_d \approx \Sigma L/A_t^2 \tag{17}$$

Solving equation (17) therefore yields a value indicative of the proportional change in local hillslope to upstream drainage area for the threshold drainage area. Because upstream drainage area increases relative to local hillslope area in the downstream direction it was expected that at the threshold $A_h/A_d < 1$. Further, intuition suggested that for basin influences to become dominant over local hillslope influences, upstream drainage area should be at least twice that of the local hillslope area and therefore, $A_h/A_d < .5$. Therefore, if consistencies between the size of the drainage area associated with the upstream limit of alluvial sediment storage and equation (17) are evident then the size of the threshold drainage area is controlled by the local hillslope and upstream drainage area.

The objective of this study was to test these implications in the field, using the upstream limit of alluvial sediment storage as an indicator of the threshold drainage area separating "small" headwater basins from "larger" higher-ordered ones.

CHAPTER III

STUDY AREA AND METHODS

Geomorphic Setting

The North Carolina Coastal Plain is generally a flat to gently-rolling surface which ranges in elevation from sea level to about 100 m or less. The Coastal Plain is composed of a sequence of sea-level terraces, ranging in age from Pleistocene/Holocene along the coast to Pliocene on the inland boundary (Daniels et al., 1978). Slope gradients of zero to two percent are common and local relief is typically 30 m or less from valley floors to upland interfluves (Daniels et al., 1984). The climate is humid, subtropical and the soils are characterized by strong leaching, low Ph, low natural fertility and formed in unconsolidated alluvial and marine sediments. The water table fluctuates seasonally in response to variations in precipitation and the degree of evapotranspiration (SCS, 1974).

The sediments of the Coastal Plain were deposited during transgressive-regressive cycles caused by eustatic sea level fluctuations. Interglacial periods were characterized by high sea-levels and the subsequent deposition of marine and strandline sediments. During glaciation a falling sea-level caused the regression of strandlines, stream entrenchment, and the subsequent dissection and erosion of Coastal Plain deposits (Soller and Mills, 1986).

The area is typically divided into three regions: the Upper, Middle, and Lower Coastal Plains. The Upper Coastal Plain is underlain by Cretaceous and Tertiary sediments and is separated from the Middle Coastal Plain by the Coats or Orangeburg Scarp. The Middle Coastal Plain is underlain by Pliocene marine sediments with local overlays of Quaternary eolian, lacustrine, colluvial and alluvial deposits. This region is separated from the Lower Coastal Plain by the Surry Scarp. The Lower Coastal Plain is underlain by Pleistocene and Holocene marine, estuarine, fluvial and local eolian deposits (Colquhoun et al., 1991).

Tar River Basin

The study area is part of the Tar River drainage basin on the Coastal Plain of eastern North Carolina (Figure 5). The Tar River rises in the Piedmont, flows across the Atlantic Coastal Plain, and discharges into the Pamlico estuary. The Coastal Plain portion of the basin is characterized by low relief. Characteristic surface features of the nearly uniform plain are gently sloping interstream areas with scattered upland depressions and slightly entrenched streams. Furthermore, streams within the Tar River watershed are characterized by a combination of parallel and dendritic drainage patterns.

Five streams tributary to the Tar River were chosen for this study: (1) Kitten Creek; (2) Otter Creek; (3) Tyson Creek; (4) Harris Mill Run; and (5) Hardee Creek (Figure 6). The streams were selected because: (1) proximity and accessibility; and (2)

Figure 5. Study area located on the Coastal Plain of North Carolina.

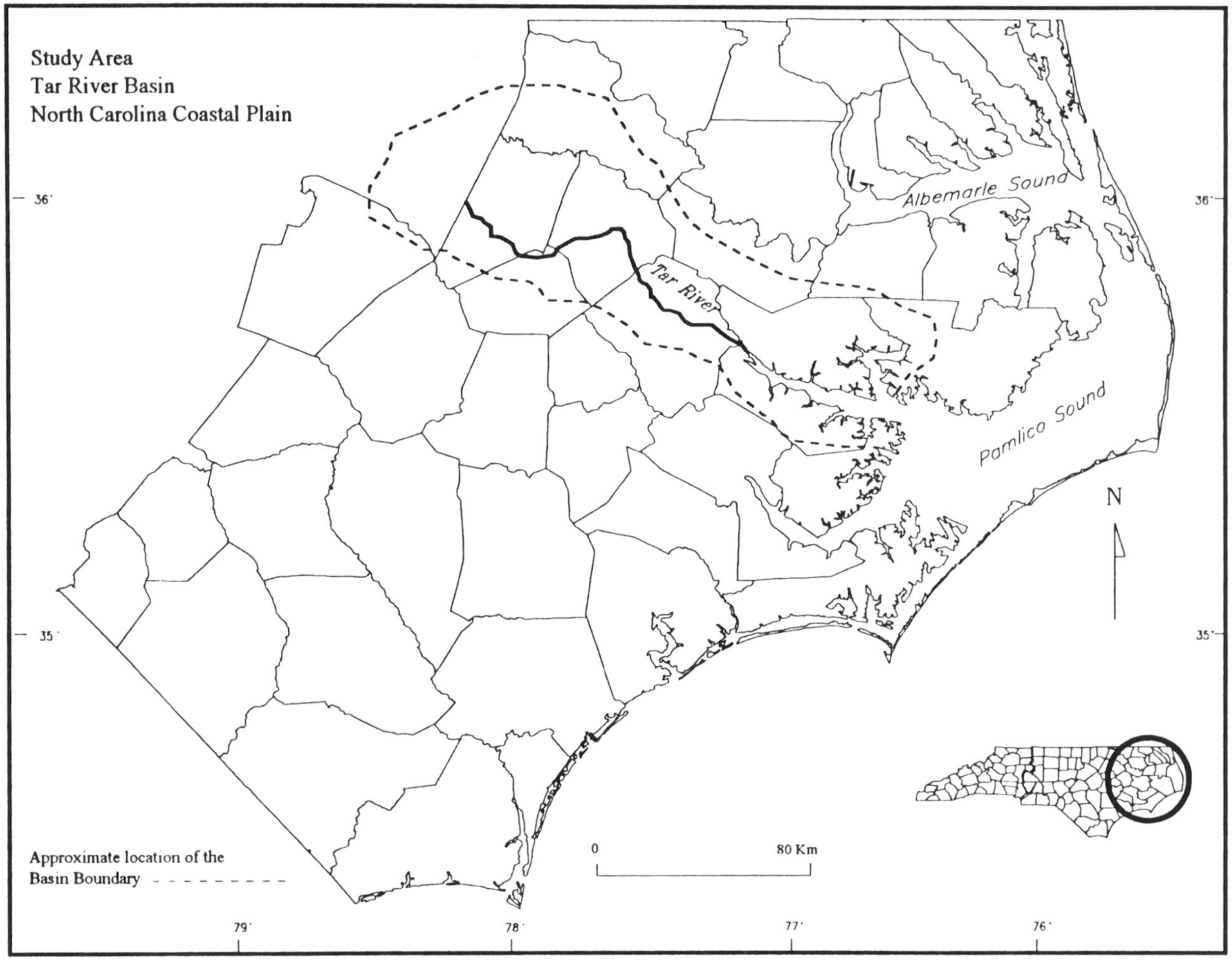
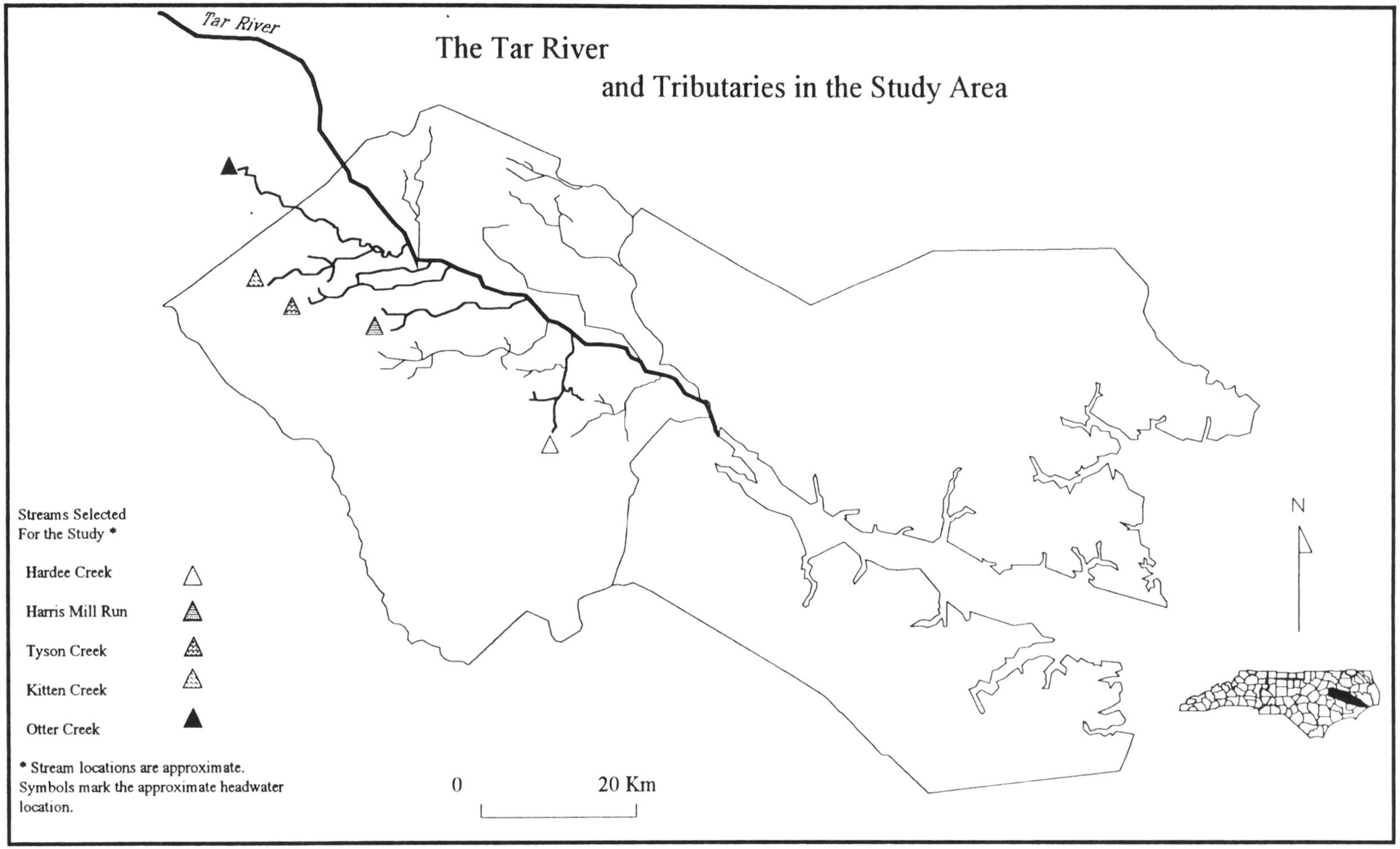


