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THE DOWNSTREAM CHANNEL FORM CHANGE OF STREAMS ON DIFFERENT GEOLOGICAL SETTINGS

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THE UNIVERSITY OF OKLAHOMA

GRADUATE COLLEGE

THE DOWNSTREAM CHANNEL FORM CHANGE OF STREAMS ON DIFFERENT GEOLOGICAL SETTINGS

A DISSERTATION

SUBMITTED TO THE GRADUATE FACULTY

in partial fulfillment of the requirements for the

degree of

DOCTOR OF PHILOSOPHY

By

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Department of Geography, College of Geosciences

Norman, Oklahoma

November 1932

THE DOWNSTREAM CHANNEL FORM CHANGE OF STREAMS

ON DIFFERENT GEOLOGICAL SETTINGS

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THE DOWNSTREAM CHANNEL FORM CHANGE OF STREAMS
ON DIFFERENT GEOLOGICAL SETTINGS

by

Mahmoud Diab Radi

An Abstract

of a dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy in the Department of Geography in the Graduate College of the University of Oklahoma

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Dissertation Supervisor: Professor Neil E. Salisbury

Abstract

This research attempts to show how geology (lithology) and man control downstream channel form change. It was hypothesized that a change from sand to shale in the Middle Permian beds of western Oklahoma produces profound differences in channel characteristics.

Three subhypotheses were constructed following a review of the literature and development of a theoretical framework. First, it was hypothesized that the force-resistance index predicts channel cross-section area more appropriately than either tractive force or resistance offered by riparian vegetation. Second, total stream length upstream is important in predicting network volume, the prediction of which is a useful parameter for comparing flood-producing potentials for watersheds treated with flood detention dams. Third, weighted mean silt-clay percent in channel perimeter is a good estimator of channel width-depth ratio. Geology was included in the analysis as an indicator of the spatial distribution of channel form.

The t-test used in the analysis of the subhypotheses revealed significant differences between the slopes and intercepts of regression lines of sandstone streams and shale streams at a 0.005 significance level.

Combining the three successfully tested subhypotheses enabled estimation of the channel cross-section area fairly consistently. However, geology as an indicator factor was important in determining the spatial variation of downstream channel form change. Evidence of the effects of varying lithology on channel form sheds light on the environmental problems in fluvial research, and signals precautions necessary when results of research in one region are applied to another.

Abstract approved by Neil E. Salisbury
Dissertation Supervisor

Title: Professor and Chairman of Geography Department

Date: November 22, 1982

CHAPTER I

THE DOWNSTREAM CHANNEL FORM CHANGE OF STREAMS ON DIFFERENT GEOLOGICAL SETTINGS

Introduction

Geomorphology in the United States has become, since the early 1970's, increasingly included within the realm of geography. This particular field of physical geography has witnessed considerable attention paid to fluvial processes and water-produced landforms (Gregory & Walling, 1973). The focus has been on the spatial variation of phenomena. Attention has also been directed to the aspects of river channel forms, processes, and change which are potentially of an applied nature (Cooke & Doornkamp, 1974).

Rivers warrant geographical study for three main reasons. Firstly, because of their existence in the physical landscape and their significance for producing fluvial landforms; secondly, because of their importance indirectly in relation to many other geomorphological processes in fluvial-dominated landscapes; and, thirdly, because of their significance for human use (Gregory & Walling, 1973). Therefore, the growth of interest in all aspects of stream channels has included greater interest in geomorphometry and stream adjustment.

The objective of this study is to determine if geology (lithology) can have an overriding control on catastrophic response and allometric change in detecting the downstream channel form change.

Several substantive and significant research works on the general morphology and hydraulics of stream channels published in the 1950's served as catalysts to the many important publications of the following decade. A research on the relationships between channel form and process was crystallized by Leopold and Maddock in 1953. In 1956, Leopold and Wolman distinguished between channels in semiarid areas, which tended to scour at high discharges, and channels in sub-humid areas in which repetitive scour and fill sequences were not demonstrated (Leopold & Wolman, 1956). Attention was also being directed towards river channel patterns at this time, and the same workers classified streams into straight, meandering, and braided forms (Leopold & Wolman, 1957), stressing that each type represents an equilibrium condition of a specific form. Complementary research on river flood plains and their formation by Wolman and Leopold (1957) yielded valuable evidence on general problems of stream channel equilibrium and change. Factors controlling stream slope in the Appalachian region of the United States were studied by Hack (1957), who established relationships between slope, distance downstream, and bed material size in the manner of the earlier work of Sternberg (1875).

Furthermore, Gregory and Walling (1973) identify three main ways in which form-process relationships have been studied. Firstly, a group of studies evolved from early work by engineers in the latter part of the 18th century of the movement of water and sediment in channels that led to the development of basic physical relationships such as those of Chezy, Manning, Froude, and Reynolds. These were later complemented by Gilbert (1914) in his flume experiments on the movement of water and sediment in the channel, and later culminated in experimental and theoretical research on the relationships between measures of channel form and process under the nomen of hydraulic geometry (Leopold & Maddock, 1953; Wolman, 1955; Leopold & Miller, 1956). Secondly, work on river pattern based on flume and field investigations has been prominent since the work of Horton (1945) on channel pattern hierarchy and Inglis (1949) on meanders. Some geomorphologists have identified a branch of study known as alluvial morphology (Wolman & Leopold, 1957) which is concerned with deposition characteristics of floodplains. Thirdly, at the drainage basin level, the work by engineers (Horton, 1945), and hydrologists (Guy, 1970), and geomorphologists (Wolman, 1967) has been carried out with the intent of understanding how runoff and sediment yields vary in response to soil erosion, changing vegetation cover, and the influence of man. This thrust is most clearly associated with drainage basin process-response systems. This research attempts to provide an understanding and sensitivity to all three of these foci with particular emphasis on the first (hydraulic geometry) and the second (alluvial morphology) of the three. The shape of the cross-section of a stream channel at any location is in part a function of the characteristics of the contributing area (drainage basin-discharge relationship), the quantity and character of the sediment in movement through the section, the character or composition of the materials making up the bed and banks of the channel, and man induced changes in the system.

A number of reviews of advances in geomorphology (Brown & Waters, 1974; Cooke, 1969; Gregory & Walling, 1973, 1974b; Jennings, 1974; Zakrzewska, 1967; Butzer, 1973; Dury, 1969a, 1972a, and 1972b), quantitative geomorphology (Salisbury, 1972), and sedimentology (Allen, 1971) have stressed the present importance of fluvial morphology, which can be defined as the quantitative expression of fluvial landform geometry. Although the movement towards morphometric studies gained momentum rapidly in the 1960's, its early history was somewhat turbulent. Statistical methods had been accepted into geological research much earlier (Krumbein & Graybill, 1965) however, and the beginnings of the 'quantitative revolution' in geography as a whole have been dated to the late 1940's (Burton, 1963). Wooldridge (1958) recognized the illustrative value of morphometric work but warned against the urge to dabble in statistics, while Baulig (1950) championed the cause for exercising caution and proclaimed that it 'remains to be proved that mathematics has ever revealed an actual relationship in geomorphology that had not been discovered without its aid.' Mackin (1964) gave approval in principle, but he feared that 'the swing to the quantitative has been too far and too fast,' whereas Chorley

(1956) welcomed quantitative geomorphology, and what he considered to be 'the conservative techniques of mathematical statistics' (Chorley, 1958). Also, Salisbury (1972) pointed out that quantitative techniques were originally introduced largely in fluvial geomorphology, then with some opposition were widely accepted into all phases in geomorphology. By the end of the 1960's, therefore, morphometric studies in fluvial geomorphology were a component part of the discipline of quantitative geomorphology (Charlier, 1968; King, 1971). The present morphometric channel study is viewed as a contribution of this trend.

An extensive study of the literature revealed that whereas a number of studies of channels had been undertaken using the hydraulic geometry approach, fewer studies are available of the spatial form and pattern of channel geometry due, in part, to spatial differences of lithology in a region. Patton and Schumm (1981) indicated [it is still unresolved whether channel cutting and filling in the alluviam is primarily caused by (a) climatic fluctuations (Bryan, 1940; Euler et al., 1979) either to a more humid or to a more arid climate, (b) changes in the relative frequency of rainfall magnitudes (Leopold, 1951), (c) poor land management practices, specifically overgrazing (Swift, 1926; Hastings, 1959; Cooke & Reeves, 1976), or (d) natural geomorphic-sedimentologic processes within a range of climates and over time spans independent of land management practices (Thornthwaite et al., 1942; Schumm

& Hadley, 1957; Patton & Schumm, 1975). Probably all of the above are causes but the nature of the prevailing climate and the geomorphic character of the drainage basin can strongly influence which causative factor will be most important]. Physical and/or biological factors are, therefore, the focus of investigation in this study.

Furthermore, human use of natural streams affects greatly the erosional and depositional mechanics. Channelization and using unnatural bank stabilization structures generally reduces the aesthetic quality of a stream. River engineering is a controversial practice used most often in an attempt to control flooding. This engineering practice, such as channel modification, deals with dynamics and properties of the river. Channelization has been used to improve control of stream-bank erosion, or improve the stream alignment relative to a bridge crossing. Terms synonymous with channelization include channel or drainage improvement, channel modification, and channel works. More recently, a new form of channelization, known as channel restoration, is being initiated on stream channels that have been disrupted by previous channelization, catastrophic events, or unwise landuse (Keller, 1976).

Man-induced changes in the natural system of a river also includes installation of floodwater retarding structures. The construction of these dams is part of the watershed treatment to reduce losses by floods. Reserach to determine the effect of floodwater retarding structures on watershed sediment yield and channel stability was initiated because excessive sediment yield and channel instability were widespread problems in the

Southern Plains (Allen, 1979). From the geomorphologists viewpoint, however, the operation of a river or stream is not so simple (Leopold ėt al., 1964; Morisawa, 1968; Cooke and Doornkamp, 1974). In addition to protecting property, the processes and environment of the river itself are of concern. The land managers, from a geomorphic viewpoint, should consider preservation of the balance of river processes while determining landuses compatible with various human needs (Palmer, 1976). A major research effort on watershed treatment by the Southern Plains Research Watershed which established its headquarters at Chickasha, Oklahoma, in 1961, has been conducted on the central segment of the Washita River basin in western Oklahoma. This extensive research effort was carried out on the effect of the flood control and watershed protection program presently in an advanced state of development. The research project has contributed more than 100 publications to the hydrologic literature. This literature reveals that the area is witnessing spatial changes in channel form characteristics caused by conservation practices of floodwater retardation, devegetation, and the differential erosiveness of materials throughout the area (Kennon, 1966; Allen, 1979).