Figure 6. Streams within the Tar River basin selected for the study.



* Stream locations are approximate.
Symbols mark the approximate headwater location.

they have not been channelized or dredged. Aerial photographs assisted in the selection of the streams included in this study. While it is recognized that agricultural activities can modify the drainage characteristics of watersheds by increasing sediment loads (Mason et al., 1990) channelization and dredging can lead to greater flow velocities and higher erosive forces. Furthermore, the clearing of riparian vegetation and debris dams during channelization can release trapped sediments and change the bed topography. Because this study was concerned with determining the upstream limit of long-term alluvial sediment storage, it was felt that streams which have been extensively channelized or dredged would produce biased results. Therefore, it was important to select streams which have not been channelized or dredged.

Methodology Overview

The approach used for testing a model for identifying threshold drainage areas in Coastal Plain Streams may be summarized as a sequence of steps. The first step involved establishing the criteria that was used to identify the upstream limit of alluvial sediment storage.

Second, the location of the tributaries and their headwaters were located on USGS 1:24,000 topographic maps. Aerial photographs and soil surveys assisted in identifying the headwater location for each stream.

The third step involved data collection. Beginning at the headwaters and moving downstream along each tributary the upstream limit of alluvial sediment storage was identified. For each location the latitude/longitude coordinates were recorded using a GPS receiver. Cross-section area, hydraulic radius, local channel slope and bankfull Manning's *n* values were recorded. The location of the long-term alluvial sediment storage sites were marked on the USGS 1:24,000 topographic maps. The upstream drainage area associated with each location was then determined and measured with a planimeter. Mainstream length and total stream length were also measured.

The final three steps for this study involved data analysis and interpretation. First, the size of the drainage area associated with the upstream limit of alluvial sediment storage for all five streams was examined. Next, data for the cross-section surveys were analyzed. Finally, the results were interpreted.

Alluvial Sediment Storage Criteria

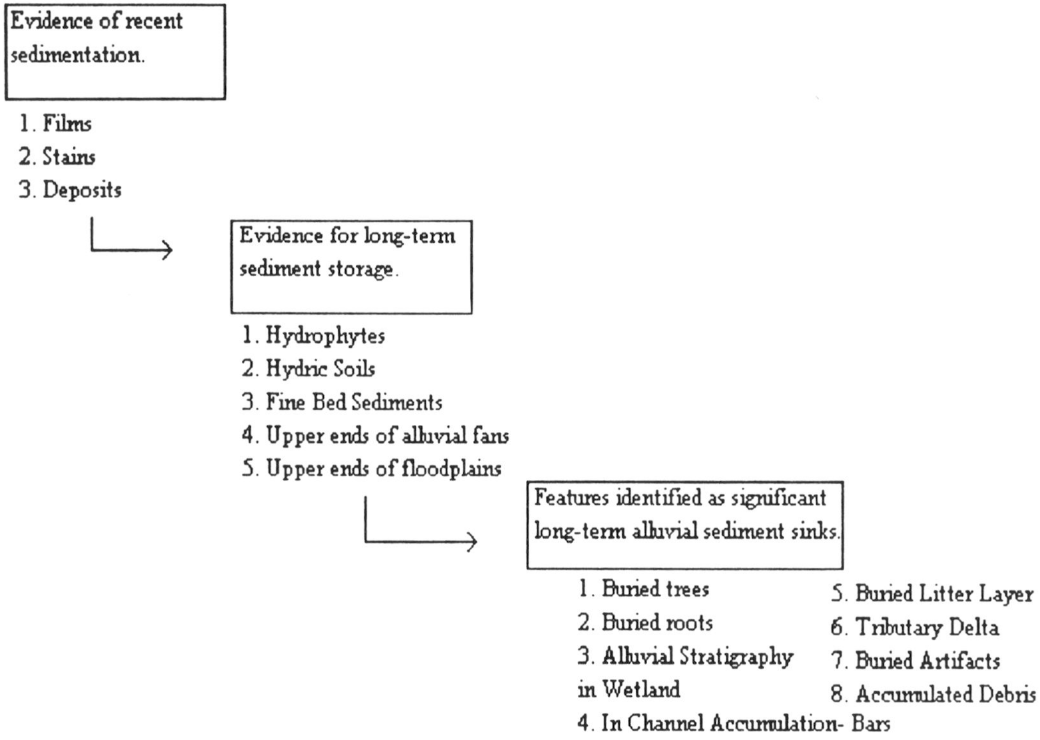
The threshold drainage area separating small, headwater basins from larger higher-ordered ones was determined by identifying sediment sinks on five streams within the Tar River watershed. Clarke and Waldo (1986) used the upper ends of floodplains, alluvial fans, or deltas; ponds, lakes or wetlands; kettles or dolines; and some man made structures to identify locations of significant alluvial sediment storage. However, a critical issue was making the distinction between significant sediment sinks or traps and transitory storage sites. Variability in sediment yield can result from intermittent bank or hillslope collapse, channel incision into alluvium of varying composition, change in land use, and variable patterns of streamflow and dredging (Gordon et al., 1992). Many bars, pool-riffle str uctures and debris dams in active channels can represent material in transient storage. It was, therefore, important to establish a set of criteria prior to data collection in the field. Evidence for a threshold drainage area is listed in Table 2. For this study an arbitrary rule was assigned such that: If a feature was submerged during bankfull flow and was unvegetated it was considered to be in-transit storage, whereas if the feature was not submerged during bankfull discharge or was vegetated it was regarded as an indicator of significant sediment storage. This rule is consistent with the findings in forested drainage areas suggesting that log steps or debris dams are rarely submerged and are important for trapping sediments (Megahan, 1982).