There is a need for a study of channel geometry in Western Oklahoma to complement the on-going geomorphological research along with the hydrological research which was initiated previously. Such an investigation appears to be necessary to further our understanding of the spatial pattern of stream channel form in a study area controlled by differential geologic formations, and modified by man. Human impact on channel form such as channel straightening are dominent in Western Oklahoma. Also land owners cut through many meander loops to alleviate bank erosion, shortening some reaches by 40 percent (Allen, 1979). This has led to the formulation of the following statement of the problem and hypotheses.

Statement of the Problem

The purpose of this dissertation is to examine the role of geology (alluvium derived from sandstones versus shales) and man in controlling channel form in the downstream direction. In other words, the study attempts to establish a causal model, for natural controls of channel form under the influence of geology, and is examined both in man-modified versus unmodified streams.

Hypotheses

This dissertation is structured around the following general hypothesis; differing stream dynamics and morphology will be found in areas of contrasting lithology. This hypothesis will be examined in Western Oklahoma using the Permian age Dog Creek and Rush Springs sandstone. Also, the consequences of direct and indirect interference by man with stream channels are increasingly important for understanding stream morphology. This requires a marriage between knowledge of natural stream behavior and investigations of induced channel changes. Three specific hypotheses are stated to evaluate current concepts of channel form change as applied to streams in the study area.

<u>Hypothesis One</u>

Streams on sandstone are best considered as candidates for the application of a process which includes abrupt change as thresholds are crossed.

Considerable rangeland in the study area was converted to cropland about the time of World War I because wheat and cotton prices were high (Allen, 1979). Landowners cut through many meander loops to alleviate bank erosion, thereby shortening stream lengths, straightening some reaches, and cleaning stream channels of riparian vegetation. The distribution (the change over space) of the present riparian vegetation along stream channels can be considered as a factor that separates modes of operation (aggradational and degradational modes) along the stream channel (Bull, 1980).

> A stream reverses from an aggradational to a degradational mode of operation in response to a perturbation; 'removal of vegetation.' Although thresholds generally cannot be forecast a priori, the passage of a threshold is readily identifiable in the field.

The change of vegetation over space, therefore, seems to be particularly well suited for studies involving the interaction of human activities with the fluvial system.

<u>Streams cut in the Sandstone are sensitive to removal of vegetation by cultivation particularly when</u>

stream channels are cleaned and lined. To test this hypothesis tractive force is calculated. If dense vegetation cover decreases the tractive force condition, then channel cross-section area is a function of the ratio between tractive force and vegetation resistant index. Regression analysis should generate significant correlation between channel capacity and the ratio between tractive force and vegetation in channelized streams.

Hypothesis Two

Streams on shale are best considered from an allometric rather than from a steady state viewpoint. The density of the drainage network is perhaps the most frequently cited single index in fluvial geomorphology studies. Sokolov (1967) noted that the density of the drainage network has a fundamental influence upon flood generation (Gregory, 1977). In addition, the type of stream channel as well as its density will affect sediment yield. The importance of drainage density arises first from the fact that channel flow is more rapid than other forms of water flow in the basin (Weyman, 1975), so that stream density reflects the potential rate at which water can be transmitted through a basin. Secondly, drainage density will reflect the climatic features of a particular area and also its basin characteristics, including rock type.

Channel cross-sections below a dam are frequently compound with an active inner channel that is separated from older pre-reservoir stream banks by vegetated berms or benches of fine sediment on one or both sides. Flow regulation by the reservoir from the time of completion of a dam should thus lead to a drop in channel capacity. Plotting bankfull cross-section areas of both inner and outer channels against catchment area, Gregory and Park (1974) showed that the reduction in channel capacity is substantial immediately below the dam, but becomes progressively less marked downstream as the fraction of the catchment controlled by the reservoir decreases (Lewis, 1981). Also, Allen (1979) indicated that small channels in the study area immediately below floodwater retarding structures tend to aggrade. This situation is just the opposite of the effect produced by large structures (Leopold et al., 1964, pp. 454-456). Aggradation occurs because peak flows are greatly reduced, reducing even further the sediment transporting capacity (transport increases with approximately the square of the discharge, Allen 1979). Also, prior to structures most of these channels were ephemeral. After structures were built these channels had prolonged base flow and a lush growth of aquatic plants began that encouraged sedimentation (Schoof et al., 1979). In this regard a hypothesis is constructed as:

<u>Channel capacity increases in the downstream direction as the total stream length, above the cross-section site, increases. Regression analysis between channel capacity</u>

(channel cross-section area) and total stream length (measurements of stream length above the cross-section site) should confirm the allometric change, of streams on the Shale Formation, in the downstream direction. Also, the regression analysis should generate low correlation between channel capacity and total stream length in streams treated by flood detention dams.

Hypothesis Three

Another concept views the dimensional characteristics of the channel as the result of erosional and depositional mechanics due to differential cohesion of the alluvial sediments (Schumm, 1961, 1963, 1977). The shape of the channel (width-depth ratio; F), therefore, should be related to the percentage of silt and clay in the sediments forming the perimeter of the channel (M).

Channel erosion into sandstone may show that the channel extends headward by headcut migration. The perpetuation of a headcut could be faced by some resistant material capping the alluvium. This resistant cap may be formed by the binding action of the plant roots, or a veneer of silt and clay overlying the coarser materials. Also, another important phenomenon that occurs during the development of a stable channel is bank carving. The relative amount of bank carving along the stream banks is in part related to the cohesiveness of sediments along the banks. In shale, however, if percent silt-clay remains approximately constant in the downstream direction then the width and depth of the channel increases at a uniform rate with discharge, excluding the influence of other variables, and width-depth ratio will remain constant and channel cross-section area increases in the down-stream direction.

Channels composed of sediment containing high silt-clay percent are narrow and deep; whereas those channels containing little silt-clay percent are wide and shallow. Regression analysis should confirm the validity of the negative relationship between width-depth ratio and weighted mean silt-clay percent. Also, the analysis should generate low correlation in streams on shale. Furthermore, a discriminant analysis of silt-clay percent should differentiate between channel shape (width-depth ratio; F) for streams on the sandstone and channel shape for streams on shale.

This study is characterized as a hypothetico-deductive system, as staed by Braithwaite (1960, p. 12) (Harvey, 1971, p. 36):

> A scientific system consists of a set of hypotheses which form a deductive system; that is which is arranged in such a way that from some of the hypotheses as premises all the other hypotheses logically follow. The propositions in a deductive system may be considered as being arranged in an order of levels, the hypotheses at the highest level

> being those which occur as premises in the system, and those at intermediate levels being those which occur as conclusions of deductions from higher level hypotheses and which serve as premises for deductions to lower level hypotheses.

Nevertheless, the study seeks to establish an inductive model. This model involves the use of probability statements within a deductive system. As Harvey (1971, p. 37) stated:

> When we speak of the scientific method as entailing deductive inference, we must necessarily exclude the problem of confirmation of theory as well as probability statements. In these two areas we are firmly in the field of induction.

The confirmation of the initial hypothesis proceeds by an alternation of deductive and inductive steps. For example, the study is proposed to test three relationships (hypotheses); and to do this by (1) making deductions which are testable, (2) designing an experiment to collect suitable data, (3) analyzing this data to a form which is comparable with the deductions from the models, and (4) testing. If the test is satisfactory, then the relationships are not disproved and may be checked against other areas: otherwise the models must be modified using fresh concepts to refine them (Carson & Kirby, 1975, p. 8).

The emphasis on both process and channel morphology will expand the explanatory power of the deduction in this study. This in turn leads to confirm the genetic approach as an appropriate method in explaining the spatial distribution of stream channel morphology. Abler et al. (1971, p. 51) pointed out a possible association between genetic and deductive explanation as:

> Whether we choose to use genetic or deductive explanation in a particular instance like this is essentially a matter of choice and temperament. At some stages of science, choice may not exist. Constructing a number of genetic explanations and then noting the similarity between them and using such similarities as bases for formulating laws is a fairly common way of developing law and theory in science. But one will never develop law and theory unless he is consciously trying to do so and is searching for what is common in the individual explanations being constructed.

Subsequently, the controlling factors of channel form adjustment are considered within the framework of a conceptual model of the research problem (Figure 1-A). In other words, channel geometry depends upon geology, man induced changes, and other related factors, such as those included in Figure 1-B. Table 1 shows the scales and environment factors of adjustment of streams on different geologic formation. Whenever any of these factors change effects are felt in the dynamics and morphology of the system. The conceptual model can be used (1) to assess the influence of geologic setting on the behavior of the fluvial system, and (2) to evaluate current concepts of channel form change such as that of allometric change (Bull, 1975; Gregory, 1977; Park, 1978; Osterkamp, 1979), of catastrophic change (Graf, 1979a, 1979b), and of fluvial system adjustments to internal stimulation (Schum, 1961, 1963, 1977) as applied to streams in Western Oklahoma (see model construction in Chapter V).