Table 2
Evidence for Threshold Drainage Areas

Location	Qualitative Evidence	Evidence of Significant Sediment Storage
Threshold drainage area is associated with the upstream limit of alluvial sediment storage located between the headwaters and the downstream reaches of streams within the basin.	Upstream limit of long-term alluvial sediment sinks, indicative of the differences in storage characteristics between small and large drainage basins.	Features are not submerged during bankfull flows, or are vegetated

Figure 7 illustrates the criteria that was used to identify significant long-term alluvial sediment storage sites. The distinction between sediment sinks or traps and transitory storage sites was problematic. Features recognized as long-term alluvial storage sites

Figure 7. Criteria for identifying the upstream limits of long-term alluvial sediment storage.



were identified by the following procedure. First, each channel was followed downstream until observed evidence of recent sedimentation such as deposits, films, or stains was encountered. Next, evidence for alluvial sediment storage was observed by locating such features as hydrophytes, hydric soils, fine bed sediments, the upper ends of alluvial fans and floodplains consistent with long-term storage. Finally, features serving as long-term alluvial sediment sinks and traps such as buried trees, buried roots, alluvial stratigraphy in wetlands, accumulated debris trapping sediments and channel fills were identified.

Upstream Limit of Alluvial Sediment Storage

The next step involved locating the upstream limit of alluvial sediment storage. First, the tributaries and their headwaters were located on USGS 1:24,000 topographic maps. Subsequently, for each tributary, locations of long-term alluvial sediment storage (i.e. wetlands, and/or floodplains) were estimated using aerial photographs and soil surveys. These reference locations were verified in the field prior to data collection. This provided a downstream reference location of long-term alluvial sediment storage. Thus, the upstream limit of alluvial sediment storage could therefore be identified at a location somewhere between the headwaters and the reference location.

Data Collection

Beginning at the headwaters and moving downstream, storage sites were identified by locating areas which demonstrated evidence of a threshold drainage area (Table 2) and corresponded with the criteria for long-term alluvial sediment sinks (Figure 7). Once the upstream limit of sediment storage was identified, the latitude/longitude coordinates were recorded using a GPS receiver. The GPS receiver displayed position coordinates to 0.01 minutes. Evidence of the upstream limit of alluvial sediment storage was then recorded.

Next, cross-section area was surveyed. The hydraulic radius was recorded as mean bankfull depth. The channel slope was measured using a clinometer and survey staff. Bankfull Manning's *n* values were estimated following Arcement and Schneider's (1984) procedure. Discharge, defined by the continuity equation is the product of cross-sectional area ($A_c = w*d$) and mean velocity (v).

$$Q = A_c*v = w*d*v \qquad (18)$$

Velocity is strongly related to flow resistance, one of the most important elements in the interaction between the fluid flow and channel boundary and was determined by the Manning (1889) equation,

$$v = \frac{K R^{2/3} s^{1/2}}{n} \qquad (19)$$

where K = 1 (SI units), R is hydraulic radius, s is the slope gradient and n is the Manning's resistance coefficient. Discharge was estimated to determine its relative control over the threshold drainage area.

A useful index for describing the erosive capacity of streams, and therefore directly relevant to sediment transport and storage in the fluvial system is stream power. Stream power refers to the rate of potential energy expenditure as water travels downslope in a channel (Rhoads, 1987). Because stream power quantifies energy expenditure in the fluvial system it was important to evaluate its relationship with the threshold drainage area. Stream power may be defined in several ways. Because the study was concerned with the upstream limit of alluvial sediment storage, it was important to determine the power per unit length at the threshold. Therefore cross-sectional stream power was used. Cross-sectional stream power is defined as:

$$\Omega = \rho g Q S \tag{20}$$

where ρ is the density of water (1000 kg/m^3), g is the acceleration of gravity (9.81 m/s^2), Q is bankfull discharge and S is the slope gradient of the channel (Rhoads, 1987).

Latitude and longitude coordinates marking upstream locations of long-term alluvial sediment storage were marked on the USGS 1:24,000 topographic maps. The drainage area associated with the upstream limit of sediment storage was then determined. A polar planimeter was used to measure the threshold drainage area.

Mainstream length, measured as the distance from the location of significant alluvial sediment storage to the drainage divide at the head of the longest stream above it, and total stream length measured as the sum of the lengths of the channels above that location were also recorded. Finally, values for mean local hillslope and upstream drainage area were determined by using equation (17).

Data Analysis

The first step in the data analysis process was to determine if the upstream limit of alluvial sediment storage identified a threshold drainage area on streams in the Tar River basin. Theoretical and empirical values of the coefficients and exponents for the relationship between mainstream length and drainage area were substituted in equation (15). Plots of equation (15) suggested that $A_t < 10$ km^2 anywhere (Figure 4). Results for the size of the threshold drainage area were then compared to plots of $1/Aa_d^b$ to evaluate the hypothesis that drainage areas associated with the upstream limit of alluvial sediment storage identifies a threshold drainage area which is < 10 km^2 for a wide range of environments.

The second step was to assess the relationship between local hillslope and upstream drainage area and the threshold drainage area. Equation (17) was used to determine the value of local hillslope and upstream drainage area at the threshold drainage area. Since upstream drainage area increases relative to local hillslope area in the downstream direction, $A_h/A_d < 1$. Furthermore, intuition suggested that for basin influences to become dominant over local hillslope influences upstream drainage are should have been at least twice that of the local area and therefore, $A_h/A_d < 0.5$. Data were then analyzed by comparing the local hillslope/upstream drainage area ratio values to the hypothesized values for A_h/A_d to determine the relative importance of local hillslope and upstream drainage area as a control over the threshold drainage area.

The dominant controls of channel form adjustment are discharge and sediment load, which integrate the effects of climate, vegetation, soils, geology and basin physiography (Knighton, 1984). Furthermore, discharge has become a primary independent variable in geomorphological approaches to the description and analysis of river channel form (Knighton, 1984). Furthermore, Clark and Waldo (1986) suggested that characteristics descriptive of the threshold drainage area may be correlated to stream order. Because A_d could be used as a surrogate for factors such as length, discharge, and stream order, the latter were examined to see whether they have any systematic relationship with A_t.

CHAPTER IV

RESULTS

Results of the cross-section surveys as well as the location and description of the upstream limit of alluvial sediment storage for each stream are summarized in Table 3. The drainage areas associated with the alluvial sediment storage locations for each tributary are shown in Figures 8-12. The size of these drainage areas ranged from 2.08 km^2 to 3.47 km^2 with a mean value of 2.66 km^2. Results are consistent with the plots of $1/Aa_d^b$ which suggest that $A_t < 10$ km^2 anywhere.