THE RESEARCH PROBLEM FLOWCHARTS

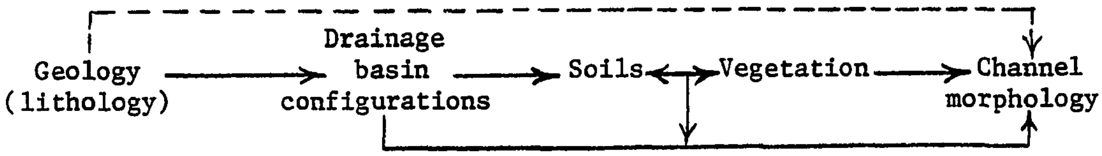
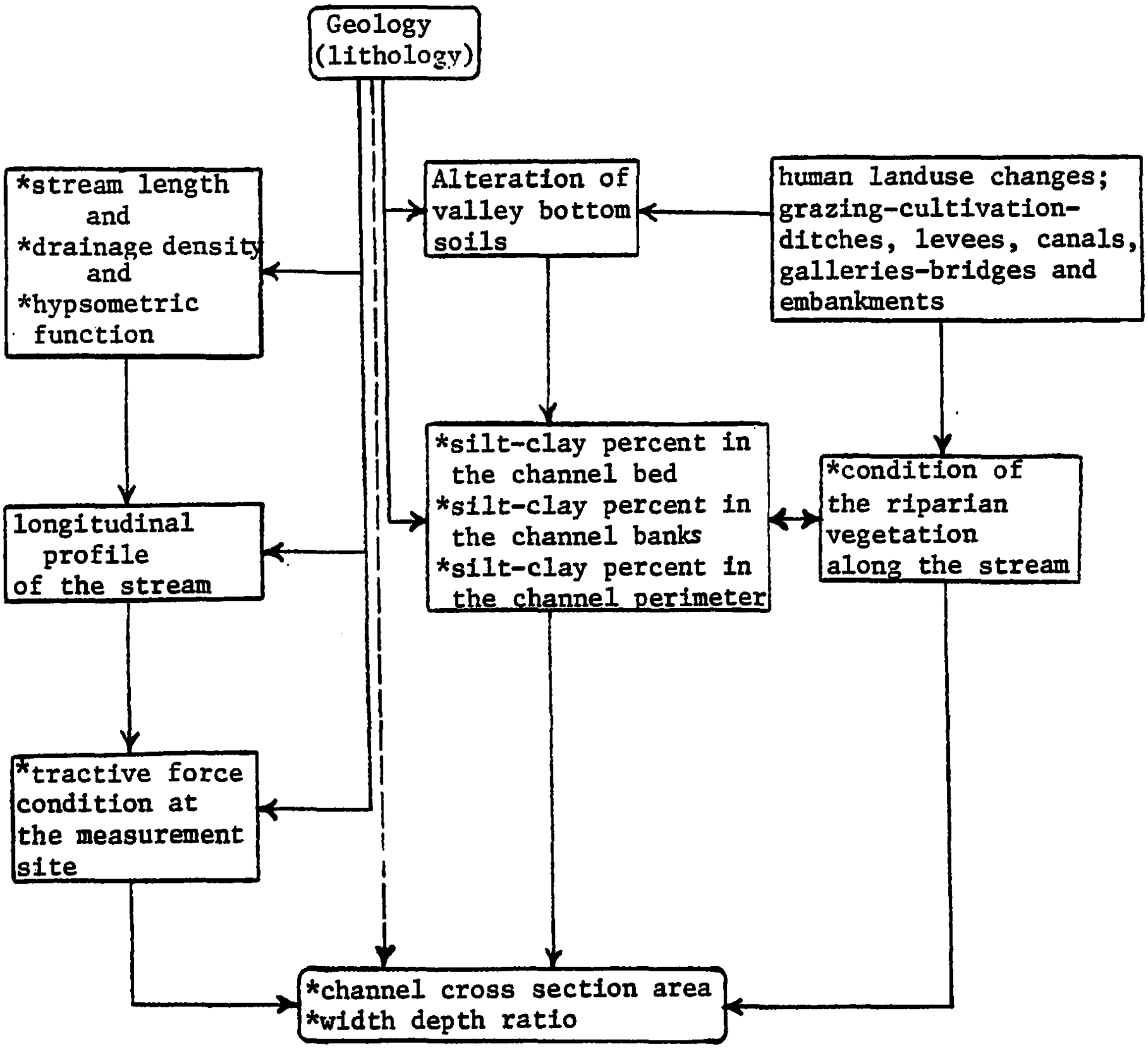


Figure 1-A



*The measured parameters

Figure 1-B

TABLE 1

THE SCALES AND FACTORS OF ADJUSTMENT OF STREAMS ON DIFFERENT GEOLOGICAL FORMATIONS

	Scale			Variables (independent)				Results (dependent)	
Geology	Type	Range of Change	Common time span for the type of change*	Biologic	Sedimentologic	Morphologic	Hydrologic	Morphologic	Sedimentologic not considered as a variable in the study
Friable Sand-stone Formation	Changes within & adjacent to channel	Less than or equal to depth of channel incision	0.01-100	Natural vegetations in species along the stream channels, and converting rangeland to cropland	Changes in reliability and availability of sediment, types of sediment, available storage along channel, facies in equilibrium with channel condition-silt-clay content in the sediment (noncohesive)	Local changes in slope, hydraulic geometry	Variations in discharge flashiness, seasonality, groundwater level	Ephemeral changes in drainages and valley floor, development of channel Width/depth ratio increases and/or channel cross-section area increases	Changes in erosion and deposition, development of sedimentary facies within the entrenched minor facies on valley floor
Shale Formation	same	same	100-10,000		Sources of sediments are upland and the channel; silt-clay proportion of sediment classify the streams as suspended load streams	Drainage basin area, drainage network index and determining the amount of discharge which will form the channel dimensions; a progressive increase of channel form downstream direction		Downstream hydraulic geometry, allometric change, cross-section area increases in the downstream direction	Increasing discharge downstream direction will increase the amount of suspended sediment
Friable Sand-stone Formation	Alluviation of the valley floor and possible local cut and fill	Less than or equal to depth of alluvial valley fill	1-10,000	Natural vegetations in species and number of individuals increases	*Adjustment of valley floor facies of local- and headwater-derived sediments, facies of channel fills (some may not be from the valley facies but transported to valley floor from upland) possible formation of muck and a layer of soil of high-cohesive materials*	*Local increase of gradient or changes in hydraulic geometry; lateral change and morphology of streams within local cycle of cut and fill*	*Variations in discharge, groundwater level increases*	*Possible phases of cycles of cut and fill entrenched channels, unentrenched channels; a decrease of channel cross-section area where deposition is dominant along vegetated reaches*	*Valley floor facies, transport of sediment out of the valley during cut phase, downstream relocation of sediment in all phases*
Shale Formation	Alluviation of the valley floor	Same	100-10	Variation and evolution of species of vegetation on the valley floor	Channel adjustment depends on climate complex response to changing discharge conditions	Previous adjustment of the gradient of the valley (equilibrium)	Large discharge conditions	Deposition on the flood plain not in the stream channel; width and depth increases at a uniform rate with discharge	Downstream increase of suspended lead with discharge, alluviation of the floodplain during periods of flooding

*that the system change its operational mode (from aggradation to degradation mode or vice versa)

CHAPTER II

LITERATURE REVIEW AND THEORETICAL FOUNDATION

This chapter seeks to provide a review of past studies of stream channels, which serves to highlight the development of individual 'threads' within the subject (Salisbury, 1972), and to evaluate substantial areas in which closer attention could be profitable. The literature reviewed is concerned with stream channel studies. The contributions of engineering and geologic threads as described by Salisbury (1972) are not evaluated within the study.

The second major topic considered in this chapter is the theoretical foundations of the research. This section will focus on the approaches of examining channel form variations.

Literature Review

The many developments in natural stream channel form studies subsequent to the Leopold and Maddock (1953) paper can be grouped into the following themes (Park, 1976).

1. Downstream channel form variations and the significance of discharge.

2. Spatial variations of channel form and sediment characteristics.

3. Channel form and catastrophic events.

4. Hydraulic geometry exponents.

5. River engineering; the dynamics and morphology of the stream channel.

Human use and man's interest in the land has historically included significant drainage modification. Examples from the Old World include canals and ditches to carry water into and sewage away from ancient cities, and diversion of streams for irrigation. In the United States, stream channelization started about 150 years ago, and thousands of miles of river and stream channel had been modified (Keller, 1976).

Downstream Channel Form Variations and the Significance of Discharge

Downstream channel form variations along individual streams have attracted more attention than spatial variations of channel form characteristics among streams in a region. Different workers have related channel dimensions to stream order and drainage area (Leopold and Miller, 1956). The use of drainage area is preferred to stream ordering schemes because of the lack of ambiguity in identification and measurement, and because sound relationships have been widely established between drainage area and discharge (Brush, 1961; Nixon, 1958; Harvey, 1969). Miller (1958) studied channel properties in the Sangre de Cristo Mountains of New Mexico, and related channel width to drainage area for perennial streams studied by Miller and for ephemeral streams studied in conjunction with Leopold (Leopold and Miller, 1956). The rate of change of width relative to drainage area was shown to be the same in the two areas, although the ephemeral channels were between three and four times wider than the perennial channels for similar drainage areas. Wolman (1967a) has demonstrated that generally larger channel sizes are associated with urban areas in a similar manner, by relating channel width to drainage area. Other illustrations are provided by Harvey (1969), who established relationships between channel size and drainage area for three streams in southern England as a measure of the adjustment of the channel cross-section to discharge. Several studies by Gregory (1974) and Park (1976) provide further illustrations of the approach based on relating form properties to drainage area.

The effects of lithology on permeability of channel beds was isolated in Coffin's (1968) study in Big Sandy Creek in Colorado, which revealed downstream increases in channel width in the upper perennial reaches and downstream decreases in width in lower ephemeral reaches, the latter reflecting channel adjustment to long term downstream decreases in flood discharge.

Other empirical studies have considered spatial variations of channel form relative to distance downstream, which has been suggested as a reasonable surrogate for discharge (e.g., Morisawa, 1969; Patton & Schumm, 1975; and Park, 1978). Close examination of variations in channel width relative to distance downstream enabled Zimmerman, Goodlett, and Comer (1967) to identify two thresholds along the streams studied, related to the influence of vegetation control; with a head-water reach in which vegetation control was strong, a down-stream reach in which the channel size was too large to be strongly influenced by vegetation, and an intermediary reach with quite marked vegetational control of channel width.

Other studies have related the spatial variations of channel form to drainage area and its associations with urbanization (Hammer, 1971, 1972; Wolman, 1967; Gregory & Park, 1976; Morisawa & Laflore, 1979), reservoir construction (Wolman, 1967; Rouse, 1950; Gregory & Park, 1974; Park, 1977), and gully developments associated with drainage schemes (Park, 1977).

On the other hand, either spatial variations, or the change through time of channel form characteristics relative to discharge were illustrated by substantial amounts of empirical field data from stream channels (Leopold, Wolman, Miller, 1964). The original work of Leopold and Maddock (1953) related downstream hydraulic geometry properties to the mean annual discharge, as inferred from gauging station records, but the discharge, or discharges, to which the channel properties are best adjusted has not been finally resolved (Park, 1978). Different workers related indices of channel form to a variety of discharge characteristics, are shown in Table 2.