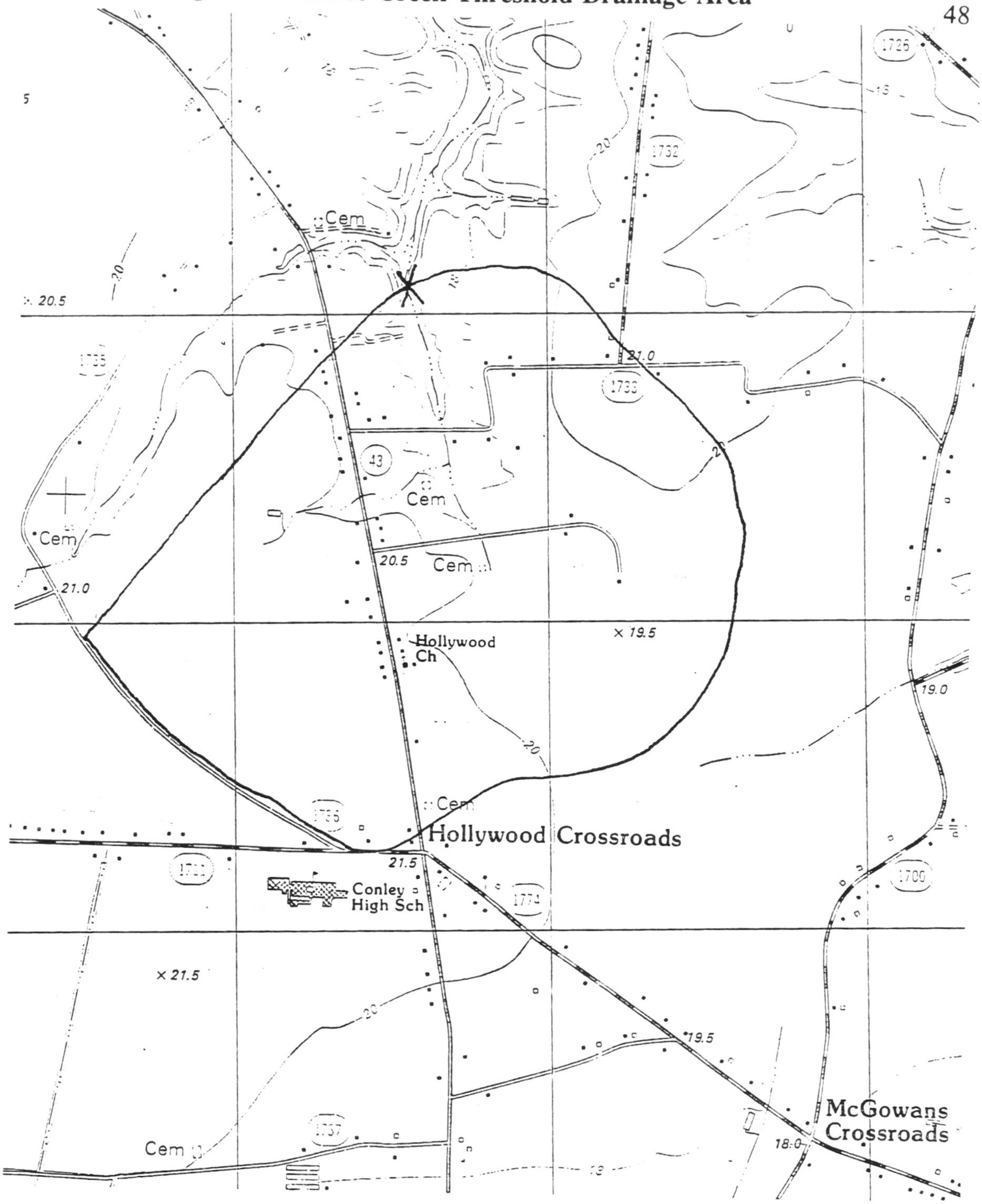
Indicators of Alluvial Sediment Storage

The threshold drainage area was identified by using indicators of long-term alluvial sediment storage to locate the zones where the fundamental energy dissipation differs with respect to areas without storage. Characteristics descriptive of the threshold drainage area for Coastal Plain streams included in this study are summarized in Table 4. The most common indicators of long-term alluvial sediment storage were log jams, debris dams, and floodplains. Several of the locations were also identified as alluvial sediment sinks because of a partially submerged feature indicative of long-term storage. Results show consistency among the indicators for all streams.

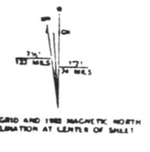
Table 3
Summary of Results:
Data Collected for the Upstream Limit of Alluvial Sediment Storage

Streams	**Hardee Creek**	**Tyson Creek**	**Otter Creek**	**Kitten Creek**	**Harris Mill Run**
Location	35° 32.50' N 77° 19.10' W	35° 39.90' N 77° 34.36' W	35° 43.23' N 77° 42.36' W	35° 41.02' N 77° 37.34' W	35° 37.27' N 77° 26.53' W
Hydraulic Radius (R) m	0.98	0.39	0.55	0.25	0.48
Bankfull Width (w) m	2.8	2.2	2.17	3.0	3.1
Local Channel Gradient (S)	2° 3% Slope	6.5° 11% Slope	4° 7% Slope	3° 5% Slope	1° 1% Slope
Manning's *n* Roughness Coefficient	.07	.06	.08	.05	.12
Estimated Bankfull Discharge **$Q = w \cdot d \cdot v$** **$v = R^{2/3} s^{1/2} 1/n$** **($m^3 s^{-1}$)**	6.46	2.57	3.09	1.28	0.75
Stream Order	Second Order	Second Order	Third Order	Third Order	Third Order
Cross-Sectional Stream Power **$\Omega = \rho GQS$** **(Watts/m)**	1899	2770	2119	627	73
Mainstream Length (km)	0.85	0.8	1.3	1.25	1.6
Total Length of the Channels (ΣL) (km)	1.5	1.65	2.15	2.2	3.05
A_h/A_d = ($\Sigma L/A_t^2$)	0.18	0.38	0.39	0.34	0.25
Threshold Drainage Area (km^2)	2.87	2.08	2.36	2.56	3.47

Figure 8. Hardee Creek Threshold Drainage Area



Mapped, edited, and published by the Geological Survey
Control by USGS, NOS/NOAA, and North Carolina Geodetic Survey
Topography by photogrammetric methods from aerial photographs taken 1977. Field checked 1978. Map edited 1982
Projection and 10,000-foot grid ticks: North Carolina coordinate system (Lambert conformal conic)
1000-meter Universal Transverse Mercator grid, zone 18
1927 North American Datum
To place on the predicted North American Datum 1983 move the projection lines 12 meters south and 27 meters west as shown by dashed corner ticks
Red tint indicates areas in which only landmark buildings are shown



UTM GRID AND 1982 MAGNETIC NORTH DECLINATION AT CENTER OF SHEET

CONTOUR INTERVAL 2 METERS
SUPPLEMENTARY CONTOUR INTERVAL 1 METER
NATIONAL GEODETIC VERTICAL DATUM OF 1929
CONTROL ELEVATIONS SHOWN TO THE NEAREST 0.1 METER
OTHER ELEVATIONS SHOWN TO THE NEAREST 0.5 METER
THIS MAP COMPLIES WITH NATIONAL MAP ACCURACY STANDARDS
FOR SALE BY U. S. GEOLOGICAL SURVEY, RESTON, VIRGINIA 22092
A FOLDER DESCRIBING TOPOGRAPHIC MAPS AND SYMBOLS IS AVAILABLE ON REQUEST

Figure 9. Tyson Creek Threshold Drainage Area

Figure 10. Otter Creek Threshold Drainage Area

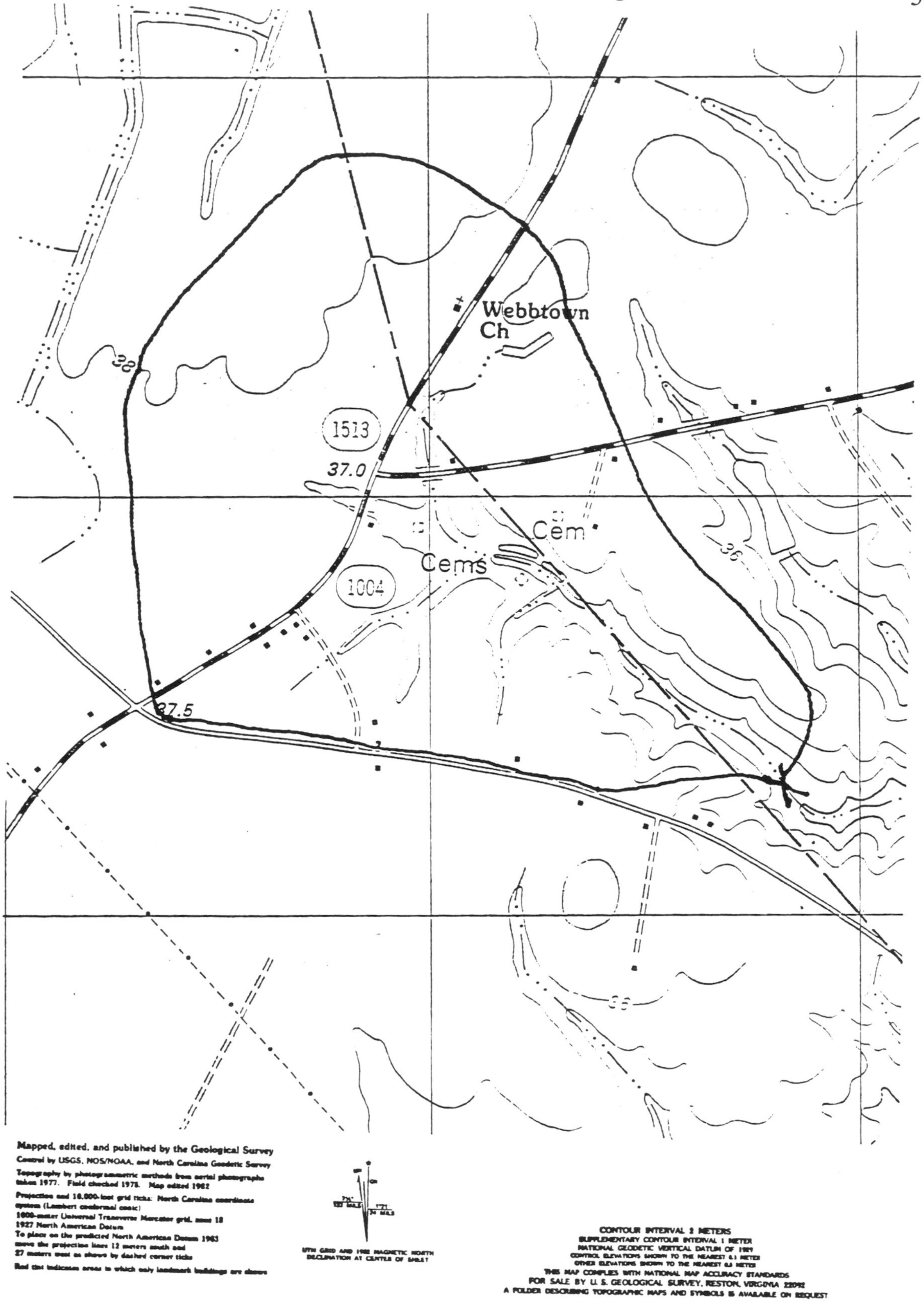
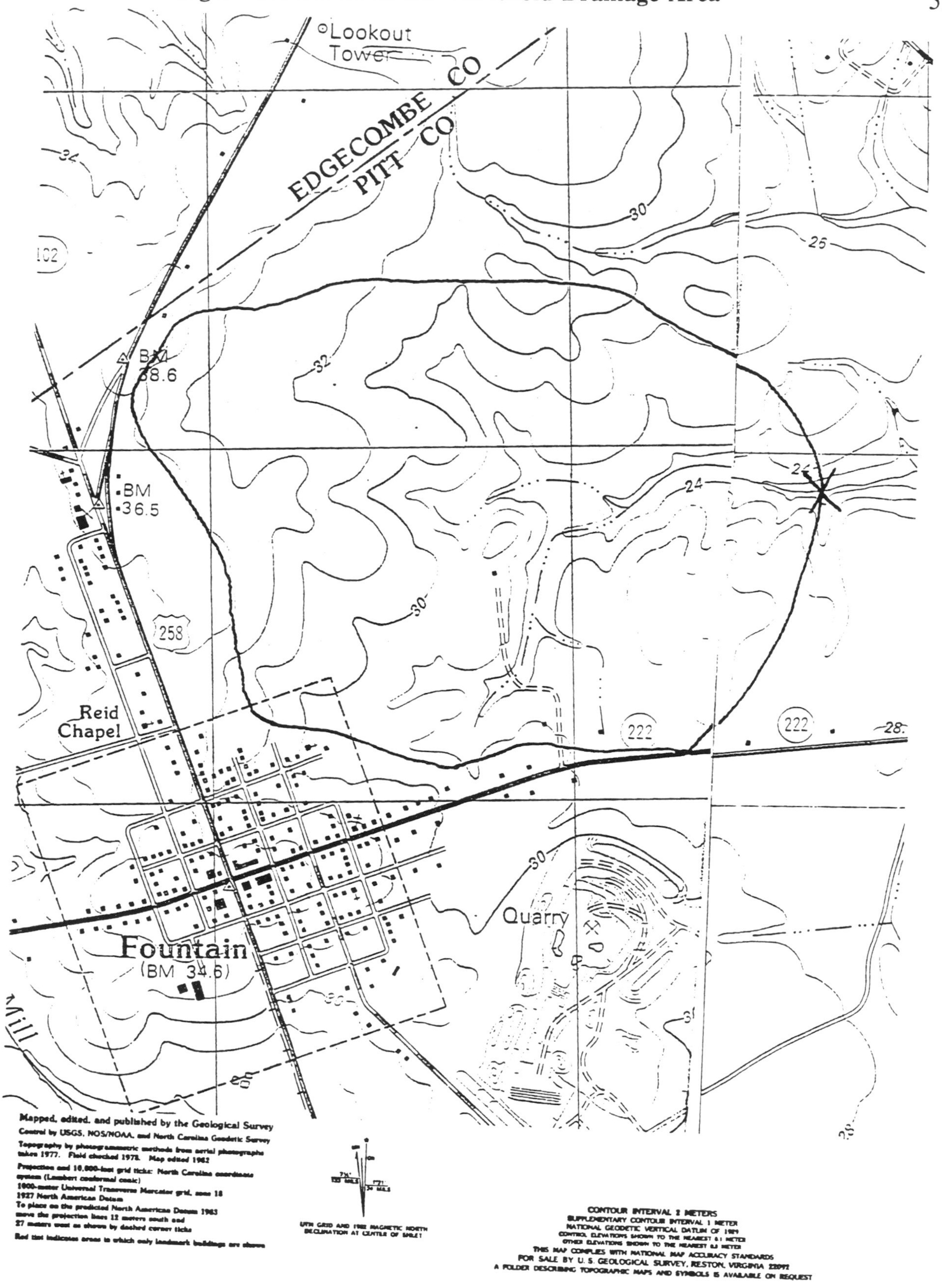


Figure 11. Kitten Creek Threshold Drainage Area



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CONTOUR INTERVAL 2 METERS
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Figure 12. Harris Mill Run Threshold Drainage Area