TABLE 2

DISCHARGE CONDITIONS TO WHICH CHANNEL FORM CHARACTERISTICS HAVE BEEN RELATED *

Discharge	Workers
a. Bankfull discharge	Wolman (1955), Wolman and Leopold (1957), Nixon (1958), Speight (1965), Leopold and Skibitzke (1967), Wilcock (1967a), Harvey (1967), Hickin (1968), Woodyer (1968), Norgan (1969), Chong (1970).
b. Discharge at time of observation	Leopold and Wolman (1956), Thornes (1970).
c. Average discharge	Moore (1968), Brown (1971), Chang and Toebers(1971).
d. Mean annual discharge	Leopold and Maddock (1953), Carlston (1965), Rango (1970), Hedman (1970), DeWalle and Rango (1972), Hedman and Kastner (1972).
e. Mean annual flood	Miller (1958), Brush (1961), Kilpatrick (1961), Schumm (1961a), Kilpatrick and Barnes (1964), Stephenson (1969).
f. Floods of other recurrence intervals	Potter and others (1968), Stephenson (1969), Brown (1971).
g. Flow duration relationships	Leopold and Maddock (1953), Wolman (1955), Wilcock (1967a), Stall and Fok (1968), Knighton (1974).

*After Park (1976)

The problem of the discharges responsible for gross channel forms has thus been accentuated rather than solved in the last two decades (Park, 1978). However, the bankfull discharge is undeniably of some basic significance. The distinction between the condition in which the channel governs the water in the channel, and that in which the water governs the channel, demands identification of some threshold position (Dury, 1975). This threshold is in all probability a function of balance between boundary shear stresses created by the flowing water, and shear resistance and stability of the channel bed and banks, and may occur at or near the bankfull stage (Bathurst, 1979). Wolman and Miller (1960) have shown that flood discharges at relatively low recurrence intervals (1-5 years) are responsible for between fifty and ninety percent of the work done by streams, and they deduce that discharges of this order probably control channel form. The bankfull discharge has been shown to have a recurrence interval in the order of 1 to 2 years in normal streams (Wolman, 1955; Wolman & Miller, 1957), which falls within the range suggested by Wolman and Miller (1960), and a recurrence interval of 1 to 6 years was found by Dury, Hails, and Robbie (1963) to be the recurrence interval of the most probable annual flood (Gumbel, 1958).

A valuable extension of the various attempts to relate channel form to discharge has been the expansion in the number of studies in which discharges in ungauged streams are predicted on the basis of observed channel form properties. Predictions of this type, based upon flow equations (Barnes, 1967) such as those of Manning or Chezy, have been attempted by engineers for several decades (e.g., Dobbie & Wolf, 1953), but geomorphological developments in this field date from the late 1960's.

In 1968, Moore described a means of estimating runoff in ungauged semi-arid areas based on the definition of relationships of runoff to altitude, for hydrologically homogeneous regions. The relations were adjusted on the bases of measurements of channel cross-sections, to ensure sound prediction in small basins. In the same year, Potter, Stovicek and Woo (1968) reported a preliminary method of estimating runoff with a recurrence interval of ten years in small streams based exclusively on channel form properties. Following this lead, a method of estimating the long term mean runoff from ungauged perennial and ephemeral channels, based on channel width and depth, was developed by Hedman (1970) and subsequently implemented by Hedman and Kastner (1972). Channel capacity was used in the estimation of mean flow by Brown (1971), who found the channel parameter to be of greater value than was drainage area. In a pilot study of the use of channel form properties in estimating runoff in 17 small forested basins in the northeast United States, DeWalle and Rango (1972) found that channel width 'explained' 37 percent of the variation in mean annual runoff. Some 78 percent of the variations could be 'explained' by the use of channel widths and drainage area together, which compared favorably with the 83 percent level of 'explanation' offered by mean annual precipitation. Similar approaches to estimating discharge on the basis of channel characteristics have also been outlined by Larson (1972) in Minnesota, and Emmett (1972) has described the application of a similar approach to estimating discharge along the route of a proposed trans-Alaska oil pipeline.

Spatial Variations of Channel Form and Sediment Characteristics

Because hydrology and other environmental conditions such as bedrock characteristics may vary in space over a river basin, the whole river system is dynamically changing spatially. In fact, the geologic factor contributes to the differences that may be found between various longitudinal profiles. The absolute value of slope at any point along a stream depends, in part, on the type of bedrock, but slope decreases in a downstream direction for most types of bedrocks. The decrease in slope with length of stream for common rock types correlates with the increase in discharge in a downstream direction (Brush, 1961; Carlston, 1969), however, it appears that a number of misconceptions have arisen regarding the nature and magnitude of variations in the downstream hydraulic geometry between areas since the Leopold and Maddock (1953) paper, due to the spatial variations in lithology (Brush, 1961).

The relationship of sediment to channel form has been widely studied since the classic flume studies of Gilbert (1914), and since pioneer work on the channel cross-section (Leopold and Maddock, 1953), slope (Hack, 1957), and pattern (Leopold & Wolman, 1957) established the relationship of water and sediment discharge to channel form. In 1960 and 1961, Schumm proposed a relationship between the weighted mean percent silt-clay in the channel perimeter, and channel form as expressed by the width/depth ratio, for stable channels. Also, the gradients of the streams show a general steepening with decreasing silt-clay. Thus, for those small ephemeral streams where the annual rainfall is in the range of 10 to 20 inches both channel shape and gradient are related to (M) where (M) is the weighted percent silt-clay in the sediment composing the channel perimeter. It is probable that a different relation exists between gradient and (M) for larger streams and those in more humid regions. The relation between (M) and width/depth ratio is valid for a large stream in different climatic regions (Schumm, 1961).

Inspired by the work of Schumm, several subsequent workers have examined the relationships between bank material and channel form, in particular as expressed in the width/depth ratio. Thornes (1970), for example, noted a complete lack of significant correlations between width/depth and silt/clay percent for streams in the Xingu Araguaia headwaters of Brazil, although he found that the width of the Casiquaire and Orinoco Rivers was closely related to the bank sediment (Thornes, 1971). Hugo and Hattingh (1971) also failed to establish a relationship between width/depth ratios and silt/clay percent in ephemeral streams in South Africa, and they attributed this to the highly turbulent nature of floodflows along the channel, rapid absorption of floodwaters into the stream beds, and the localized nature of storms and floods in the area. Lack of any systematic downstream variation in the silt/clay percent of bank material on the River Bollin Dean in Cheshire, and the low percentage of the silt/clay fraction in the bed material, prevented Knighton (1974) from recognizing any systematic relationship between downstream channel form changes and sediment factors. He did find, however, that the rate of change of width with discharge at-a-station was largely controlled by bank material composition, and in particular the silt/clay fraction (Park, 1976).

In terms of sediment transport, the significance of both suspended load and bed load have been studied by various workers since Gilbert (1914), and Leopold and Maddock (1953) demonstrated how the shape of the channel cross-sections, and its progressive changes downstream, are conditioned by two characteristics of suspended load transport. These are the increasing concentrations in load relative to increasing discharge at-a-station, and slight decreases in concentration downstream at flows of constant frequencies. However, the type of materials transported or the mode of its transport as a bed load or suspended load appears to be a major factor determining the character of a stream channel, and the classification of channels is based on this relation (Schumm, 1977). Although variations in the proportions of bed load and suspended load are transitional, it is possible to think in terms of a suspended load channel transporting less bed load, a bed load channel transporting a greater bed load, and a mixed load channel transporting bed load greater than suspended load channels and less than bed load channels. The boundaries between these groups are based on the relation between (M) and the percentage of bed load and on the known characteristics of streams having varying amounts of silt and clay in their channels (Schumm, 1963). The type of materials transported as bed load or suspended load have several effects on the stream and the valley. If the stream channel is characterized as a suspended load stream, perhaps the channel has stable banks. In contrary, a bed-load stream is associated with unstable (eroding or depositing) segments along its channel. Channel entrenchment has a detrimental effect on the flat and fertile areas in the valley floor desirable for settlements and economic activities (Cooke & Reeves, 1976).

Park (1976) developed the following argument about the relationships between channel form and sediment in the channel.

> Few theoretical considerations of the effects of sediment on channel form have been reported, and this lack can perhaps be attributed to several factors. Firstly, there have been few attempts to drive hydraulic geometries on theoretical grounds, and the work of Smith (1974) was offered as an alternative to earlier work by Langbein (1964). Secondly, theoretical models of sediment movement must be derived either from the various mathematical bed load functions and formulae--which tend to yield completely different results for a single channel section (Gregory & Walling, 1973)--or from theoretical modeling of the physical processes involved (e.g., Einstein, 1950; Bagnold, 1966). In either case assumptions are involved, many of which may either remain unsubstantiated or are open to debate (Yang, 1972). Thirdly, consideration of the inter-relationships of sediment movement and channel form--which must logically proceed via the influence of roughness--may not yet be

> sufficiently well understood (Richards, 1973) to allow the derivation of a suitable model with either firm statistical or physical foundations. Nonetheless, Wilson (1973) has identified two simple models describing the equilibrium state of channel cross-sections. These are a 'static' model, which is applicable when all bedload grains on the bed are just at the threshold of movement, and a 'dynamic' model which describes channel bed has a sideways component up the channel banks which exactly compensates the tendency of these sediments to move downslope under the influence of gravity. This latter model is used by Wilson to explain why rivers with coarse grained beds tend to be wide and shallow, whereas those with fine beds tend to be deep and narrow.

Channel Form and Catastrophic Events

Considerable interest has also been focused on channel changes caused by extremely large floods, such as those associated with infrequent but devastating hurricanes in the United States (Williams & Guy, 1973). This approach was introduced by Graf (1979) to evaluate the potential utility of catastrophe theory in fluvial geomorphology, as a means of describing change in fluvial systems. Catastrophe theory was designed as a language to replace calculus (a mathematical language used by geomorphologists to describe changes observed in geomorphic variables especially for continuous processes) especially for those situations where discontinuous processes or abrupt changes are common (Graf, 1979).

However, the analogy of Leopold and Maddock (1953) was interpreted to demonstrate that:

> . . . the average river channel system tends to develop in a way to produce an approximate equilibrium between the channel and the water and sediment it must transport.

A few years later Strahler (1957), Hack (1960), and other geomorphologists used the term "dynamic equilibrium" in referring to an open system in a steady state in which the form or characters of the system remains unchanged. In other words, a mean or average elevation of the bed of a river channel is maintained over a period of time because the inflow of sediment to the reach equals the outflow. This balance is determined by the ratio of sediment and water derived from the drainage basin as a function of runoff, vegetation, and soil type (Leopold, Wolman, and Miller, 1964).

The emphasis on stable time periods has led the science away from analysis of change or adjustment, and only recently have geomorphologists turned their attention to those periods between the equilibrium. Schumm (1973), for example, has identified the importance of thresholds and complex responses intrinsic in some fluvial systems. Bull (1975) has shown that some systems do not trend toward a steady state, and Graf (1979a) has analyzed response times during periods of fluvial adjustment. A major symposium addressed the question of change in fluvial systems rather than their stability (Gregory, 1977). This new direction of research, with its emphasis on change rather than stability, has its roots in and is a logical outgrowth of the previous work dealing with equilibrium. The two threads of works are complementary rather than opposed (Graf, 1979).

Hydraulic Geometry Exponents

Hydraulic geometry is the graphical and statistical analysis of the relationships of characteristics of a stream channel to the river hydraulic condition (Chorley, 1972). These include width, depth, slope, discharge, velocity, bed material, roughness, and load. Discharge can be regarded as the only fully independent variable. Bed material can also be regarded as an independent variable, since its properties are controlled by the geology of the drainage basin (Brush, 1961). The remaining hydraulic characteristics interact upon one another in a complex fashion (Morisawa, 1973; Morisawa & Laflure, 1979). The concept of hydraulic geometry as stated by Leopold and Maddock (1953, p. 18) is as follows:

> The channel characteristics of natural rivers are seen to constitute, then, an interdependent system which can be described by a series of graphs having simple geometric form. The geometric form of the graphs describing these interactions suggests the term 'hydraulic geometry.'

Although some variations in the exponent and constant values of the hydraulic geometry relationships for streams in the midwest United States were apparent in the original analysis by Leopold and Maddock (1953), mean values of the exponents were calculated, and these have subsequently been accepted as a 'standard' for midwest streams (Park, 1978). When downstream variations in the exponents were examined by Park (1976) with the aid of tri-axial distribution graphs, several points emerge. Humid temperate streams tend to have medium to high width exponents (0.4 to 0.8), medium depth exponents (0.2 to 0.6) and low velocity exponents (-0.5 to 0.3). These contrast with the tropical streams. A distinction can be drawn between most of the perennial channels which have exponents similar to those in humid temperate streams, and the few ephemeral streams available at that time, which have low width exponents (below 0.3) and relatively high velocity exponents (0.4 to 0.6). The two anomalous perennial sites in this analysis by Park (1976), which have extremely high velocity and low width exponents, are the Pojoaque and Santa Cruz rivers in New Mexico described by Miller (1958). Contrasts in the hydraulic geometry exponents between the major climatic environments can thus be identified and these are summarized in Table 3 (after Park, 1976).

TABLE 3

SUMMARY OF DOWNSTREAM HYDRAULIC GEOMETRY EXPONENTS BETWEEN MAJOR CLIMATIC ENVIRONMENTS

Climatic Environment	Exponents: Width	Depth	Velocity
a. Humid temperate	medium-low	medium	low
b. Semi-arid perennial	medium-high	medium	low
c. Semi-arid ephemeral	low	medium	high
d. Tropical	medium	medium	medium
e. Pro-glacial	Insufficient evidence		

This does not point directly to the conclusion that climate directly controls the downstream hydraulic geometry exponents, but rather it suggests that the exponents are affected by major climatic differences as expressed through differences in water and sediment yield as well as through gross topographic differences, which were taken into consideration by Morisawa and Laflure (1979).

Park (1978) pointed out that there is a considerable range of exponents in the at-a-station and downstream geometries, and so this casts doubts on the use of mean values samples of exponents to characterize the hydraulic geometries of particular areas (e.g., Leopold & Maddock, 1953). In most of the literature cited in this chapter no attempt was made to explain the causes of variations in the hydraulic geometries as reflected in the exponents and the constants.

River Engineering:
The Dynamics and Morphology of the River

As in a drainage basin, so along the river variations of valley slope, sediment load, and discharge cause local reaches to be unstable. If, for a river, a relation between stream power and sinuosity or between valley slope and sinuosity can be established, then a plan for stabilization may include an increase of sinuosity as a means of decreasing stream power. Such an approach may be a valid substitute for the bank stabilization and realignment techniques currently in use, but it would require the construction of dikes to begin development of a sinuous pattern (Schumm, 1977, p. 327).

For natural stream channels, it is easier to find examples of meandering or braided channels than to find long straight reaches. Even where the channel is straight it is usual for the thalweg, or line of maximum depth, to wander back and forth from near one bank to the other. Opposite the point of greatest depth there is usually a bar or an accumulation of sediment along the bank, and these bars tend to alternate from one side of the channel to the other. In an idealized sense, flow and depositional pattern seen in the plan view is similar in straight and in meandering patterns (Leopold et al., 1964, p. 281). This condition is described as the morphologic recovery of a modified channel, defined as a return to a more natural distribution of bed forms such as pools and riffles. This morphologic recovery is likely to be more rapid than biologic recovery, because the diversity of flow provided by pools and riffles is necessary for maximum biologic productivity. Although incipient pools may begin to form soon after channelization (Morisawa, 1974; Keller, 1975), the time necessary for the formation of a morphologically stable series of pools and riffles is not known, in fact, if the stream is severely disturbed and periodically remodified, they may never form.

> One of the requirements for pools and riffles in nonmeandering streams appears to be some degree of heterogeneity of bed material size. Channels that carry uniform sand or uniform silt appear to have little tendency to form pools and riffles. In part this might be attributed in sand channels to the concomitant high width-to-depth ratio associated with relatively noncohesive bank materials,

> and this ratio in turn tends to promote a braided pattern. Pools and riffles are less well developed in braided than in straight but nonbraided channels. Braided or anastomosing channels are often but not always associated with sandy or friable bank materials. Also, vegetation has similar effects; a change from nonbraided to braided character is sometimes associated with a change from dense vegetation along the channel banks to sparse or nonvegetation. Whether these coincident changes are causally related cannot usually be ascertained, although the coincidence is suggestive. (Leopold, et al., 1964, p. 282)

Channelization is a controversial issue, and, although more environmental aspects are being considered today than ever before, the pressure to modify channels will increase in the future. Therefore, it is important to discover ways to modify channels that will minimize environmental degradation and maximize the overall return to man. Several aspects of fluvial geomorphology and hydrology that are now or may eventually be helpful in designing more natural channels are: (1) the concept that streams are open systems in which channel form and process evolve in harmony, (2) utilization of the convergent-divergent criterion, (3) identification of geomorphic thresholds for stream channel, and (4) recognition of the complex relationships between deposition, erosion, and sediment concenteration. In general, increased slope of the straightened channel will normally cause incision and rejuvenation upstream, thereby producing a high sediment load that will probably enlarge and aggrade the aligned channel and cause flood plain destruction. Again, the channels that will benefit most from partial or full channelization probably can be identified by a careful study of channel morphology.

Methodological Aspect of the Problem

The literature review revealed that investigators sought to explain areas of erosion and deposition among segments of a stream channel and among streams in the region, as a consequence of 'human' or 'climatic' impact. But very few have attempted to explain degradation and aggradation along streams according to environmental variations (i.e., differential geologic formations, unusual flood or rainfall events, differential flows through the drainage net, and inherent disadjustments in stream profiles). Therefore, the basic method that will be used in the study is to adopt different approaches for different lithology to predict changes in channel form. Firstly, adopting an approach in the study to use the arroyo-development process as an indication of channel form change as the fluvial system adjusts to internal or external stimuli. Secondly, adopting an approach in the study to use the relationships between cross-sectional from variables and total stream length above a particular cross-section as a means of characterizing the variations of channel form along individual streams. Thirdly, adopting an approach in the study to use the relationship between cross-sectional form variables and sediment characteristics as a means of describing the type of geomorphic thresholds in the system, whether they are thresholds controlled by external variables or are thresholds controlled by internal variables.

Approach One

Approach one is to test catastrophe theory in the system using data from entrenched stream segments. The entrenched stream segments are best considered as candidates for the application of catastrophe theory because the process includes abrupt changes as thresholds are crossed. The system controls (independent variables) are readily identified as (a) the force of flowing water operating against (b) the resistance of earth materials. The behavioral variables (dependent variables) reflecting this operation is the area of the channel cross-section at a given point in the stream network (Graf, 1979b). However, the process of arroyo development is a longstanding question in geomorphic research (Cooke & Reeves, 1976). That is, the same forms have resulted from a wide variety of causes, some of which are detectable after their operation and some of which are not. Emphasis in arroyo-related research is now turning to the process of channel changes as the fluvial system adjusts to internal (Schumm, 1977) or external (Cooke & Reeves, 1976) stimuli.

Approach Two

Approach two will test the variations in channel size properties relative to drainage area or total stream length above a particular cross-section, which will be employed both as a measure of spatial location, and as a surrogate for discharge. The hydraulic geometry approach is important to relate the characteristics of the total area or the total stream length above sampling points to channel morphology.

Approach Three

Approach three is to examine the relationship between channel morphology and lithology. That is, the characteristics of superficial deposits should be numerically described. In essence, alluvial stream channels have been classified on the basis of one of the two independent variables influencing stream morphology. Discharge cannot be used as a basis for classification because it controls mainly the size of the stream channel. The type of materials transported by the stream is a major factor determining the character of stream channel. Absolute size of the sediment load may be less important than the manner in which it moves through the channel. Channel stability depends on the balance or lack of balance between sediment load and transportability and three classes of channel result: stable, eroding, and depositing (Schumm, 1977).