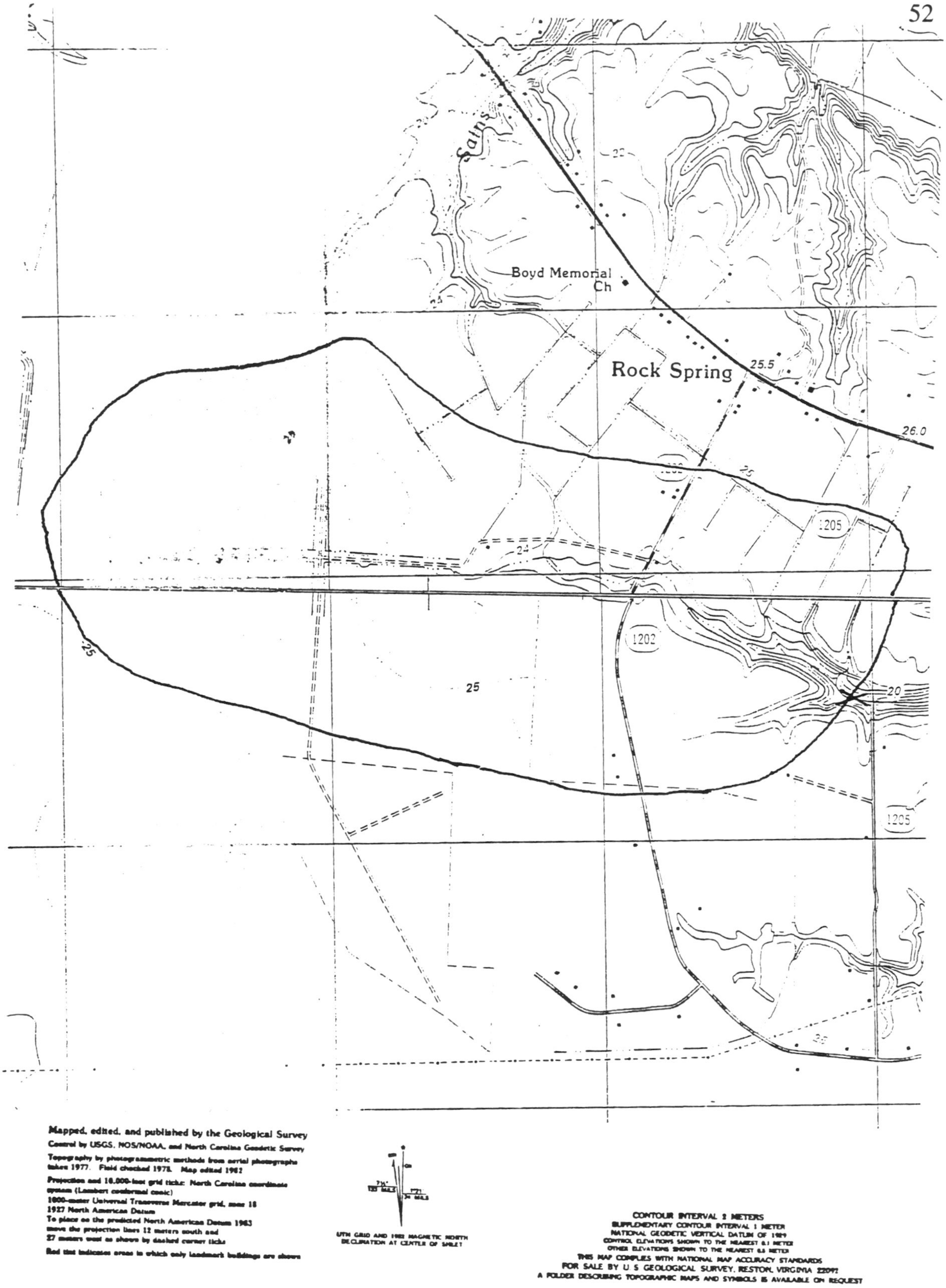

Table 4
Descriptions of the Upstream Limit of Alluvial Sediment Storage

Streams	Description of the Headwater Channels	Evidence for the Upstream Limit of Alluvial Sediment Storage in Coastal Plain Streams
Hardee Creek	The channel of this first order stream flowed through an agricultural field; Channel bottom consisted of course- grained sediments; Channel pattern was straight.	Downstream from the headwater location the stream began to meander through a densely vegetated forest. Indicators of long-term alluvial sediment storage were identified by the following: **1) Tributary delta of an aggrading stream 2) Delta consisted of fine-grain and organic sediments and 3) The tributary delta was vegetated.**
Tyson Creek	Tyson Creek headwaters originated in an agricultural field. The channel flowed into a forested area characterized by local upland depressions with seasonal high water tables. Channel incision into alluvium of varying composition was evident. There were no indicators or evidence of alluvial sediment storage through this reach and the channel bed consisted of a mixture of course- and fine-grain sediments.	Downstream, this second order creek was characterized by an anastomosing pattern. Channel bed consisted of fine-grain sediments. The upstream limit of alluvial sediment storage was identified by the following: **1) A vegetated point bar composed of organic fine-grain sediment 2) A partially submerged log indicative of long-term storage 3) The upper end of a floodplain.**
Otter Creek	The Otter creek headwater channel was located in an upland forest. The channel pattern was straight. No evidence of significant long-term alluvial sediment storage within this reach.	The channel began to meander downstream. Channel bed consisted of a mixture of course- and fine-grained sediments. Evidence of alluvial sediment storage for this third order stream was indicated by the following observations: **1) A root obstruction in the middle of the channel trapping sediment 2) A buried tire 3) Localized ponding created by the root obstruction 4) Floodplain deposition**
Kitten Creek	Kitten Creek headwater channel flowed through a densely vegetated forest. There were several depressions through this reach with seasonal high water tables. The channel bed of this first order stream consisted mostly of course-grain sediments. Roots from adjacent trees within the channel were exposed. There was no evidence of long-term sediment storage.	Downstream the channel pattern was meandering. The channel was characterized as a third order and the bed material was composed of coarse- to fine-grain sediments. The upstream limit of alluvial sediment storage was indicated by the following:**1) A vegetated point bar composed of fine-grain sediments 2) Upper end of a floodplain 3) Partially submerged log trapping sediments.**
Harris Mill Run	The Harris Mill Run headwater channel was located in an agricultural field. The channel pattern was straight and the bed material was composed of course-grain sediments. The channel was very narrow (< 1m) with little incision. There was no evidence of significant sediment sinks and roots from adjacent trees within the channel were exposed.	Downstream the channel meandered through an upland forest with seasonal high water tables. The bed material was composed mostly of fine-grain sediments. The upstream limit of long-term alluvial storage was identified by the following: **1) A partially submerged log that crossed the channel, which would not be submerged during bankfull flows 2) Floodplain deposition 3) A series of log steps trapping sediments downstream of this location.**

Discharge and Stream Order Analysis

There was an eight-fold variation in estimated bankfull discharge among the sites. Discharge ranged from 0.75 m^3 s^{-1} to 6.46 m^3 s^{-1}. Harris Mill Run and Hardee Creek had the largest threshold drainage areas with the lowest and highest discharges, respectively. Furthermore, there was no pattern with respect to A_t and discharge. Results indicate that discharge exerts no control over the threshold drainage area for streams in this study.

There was also no systematic relationship between characteristics descriptive of the threshold drainage area and stream order. Harris Mill Run, Kitten Creek, and Otter Creek were third order streams, while Hardee Creek and Tyson Creek were second order streams.

Mainstream and Total Stream Length

Mainstream length measured as the distance from the location of long-term alluvial sediment storage to the drainage divide at the head of the longest channel above it, ranged from 0.8 to 1.6 km. Total stream length, measured as the sum of the channels above the location of long-term alluvial sediment storage upstream, ranged from 1.5 to 3.05 km. With only a two-fold range in values for mainstream and total stream length of the channels, results suggest that distance from the headwaters to the upstream limit of alluvial sediment storage was consistent between all streams investigated. Additional analysis suggests that mainstream length was approximately one-half that of total stream length in each case.

Cross-Sectional Stream Power

Stream power is related to the time rate at which either work is done or energy is expended in the fluvial system. Stream power is dissipated in maintaining the fluid flow against flow resistance and in doing work by moving the saltating bedload (Bull, 1979). Scour of alluvium occurs where stream power is ample to transport an imposed sediment load. Where stream power is insufficient the stream bed will aggrade. Results show that at the threshold there was no relationship between stream power and the upstream limit of alluvial sediment storage.