Theoretical Foundation of the Problem
Concepts of Channel Form Change

Attention could be subsequently focused on three concepts which are: (1) catastrophe theory, (2) the concept of allometric change, and (3) sediment properties as a geomorphic factor determining stream channel shape and differences in mechanics of erosion and deposition between areas (Table 4). By applying these approaches it will be possible to examine areas of erosion and deposition along stream channels, and to

TABLE 4

CONCEPTS OF CHANNEL FORM CHANGE

Key Concepts	Objective	Method	Source	Units	Function	Variables
Catastrophe theory or catastrophic change applied to streams on the sandstone formations	Tractive force	DuBoys	Graf (1979a)	dynes	$\tau_o = (\gamma_f)(d)(\theta_c)$	τ_o=tractive force γ_f=density of fluid d=depth of flow θ_c=slope of the channel
	Depth of flow	modified manning equation	Graf (1979a)	feet	$d = \left[\dfrac{Q_{10}}{\frac{1.49}{n}(0.55)(W_m)(\theta_c)^{0.5}}\right]^{0.6}$	d=depth of flow Q_{10}=10-years discharge n=roughness parameter w_m=mean width of channel θ_c=slope of the channel
The concept of allometric change applied to streams on the shale formations	Relationship between channel geometry and network morphometry	Gregory	Gregory (1977)	sq. feet	$C = a(\Sigma L)^6$	C=channel capacity (width x depth) ΣL=Total stream length above a particular cross-section
Sediment properties as a geomorphic threshold applied to streams on the sandstone and shale formations	Weighted mean percentage silt-clay for each cross-section	Schumm	Schumm (1960)	Weighted mean percent	$M = \dfrac{Scxw+Sbx2d}{w+2d}$	M=weighted mean percentage silt-clay S_c=percent silt-clay in channel alluvium S_b=percent silt-clay in bank alluvium d=channel depth w=channel width
	Relationship between channel shape and weighted mean percentage silt-clay	Schumm	Schumm (1961)	ratio	$F = a(M)^b$	F=width-depth ratio M=weighted mean percentage silt-clay a+b:are constants

establish the magnitude of spatial variations in various channel cross-section form characteristics along individual streams and between streams among different geological formations. Throughout the study it may be possible to confirm that geology plays an important role in determining the shape of the channel cross-section. Then, it is possible to synthesize the findings from the various approaches which will be applied into a general model.

Concept One:
Catastrophe Theory -

In fluvial geomorphic applications, the catastrophe concept is readily acceptable and has been explicitly employed by researchers for nearly two decades (Chorley, 1962; Chorley & Kennedy, 1972). Control variables (independent variables) could be defined in a number of ways, but the most basic are force and resistance. Possible behavior (dependent variables) variables include channel characteristics such as width, depth, slope, and roughness (Graf, 1979b).

Research into questions of arroyo processes has only begun to explore the nature of changes that take place within arroyos and their drainage basins. If processes are measured by changes in observed variables in a system (Thorness & Brunsden, 1977), basic quantitative research into arroyo processes is limited to a few works, including those by Leopold and Miller (1956), Schumm and Hadley (1957), Melton (1965), Cooke and Reeves (1976), and Malde and Scott (1977). La March (1966), Womack and Schumm (1977), and Graf (1979a) provided quantitative data on rates of process. The general suggestion by Horton (1945) and Strahler (1952) that geomorphologists concentrate on the interplay between force and resistance has been largely overlooked by researchers studying arroyos.

The idea behind arroyo cutting (channel entrenchment) is that the force in the channel exceeds the resistance offered by vegetation cover. For example, a 10-year flood discharge may have a force that the resistance offered by vegetation cover is unable to withstand. On the other hand, biomass of vegetation on the valley floor exerts significant control on the trenching process (Graf, 1979a).

Data of discharge at various return intervals are needed to determine flow depths for tractive force calculation. The method used by Graf (1979a) would be appropriate for determining discharge by indirect ways when required detailed meterological data are not available (where basin areas determined for each measurement site are small).

If the discharge for the channel is known, a modified form of the Manning equation provides depth of flow:

$$d = \left(\frac{Q_{10}}{\frac{1.49}{n}\,(0.55)\,(W_m)\,(\theta_c)^{0.5}}\right)^{0.6} \quad \ldots\ldots\ldots\ldots \quad (1)$$

where n = Manning roughness coefficient as observed in the field and evaluated according to Barnes (1967), W_m = mean channel width, measured in feet to maintain consistency with the values for Q(cfs), and θ_c = channel gradient.

In many applications of variations of Manning Equation, depth is directly substituted for hydraulic radius if the channels are wide and shallow.

The general form of the tractive force relationship in an open channel is the DuBoys equation (Rouse, 1950, p. 194):

$$\tau_o = (\gamma_f)(d)(\theta_c) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad (2)$$

where τ_o = tractive force in dynes, γ_f = specific weight of the fluid, and d = depth of flow as derived above, converted to Cm for compatibility with the measure of the force in dynes (see Bogardi, 1974 for discussion of the strengths and weaknesses of the function).

Surficial materials and vegetation on valley floors resist the force generated by flood flows. The study area selected for application should not seriously violate the assumption that the materials of the valley fills are homogeneous and their cohesion is considered constant. Vegetation on valley floor, however, could be found highly variable because it is always subject to manipulation by human activities. Therefore, biomass on the valley floor (B_v) is a useful measure of resistance at measurement sites.

The comparison between τ_o and B_v may demonstrate why channel entrenchment is sensitive to vegetation changes of both broad and narrow extents. Reduction of vegetation throughout the basin results in increases in forces, which might initiate entrenching.

Therefore, the amount of excavation that occurs depends on the amount of force available, the vegetative resistance, the nature of the materials, and the amount of time involved. However, the time element of channel entrenching in the study area depends on the historical data available.

William Graf (1979b) provided the language of catastrophic change and introduced what is called catastrophe theory as a model for change in fluvial systems. Controversies around catastrophe theory were, also, created. That is not a product of its development, but rather results from attempts at application. The mathematical language used by geomorphologists to describe changes observed in geomorphic variables has been based on calculus, a body of mathematics that is extremely flexible and adaptable, especially for continuous processes and those that have smooth transitions from one state to another. Geomorphic research concerned with equilibrium has made effective use of calculus-based mathematics, which have yet to be fully exploited for this purpose. The flexibility of calculus is limited, however, when discontinuous operations or abrupt transitions are considered (Zeeman, 1976). Catastrophe theory was designed as a language to replace calculus, especially for those situations where discontinuous processes or abrupt changes are common. In the light of the characteristics of the theory, it could be a useful representation of change in those geomorphic systems definable by two control factors (force and resistance) and behavior dimension.

Concept Two:
The Concept of Allometric Change -

The concept of allometric change, developed and widely applied in biology, is an approach used to examine the nature of static allometric changes in stream channel size relative to drainage basin area. In other words, it suggests a tendency for streams to reach a steady-state allometric situation when channel capacity increases at the same rate as drainage area (Bull, 1975; Park, 1978).

The removal of materials from the valley floor results in the creation and modification of the channel form. The width and depth of the channel are related to the amount of materials excavated, but these two basic dimensions change at different rates. Width is controlled by mass movement processes, especially soil fall and slumping from the channel banks, while depth is primarily controlled by fluvial processes on the entrenched channel floor. If rates of change for the dimensions are related to a third variable (amount of material excavated, discharge or tractive force), then the dimensions are related to each other by a power function (Graf, 1979a):

$$W = a(d)^b \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad (3)$$

where W = width, d = depth, a = constant, and b = the rate of change. If the rate of change for both variables is the same, b = 1.0, and there is an isometric change characterizing the form. If the rates of change are different, a more likely situation since two different processes are at work, $b \neq 1.0$, and allometry prevails. See Woldenberg (1968) and Bull (1975) for discussions of allometry in geomorphology. Although Woldenberg (1968) and Bull (1975) do not explain why changes occur in the entrenched channel forms, the allometric concept provides a useful descriptive method for the different rates of change of channel width and depth.

Woldenberg (1972) has pointed out that the concept of hydraulic geometry as advocated and illustrated by Leopold and Maddock (1953), is one of the more easily visualized examples of allometry in geomorphology. Two recent evaluations of the nature and magnitude of variations in hydraulic geometry exponents, based on analysis of large bodies of data from the literature, have revealed substantial differences in the exponents when the b, f, and m exponents are considered simultaneously by means of triaxial scatter plots (Park, 1976). These suggest that the classic approach of characterizing variations in flow width, depth, and velocity relative to discharge both at individual sites (at-a-station hydraulic geometry) and in the downstream direction (downstream hydraulic geometry) may not be really suitable for identifying regional spatial patterns of channel morphology.

The independent variable against which variations in channel form variables can be related should logically provide both a surrogate for discharge and a measure of spatial locations (although these are highly interrelated). There would appear to be at least four possibilities (Park, 1978):

(1) Stream Order: A basic measure of relative location within stream systems, which has shown to relate to stream discharge (Leopold and Miller, 1956; Leopold, Wolman, and Miller, 1964).

(2) Distance Downstream: A simple measure of relative location, which is shown to be related to discharge (Leopold et al., 1964) and less ambiguous to determine and interpret there order.

(3) Total Stream Length Upstream: This measure, which circumvents the problems of ignoring network structure upstream, has been used by Gregory (1977).

(4) Drainage Area: A standard measure of relative location, which is known in many studies to be related to discharge (Leopold et al., 1964).

The original study which related the hydraulic geometry of stream channel to drainage basin network was Leopold and Miller's 1956 paper. They stated that the quantitative description of drainage nets developed by Horton (1945) related stream order to the number, average length, and average slope of streams in a drainage basin. Their purpose was to show that this useful tool may be extended to include the hydraulic as well as drainage-net characteristics. The interrelationship between the drainage network and hydraulic geometry was established via the concept of proportionality in the following manner. Horton (1945) demonstrated that the relationship between stream order (O) and drainage area (Ad) can be defined using the following expression:

$$O = \log Ad \quad \text{(4)}$$

Furthermore, it is well established that Q (discharge) is related to drainage area by the simple power function:

$$Q = Ad^{K} \quad \text{(5)}$$

Therefore, it would follow that discharge is related to stream order in the form:

$$O \propto \log Q \quad \text{(6)}$$

Using the hydraulic geometry equations of Leopold and Maddock (1953):

$$W \propto Q^{b} \quad \text{(7)}$$

$$d \propto Q^{f} \quad \text{(8)}$$

$$V \propto Q^{m} \quad \text{(9)}$$

$$L \propto Q^{j} \quad \text{(10)}$$

$$S \propto Q^{z} \quad \text{(11)}$$

$$n \propto Q^{y} \quad \text{(12)}$$

where Q is the discharge, W is the channel width, d is the channel depth, V is mean velocity, L is the suspended load, S is the local channel slope, n is the channel roughness, and b, f, m, j, z, and y are exponents for each equation. Leopold and Miller (1956) suggested that the following multivariate relationship should "definitely" exist in the form:

$$O \ \alpha \ \log (W, d, V, L, S, n) \qquad (13)$$

The empirical confirmation of Equation (6) was based on the following relationships:

$$O = K \log W \qquad (14)$$

$$O = K \log S \qquad (15)$$

Channel width was then regressed against channel slope, and the equation was expressed as follows:

$$S = 0.12\ W^{-0.5} \qquad (16)$$

Equation (16) along with the well-accepted nature of Equation (5) is the foundation by which Leopold and Miller (1956) integrated the geomorphic and physiographic characteristics. Also, they indicated that ephemeral streams in New Mexico are characterized by a uniform downstream increase of width, depth, and velocity, with stream order and also with drainage basin size. Increasing size of a drainage basin is combined by a downstream increase of discharge. Furthermore, the interrelations among all of these factors, both hydraulic and physiographic, may be expressed in simple terms, either as exponential or power functions. The method of combining the hydraulic variable with factors measured on a map or obtained in the field from a dry stream bed allows a simple means of obtaining interrelations which cannot be measured directly.

To advance this approach one must define the independent variable which will be regressed against channel form variables. That provides a logical surrogate for discharge and a measure of spatial variation. It is desirable to have a volumetric index of the stream network (Gregory, 1977). The reasons of selecting that index are its applicability on drainage basin impounded by flood-water structure. Also, the index is applicable to drainage basins characterized by high differential flows through the drainage net, as well as drainage basins characterized by an increase in channel size relative to drainage area.

To obtain an index of network volume, it is necessary to relate the length of the network to the capacity of the channels to which the network is tributary. Within any one basin, measurements of the channel cross-section can be made at a number of locations and related to measurements of length of stream channel above the cross-section. These measurements can then serve as a basis for an index of stream network volume. Several workers have related stream channel capacity to drainage area (Harvey, 1969; Gregory et al., 1974) or stream channel width to drainage area (Day, 1972), and stream capacity to stream network volume (Gregory, 1977). A linear regression relating channel capacity (C) as the dependent variable to the total stream length (ΣL) above a particular channel cross section provides a means of relating channel geometry to network morphometry in the form:

$$C = a\ (\Sigma L)^b \quad \text{. (17)}$$

As it represents a combination of channel geometry and network morphometry, an index of channel morphometry may be developed by integrating the expression between the limits of the stream network (Gregory, 1977). Thus:

$$\int C = \int_{n_1}^{n_2} a(\Sigma L)^b \cdot d(\Sigma L) \quad \text{. . . . . (18)}$$

where n_1 is the upper limit and n_2 is the lower limit of the channel network considered. The integrated expression can then provide an index of the volume of the channel network between limits n_1 and n_2. Because water flow in channels is at least ten times more rapid than water flow by overland or subsurface routes, the volume of the stream network calculated in this way will index the potential of the basin to transmit water rapidly. The value of channel morphometry calculated for several basins could be used in relation to water and sediment production from the basins, it could be used as an index to compare the fluvial morphology of the networks, and it could also be employed to indicate the magnitude of changes which have taken place. Therefore, the choice of the allometric model is based on whether the model can be proved to exist in the field, and whether it provides flexibility of the channel form change. In any case, relation of channel form to total stream length in the form $y = ax^b$ can be defined, which will vary depending on the spatial scale of the streams selected for the study.

Concept Three:
Sediment Properties as a Geomorphic Factor Determining Stream Channel Shape -

The width/depth ratio of a channel defined as stable (because there is no progressive channel adjustment) is related to percentage of silt/clay in the perimeter of the channel (according to Figure 5.2 in Schumm, 1977, p. 109). Also, the gradient of stream channel is directly related to both bed material load and grain size and inversely related to water discharge. Hack (1957) found the relation between channel gradient and median grain size (D_{50}) using the drainage area as an index of discharge. Brush (1961) also concluded that downstream decreases of stream gradient depend upon the resistance of the rock being eroded and supplied to the channel. The gradient of a stream, therefore, is strongly influenced by the hydrology and geology of the drainage basin. The latter provides sediment of various quantities and sizes.

Complete description of the characteristics of the drainage basin must include reference to the rocks and sediment beneath the basin. This is necessary because the type of rock underlying the basin will determine the nature and extent of groundwater storage and also the type of material which is available for erosion and transport within the drainage basin. Geologists have provided numerous techniques for describing the character of rocks and similarly pedology has devised methods of soil description. Both sciences have produced maps at a variety of scales. Such maps are necessarily governed in their production by the needs of the disciplines from which they arise and so the basis of classification of rock type, soil type, and superficial deposits on published maps are frequently based primarily upon genesis, which is not necessarily the most relevant and appropriate basis for the fluvial geomorphologist. Ancillary techniques, therefore, have to be utilized to express the character of rocks and sediment relevant to the form and function of the drainage basin. Such techniques have been incompletely explored by geomorphologists compared with the techniques denoting topographic characteristics of drainage basins or describing the fluvial sediments within the basin. These techniques are going to be used in the research to describe the characteristics of rock, of superficial deposits and soils, and of fluvial sediments.

Although in most studies of rivers the median grain size (D_{50}) has been used as a descriptor of sediment type, some studies revealed that there is a great change in sediment character within the range 0.05 to 0.1 mm. Any grain size between 0.05 mm and 0.1 mm could be used as the boundary between silt-clay and sand. However, it is convenient to use either the 200-mesh (0.074 mm) or 230-mesh (0.0625 mm) sieves (Schumm, 1960). Schumm used the weighted mean percentage silt-clay for each cross section (M) as a parameter to express the character of the sediment forming the perimeter of the channel where:

$$M = \frac{S_c \times w + S_b \times 2d}{w + 2d} \quad \ldots\ldots\ldots\ldots\ldots \quad (19)$$

in which:

S_c = percent silt-clay in channel bed alluvium

S_b = percent silt-clay in channel banks alluvium

d = channel depth

w = channel width.

As the percentage of silt-clay in channel bed and banks increases, the shape of stream channels expressed as width-depth ratio (F) varies according to the equation:

$$F = 255\ (M)^{-1.08} \quad \ldots\ldots\ldots\ldots\ldots \quad (20)$$

Also, the gradients of the streams show a general steepening with decreasing silt-clay. Thus, for those small ephemeral streams where the annual rainfall is within the range 10-20 inches, both channel shape and channel slope are related to (M) (Schumm, 1961). It is probable that a different relation exists between gradient and (M) for larger streams and more in humid regions. The relation between (M) and width/depth ratio is valid for a large range of streams in different climatic regions (Schumm, 1960).

The type of materials transported or the mode of its transport as a bed load or suspended load appears to be a major factor determining the character of a stream channel, and the classification of channels is based on this (Schumm, 1977). Streams having high content of silt-clay are characterized by low width/depth ratio and the mode of sediment transport is suspended load. On the other hand, streams having a log silt-clay content are better described as bed load streams where transportation depends on the balance between load and stream competence. Absolute size of sediment, therefore, may be less important than the manner in which it moves through the channel. Furthermore, channel stability depends on the balance or lack of balance between sediment load and transportability and three classes of channels result; stable, eroding, and depositing (Schumm, 1963).

Lane (1955) depicted the relationships between water discharge, Q_w, channel slope, S, sediment size, D, and bed-load discharge, Q_s, as:

$$Q_s\ D\ \alpha\ Q_w\ S \qquad \ldots\ldots\ldots\ldots\ (21)$$

Accordingly, if water discharge is abruptly increased relative to bed-load sediment discharge and if slope does not immediately change, a stream will react either by entraining more of the bed sediment to meet the increased capacity, or by entraining larger sizes of bed materials to meet the increased competence. Usually both adjustments occur.

To include properties of channel geometry, Schumm (1969, 1971) extended Lane's qualitative relationship to the form of:

$$Q_w \; \alpha \; \frac{W d L}{S} \qquad \text{. . . . . (22)}$$

where W is channel width, d is channel depth, and L is meander wave length. This relationship indicates that the channel should respond to the increased discharge by eroding its channel and thus increasing its width and depth. This relationship was employed in a study by Shepherd (1979) of Sandy Creek watershed in Central Texas. Because the channel of Sandy Creek was bed rock and alluvial, the width and depth increases occurred not by erosion but by the increased discharges being conveyed in larger cross-section areas of the valley. Furthermore, the constricting effect of the bed rock bed and banks of the channel of Sandy Creek induced incision such that the increase of depth was greater than that of width (Shepherd and Schumm, 1974).

Schumm (1969) depicted the relationships between the increase in bed-load and channel variables as

$$Q_s \; \alpha \; \frac{W L S}{d P} \qquad \text{. . . . . (23)}$$

for alluvial streams, where P is sinuosity and the other variables are the same as in Equations 21 and 22. The combined relationships are (Schumm, 1969, p. 261):

$$Q_w^{\,+} \, Q_s^{\,+} \; \alpha \; \frac{W^+ L^+ (W/d)^+}{P^-} \, S^{\pm} \, d^{\pm} \qquad \text{. . . . . . . (24)}$$

where + and - indicate the direction of change. Schumm suggested that although S and d could either decrease or increase, S will usually increase and d will usually remain constant or decrease in alluvial channels (see also Cook and Warren, 1973).

Consequently, Schumm's (1963) classification of alluvial river channels was through distinctions of width-depth ratio, gradient, sinuosity, and relative amounts of sediment transported as bed load and suspended load. According to specific ranges of these factors Schumm distinguished suspended-load, mixed-load, and bed-load alluvial channels, and characterized their dominant processes of erosion and deposition. Two important factors limit the concise applicability of Schumm's classification in such a study area. First, the availability of precise data on the relative amounts of sediment transported as bed load or suspended load. Second, the availability of alluvial stream channels in the study area. Only the type of the channel bed sediment (sand bed channel or silt-clay bed channel) is a major independent variable because it is readily measured in the field. Therefore, the classification is (bed load stream or suspended load stream based on the percentage of silt-clay in the sediment in the channel) analogous to that of Schumm (1963), and his terminology will be employed.

CHAPTER III

THE STUDY AREA

General Deliniation

The area considered in this study is in the Osage plains of the Central Low Land Province and is centered in west-central Oklahoma. It includes all the sections from T. 7 N. through T.11 N., R.6 W. through R.10 W. in Grady and Caddo counties, Oklahoma. The location of the streams selected for this study within the area are shown in Figure 2 and will be referred to herein by their respective names.

Rolling plains developed by erosion of shales and siltstones characterize the eastern part of Grady County. The western part of Grady County and most of Caddo County, north of the Washita River, is cut by deep channels eroded in sandy shales and sandstones (Fenneman, 1938, pp. 455, 605-630). The upland prairie of the area between the Canadian River and the Washita River in Caddo County is essentially a terrace-leveled surface on the underlying Rush Springs Sandstone. On this surface, stream erosion has incised a dentritic pattern of steep walled canyons, badlands, and buttes. Near the contact of the Rush Springs, sandstone and the siltstone and shales of the underlying Marlow formation, the canyon walls lose their steepness and grade into gentle slopes.