Local Hillslope and Upstream Drainage Area

Consistent with the expectations of the model the threshold drainage area has a characteristic local hillslope to upstream drainage area ratio. Values for the ratio of local hillslope and upstream drainage areas for the streams included in this study ranged from .18 to .39 with a mean value of .30. The data confirm the notion that at the threshold, upstream drainage area increases relative to local hillslope area, since $A_h/A_d < 1$.

Results also suggest that, at the threshold, basin influences dominate over local hillslope influences, since $A_h/A_d < .5$ for all streams. These findings suggest that the ratio of local hillslope to upstream drainage area is a significant control over the size and location of the upstream limit of alluvial sediment storage.

CHAPTER V

INTERPRETATION AND DISCUSSION

Threshold Drainage Areas in Coastal Plain Streams

The upstream limit of alluvial sediment storage identifies a threshold drainage area between "small" headwater basins and "larger" higher-ordered ones. The size of the threshold drainage area represents an explicit spatial boundary indicative of the upstream-downstream differences in energy/sediment storage. The consistency in size, and apparent control by relationship between local hillslope and upstream drainage areas confirms the applicability of the threshold drainage area as a means for distinguishing between basin sizes for streams included in this study.

Influence of Discharge and Cross-Sectional Stream Power

With an eight-fold range in discharge between the sites and no apparent relationship with the size or location of the threshold, results demonstrate that discharge and stream power do not control the threshold drainage area. This may be viewed as an unexpected result, as discharge is largely responsible for the adjustment of channel geometry to imposed conditions and has been highly correlated with drainage area (Knighton, 1984). This finding may be explained on the basis that, even in small basins, discharge is not necessarily of constant frequency (Pickup and Warner, 1976). However, a likely explanation concerning the variation in estimated bankfull discharge involves the slope characteristics of the channels. In the North Carolina Coastal Plain slope gradients of zero to two percent are common and slopes greater than six percent are rare with the exception of valley sides and other isolated situations. Results for the channel slope gradients of the streams reveal an eleven-fold range of values from 1% to 11%. Therefore, results suggest that the spatial variation of local channel forms within the stream beds partially explains why there was no relationship between estimated bankfull discharge and the threshold drainage area.

Cross-sectional stream power is partially dependent on discharge and channel gradient. This also explains why there was no relationship with A_t. These results were also unexpected since cross-sectional stream power is the power per unit length of a stream available to transport the sediment load. At the threshold, where there is long-term alluvial sediment storage, cross-sectional stream power was expected to be consistently low for all streams, indicative of aggrading conditions. Results show that there was no relationship between the amount of power per unit length of a stream and the threshold drainage area.

While it is recognized that investigations which include a larger sample of basins might reveal correlations between bankfull discharge, cross-sectional stream power and the threshold drainage area, the data here suggest that there are no apparent relationships between these variables. However, this does not suggest that discharge and stream power should be disregarded. They were collected to determine whether their were any systematic relationships with A_t, since A_d is often used as a surrogate for such factors.

Effects of Local Hillslope and Upstream Drainage Area

One of the main objectives of this research was to test a proposed model which suggests that the upstream-downstream distribution of energy/sediment storage is a function of local hillslope and upstream drainage area. Furthermore, the model developed for this study suggests that the size of the threshold drainage area should demonstrate some general consistency where Hack's (1957) empirical proportional relationship holds (Equation 15). Relationships between the plots of $1/Aa_d^b$ and the size of $A_t < 10$ km^2, demonstrate this consistency.

It was also hypothesized the threshold, $A_h/A_d < 0.5$ for all streams. These values were consistent with the expectation that upstream drainage area increases relative to local hillslope area in the downstream direction and the intuitive proposition that basin influences begin to dominate over local hillslope influences at the threshold. This suggests that the size and location of the threshold drainage area is strongly influenced by the local and upstream hillslope subsystems. Comparisons between the size and extent of the threshold drainage areas indicate consistency among the streams that were analyzed. The threshold drainage area for all of the streams was < 10 km^2 with a range of less than two-fold between the streams, as expected from Hack's (1957) relationship. The hypothesis concerning A_h/A_d was supported. Larger values were associated with smaller threshold drainage areas. Smaller values represented larger threshold drainage areas. This demonstrated a consistent relationship between A_h and A_d at the threshold, while there is no evidence of a systematic relationship between A_t and bankfull discharge stream order or stream power. Overall, the results suggest that the upstream limit of alluvial sediment storage is controlled by the relative importance of local and upstream hillslopes, and that a threshold drainage area occurs where $A_h/A_d < 0.5$.

Clarke and Waldo (1986) and Graf (1987) failed to acknowledge the role of energy dissipation as an indicator of differences between small and large drainage basins. However, fluvial geomorphologists have commonly recognized the importance of energy distribution in the development of theoretical models concerned with the search for a general principle governing stream behavior and channel form adjustment (Langbein and Leopold, 1964; Yang, 1971; and Knighton, 1984). Clearly, this theme is manifested in the model developed and tested for this study. Thus, identifying threshold drainage areas has several important implications.

Implications for Modelling Sediment Storage in Fluvial Systems

The modelling of sediment movement on a river-basin scale is in a primitive state (Meade, 1982). Many scientists have recognized that the storage and routing of sediment eroded and transported by fluvial processes is a complex problem (Meade, 1982; Walling, 1983; Phillips, 1992; Sutherland and Bryan, 1991). Furthermore, Meade (1982) has recognized that for time spans of years to centuries scientists are least prepared to construct predictive models concerning sediment movement. It has been suggested that the storage of sediment and the time periods over which the sediment goes into and out of storage are the most important factors in understanding the movement of sediment in rivers (Meade, 1982).

Identifying threshold drainage areas has implications for understanding sediment storage patterns and can perhaps be a useful tool for developing predictive models of sediment storage and movement in fluvial systems. Because the results of this study suggest that the threshold drainage area is controlled by the local and upstream hillslopes the location of the upstream limit of alluvial sediment storage is strongly influenced by the variables that influence the transport or storage of sediments (i.e. gradient, stream power, velocity) within the local and upstream hillslope subsystems. Therefore the upstream location of alluvial sediment storage will be sensitive to changes in the local and upstream hillslopes. Changes in the local and upstream hillslopes might reflect changes in the feedback relationships between streamflow, and channel characteristics, as well as sediment transport and storage effectively altering the upstream location of alluvial sediment storage. Simulating changes in variables which control the storage and transport of sediment within the upstream and local hillslope subsystems might cause corresponding changes in the amount of sediment being transported or stored at the threshold. Applying these simulations to the local and upstream hillslopes could perhaps be beneficial to estimating changes in the location of the upstream limit of alluvial sediment storage under a variety of human or environmentally induced impacts over a broad range of time scales. Scientists could then use the threshold drainage area as a region where storage changes could be calculated within a watershed.

Implications for Environmental Management

Fluvial geomorphologists concerned with the study of sediment erosion, storage, and transport in watersheds, commonly use sediment budgets to account for sediment entering, leaving, and being stored within the fluvial system. Fluvial sediment budgets are useful for regulating soil conservation and sediment pollution control needs and are based on two assumptions. The assumptions suggest that the portion of eroded sediment delivered to the channel system and the river system's sediment conveyance capacity remain constant under extremely diverse land management scenarios (Phillips, 1985). However, human impacts on stream channels that affect the fluvial hydrology and result in changes in stream power may invalidate these assumptions (Phillips, 1985). Phillips (1985) maintains that the sediment budget is a useful tool if the fluvial sediment system can be assumed to be qualitatively stable and the stream sediment transport capacity is constant. However, Phillips (1985) also suggests that the sediment budget can be a more powerful tool with better information about sediment storage within watersheds. Results of this study provide insight into the storage of sediments in watersheds since the threshold drainage area is indicative of the upstream-downstream differences in energy/sediment storage. Furthermore sediment budgets should be specific and distinguish between alluvial floodplain storage and channel storage because floodplain ecosystems often have the capacity to store, remove, or buffer nonpoint-source pollutants (Phillips, 1985). Differences between long-term alluvial floodplain storage and short-term transitory channel storage are substantiated by identifying threshold drainage areas. Thus, identifying threshold drainage areas has implications for fluvial sediment budgets and improving environmental management techniques.

Future Studies

Numerous studies have investigated relationships concerned with the morphometry of drainage basins and have shown how process-response relationships may vary with spatial scale. This study has addressed the question of defining small and large drainage basins through the concept of the threshold drainage area. Specifically, the purpose was to show that differences in energy/sediment storage between small and large basins suggests there is a unique threshold drainage area.

Although techniques of analysis in geomorphology have become more sophisticated, methods remain largely inductive with the attendant problem of making generalizations from empirical results founded usually on a limited sampling base (Knighton, 1984). Despite these shortcomings, the empirical approach has been dominant in geomorphology and has provided valuable insights into the working of the fluvial system (Knighton, 1984). Similarly, this study was based on a limited sample of five basins where results showed that the upstream limit of alluvial sediment storage is indicative of a threshold drainage area, controlled by local and upstream hillslopes. This study was established on: (1) the theoretical grounds of thermodynamic principles which explicitly demonstrate the role of energy/sediment storage as an indicator of differences in system behavior, and (2) the empirical power-function relationship between mainstream length and drainage area which suggests that at the threshold there is a characteristic local hillslope/upstream drainage area ratio. The latter is an example of a formal analysis used for investigations that are not deductively rigorous, either because there is insufficient theoretical basis, or because the theory leads to intractable equations (Church and Mark, 1980). However, Church and Mark (1980) suggest relations derived in this way may represent prologues to theory formulation and thus future studies concerned with the threshold drainage area will be important for advancing the results of this research.

Prior to this study, little was known about the size and extent of the threshold drainage area. Scientists recognize that there are many problems associated with modeling a complex physical system. With the ultimate objective of explaining the behavior of natural streams, a great challenge for fluvial geomorphologists is to collect the necessary data to establish empirical relationships and to test models (Knighton, 1984). Therefore, future research concerning the threshold drainage area should involve collecting data for a large sample of basins. This will provide researchers with an opportunity to test the model presented in this study and determine if the results are reliable and accurate.

There are several issues that need to be explored. The first concerns the relationship between the location of the threshold drainage area and the influence of local hillslope and upstream drainage area. In one sense local hillslope and upstream drainage area reflect the complex interactions which influence the transport and storage of sediment over time and space in fluvial systems. In another, they have causative significance for determining the efficiency for which the controls of energy/sediment are stored or dissipated in fluvial systems. This study stresses the latter, more broad viewpoint. However, what remains to be explored is the question of whether the threshold drainage area is specifically controlled by individual or a combination of the detailed interactions which influence sediment transport and storage. Results of this study indicate that discharge exerts no systematic control over the threshold drainage area. However, applied to an adequate sample of basins, bankfull discharge and/or other factors that influence the transport and storage of sediments might reveal different results. Furthermore, studies should be concerned with recognizing the spatial and temporal patterns of discharge in relation to the threshold drainage area.

Second, future work concerning the threshold drainage area might involve establishing relationships with the threshold of critical power in streams. Bull (1979) has developed the concept of the critical threshold of stream power, which permits the determination of whether degradation, aggradation, or a graded condition may exist in a stream channel for a given imposed sediment load. The critical-power threshold occupies a key position in the complex interactions between the hillslope and stream subsystems and is affected by feedback mechanisms and complex responses operating in either subsystem (Bull, 1979). The critical-power threshold in relation to the threshold drainage area is significant for understanding the relative control and importance of local and upstream hillslopes. Furthermore, Bull (1979) suggests that thresholds can be used in studies involving minutes to million of years. If relationships could be established with the threshold drainage area it would be directly beneficial for understanding channel characteristics and sediment storage in watersheds on time spans of years to centuries over which scientists are least prepared to construct predictive models (Meade, 1982) and over which little is known about sediment movement through watersheds (Meade, 1982; Knighton, 1984).

Finally, there is a clear need to identify threshold drainage areas in a variety of environmental and climatic settings. This would allow for a comparison of threshold drainage areas in locations with different environmental controls. Furthermore, this would provide the identification of the range of threshold drainage areas. The results presented here indicate that the average size of the threshold drainage area was 2.66 km^2. Future studies are needed to verify if these findings are accurate and reliable. It is recommended that future research focuses on relationships between the threshold drainage area and the factors which influence sediment transport efficiency within the local and upstream hillslope subsystems such as climate, relief, channel slope, lithology, land use, and vegetation. Still, little is known about the size of threshold drainage areas and therefore their identification in different environmental and climatic regions would provide additional insight into their relative size and existence.

CHAPTER VI

SUMMARY AND CONCLUSIONS

The primary focus of this study was to determine whether threshold drainage areas separating "small" headwater basins from "larger" higher-ordered ones were evident in Coastal Plain Streams within the Tar River Watershed. The size of the threshold drainage area ranged from 2.08 km^2 to 3.47 km^2 with a mean size of 2.66 km^2. Because $A_h/A_d < 0.5$ for all streams, results confirm that at the threshold upstream drainage area increases relative to local hillslope area and basin influences dominate over local hillslope influences. Furthermore, these results suggest that at the threshold there is a characteristic increase of upstream drainage area relative to local hillslope area. This indicated that the size of the threshold drainage area is controlled by the ratio of local hillslope to upstream drainage area. Hack's (1957) area-length relationship is an important theoretical component of the local hillslope/upstream drainage area model, and suggests similar threshold areas in the wide range of environments for which $L = Aa_d^b$ holds.

Characteristics descriptive of the threshold drainage area were consistent among all of the streams regardless of stream order. Therefore, it is suggested that stream order had no relationship with the threshold drainage area for the streams investigated in this study. Furthermore, variation in discharge and cross-sectional stream power estimated for the five streams demonstrated no apparent relationship between the size or location of the threshold drainage area. This suggests that at the threshold it is indeed the ratio of hillslope and upstream drainage area and not just a matter of area as a surrogate for discharge and stream power.

Identification of the upstream limit of alluvial sediment storage in Coastal Plain streams within the Tar River watershed resulted in several conclusions concerning threshold drainage areas. The conclusions are summarized below.

- Drainage areas associated with the upstream limit of alluvial sediment storage identify threshold drainage areas that are representative of the upstream-downstream differences in energy/sediment storage between "small" headwater basins and "larger" higher-ordered ones.
- The size of the threshold drainage area is controlled by local hillslope and upstream drainage area. Larger threshold drainage areas were related to a greater increase of upstream drainage area relative to local hillslope area while smaller thresholds were a result of a slight increase of upstream drainage area relative to local hillslope area.
- Consistencies between the size of the threshold drainage area and the proportional increase of upstream drainage area relative to local hillslope area suggests that local hillslope and upstream drainage area have causative significance for determining the location of long-term alluvial sediment storage within the fluvial system.
- Hack's (1957) mainstream length/basin area relationship implies, that the local hillslope/upstream drainage area model may be applied to a wide range of environments for which $L = Aa_d^b$. Hack's (1957) relationship is therefore an important theoretical component of the local hillslope/upstream drainage area model.
- Estimated bankfull discharge and cross-sectional stream power exerted no control and demonstrated no apparent relationship with the threshold drainage area which suggests that at the threshold area was not a surrogate for those factors.

This thesis has attempted to provide additional insight into the size and extent of threshold drainage areas. Scientists have recognized that there are many problems associated with modeling a complex physical system. With the ultimate objective of explaining the behavior of natural streams, a great challenge for fluvial geomorphologists has been collecting the data necessary to establish empirical relationships and test models (Knighton, 1984). Relationships established in this study suggest that the threshold drainage area is controlled by the relative importance of local and upstream hillslopes and a threshold drainage area occurs where $A_h/A_d < 0.5$. Future research is needed, however, to advance the ideas presented in this thesis to determine if there are specific controls within the hillslope subsystem which control the size and extent of the threshold drainage area.

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