Tributaries to two of the major rivers in Oklahoma, the Washita and Canadian, drain the area. Buggy Creek, draining the northern part of the area, is the only tributary of the Canadian River present here. The other principal streams in the area are Sugar Creek, Spring Creek, and West Ionine Creek. These streams head in the outcrop area of the Rush Springs Sandstone and are spring fed. Other major tributaries of the Washita River including Salt Creek head in the outcrop of the Dog Creek Shale.

Some streams of the central segment of the Washita River and the Canadian River were selected for the study. Devil's Canyon and Kickapoo Creeks drain into Sugar Creek which is a tributary of the Washita River. Buggy Creek heads in the southwestern township of Canadian County, Oklahoma, and flows through the northwestern part of Grady County to empty into the Canadian River east of Minco, Oklahoma.

> Although the area if geographically centered between the two major reivers, the larger amount of drainage goes to the Washita River due to a 160 foot difference in grade level between them (Norris, 1951). Within the Anadarko-Alex reach of the Washita River, West Fork of Salt Creek and Salt Creek are drained southward into the Washita River which is a tributary of the Red River (Figure 2).
>
> Many streams are incised in a broad valley floor of a former drainage system. Rapid headward

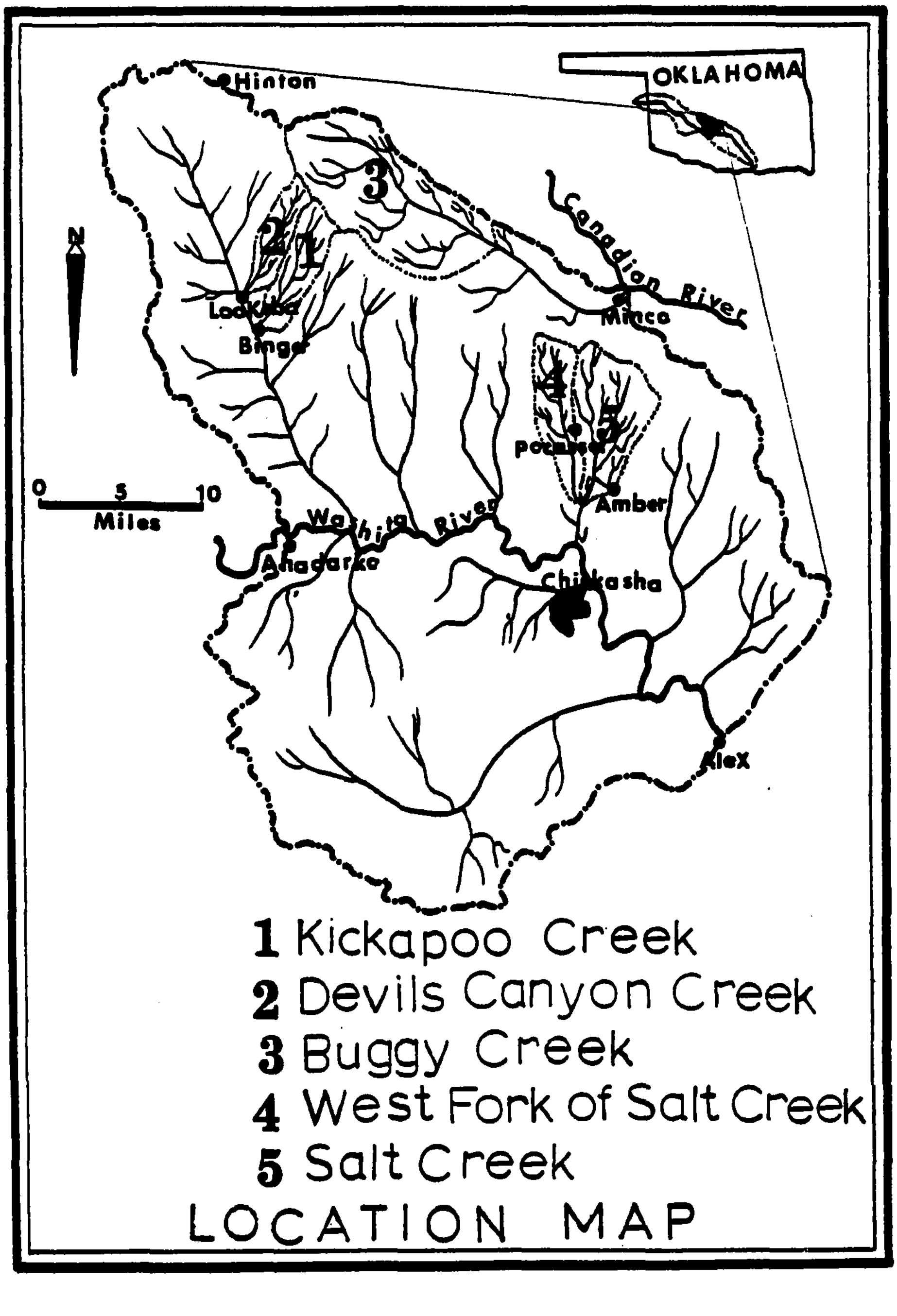

Hinton
OKLAHOMA
N
Canadian River
Amber
0 5 10
Miles
1 Kickapoo Creek
2 Devils Canyon Creek
3 Buggy Creek
4 West Fork of Salt Creek
5 Salt Creek
LOCATION MAP

Figure 2

erosion in canyon fill sediments has re-excavated several buried canyons. All stages of canyon re-excavation are present in this area (Norris, 1951). Knickpoints and notched box head canyons are indicative of multi-cycle erosion (Howery, 1960, p. 11).

The dendritic pattern of the streams affords no consistent relationship to structure. Comparison of the geologic map and the structural contour map indicates that Sugar Creek, for example, is paraclined as well as diaclinal. Buggy Creek has some relation to structure but no definite pattern exists for the entire stream system. This consequent stream pattern is probably more affected by regional surface slope than by structure of the underlying strata which dip only five to fifty feet per mile. (Davis, 1955 and Howery, 1960, p. 11 & 12)

In the area the wind plays an important role in the type of vegetation that protects the soil. Its desiccating effects limit the amount of moisture the vegetation receives, and its velocity affects the type of trees that can exist there. A marked contrast in vegetation is observed at the contact of the Rush Springs Sandstone and the Marlow Formation. Dense growth of scrub oak and juniper flourish on the Rush Springs as a result of their natural "preference" for sandy soils. On the Marlow Formation native grasses are predominant. Along the valley floors the streams are bordered by a fringe of willow, sycamore, cottonwood, and oak trees as well as by underbrush. Inside the canyons a relict flora of sugar maple and other large trees exotic to this area exist. This occurrence is probably enhanced by the protection afforded by the canyon walls and a water supply afforded by the stratum springs, typical of Devil's Canyon Creek.

Water Balance

Northern Caddo County has a climate subject to moderate to extreme climatic changes. The annual mean temperature is 62°F and the annual average precipitation is 29.6 inches, far below the average evaporation per year (See State Climatologists, 1957; Thornthwaite, 1942). There can be extended drought in this area, as exemplified by the "Dust Bowl" period of the 1930's, as well as by thunderstorms during which six to seven inches of rain fall in one twenty-four hour period. The upland prairie is continually subject to winds averaging more than 14 miles per hour with occasional gusts capable of uprooting full-grown trees. However, in the narrow, steep-walled canyons, a relict flora of tall trees exists, well watered and protected from the drier, more windy environment above (Howery, 1960, p. 13).

In Grady County north of the Washita River the area has a moist-subhumid climate (State Climatologists, 1957; Thornthwaite, 1942). In winter the temperatures generally are moderate with occasional short periods of severe cold and relatively little snow. In summer temperatures often are uncomfortably hot during the day, but usually the nights are cool. The distribution of precipitation is generally such that it exceeds the losses by evaporation and plant needs, resulting in a moist subsoil and acceretion to ground water. Farming without irrigation is practical in most years, but several years of average or above average rain may be floowed by several dry years during which irrigation would be necessary for optimum crop growth.

Tanaka et al. (1963, p. 34) summarize the natural routes of groundwater discharge as follows:

> Discharge of groundwater by transpiration and evaporation from the zone of saturation is negligible over most of the area because the average depth to the water table in the upland area is more than 40 feet. Only where the water table is close to the land surface, such as along the major drainage systems, does discharge of groundwater through evapotranspiration become significant.

The losses due to evaporation and transpiration are large, however, for that part of precipitation that does not infiltrate to the water table. The magnitude of this loss is illustrated by the study of the Cobb Creek basin (Clark, 1956). The average annual precipitation over the basin is 28.10 inches, which is most of the water that enters the basin. If the yearly recharge to the aquifer is 2.8 inches and surface runoff of Cobb Creek is 2.1 inches, evapotranspiration must account for about 23 inches, or 80 percent of the total water that enters the basin. This figure for evapotranspiration probably would apply over that part of the study are in Caddo County (i.e., Kickapoo Creek, Devil's Canyon Creek, and the upstream part of Buggy Creek).

> The rate of groundwater movement varies with the gradient of the water table if transmissibility and storage in the aquifer are assumed to remain about constant. Water levels measured over a period of years in Caddo County (Tanaka, 1963) indicate that groundwater moves toward the larger streams at a fairly constant rate. No significant yearly change occurs in the groundwater gradient except near the major drainage areas, where groundwater discharge is affected mainly by evaporation and transpiration by plants during the summer months. (Tanaka et al., 1963, p. 35)

Tanaka, et al. (1963, p. 34-35) reported the monthly groundwater discharge of Cobb Creek computed from base-flow rating curve, 1953-1956 as follows:

> Groundwater discharge was computed by Meinzer and Stearns (1929) in their study of the Pomperaug Basin, Connecticut, by relating stream base flow to the average water level in several wells. A similar approach is used to compute groundwater discharge by relating average water levels in wells near Cobb Creek to its base flow.
>
> The depths of water in seven wells near Cobb Creek were averaged. Groundwater

> stages were selected only during the season when evaporation and transpiration losses were low (November through February) so that the effects of evaportranspiration on stream base flow would be at the minimum, and the selection was further restricted to periods when the flow in Cobb Creek consisted only of groundwater flow, as indicated by stream hydrographs and precipitation records. From these selected points a base-flow rating curve was constructed by plotting mean groundwater stages against stream-discharge rates in corresponding months. From this curve approximations of the monthly and yearly base flow of Cobb Creek between 1953 and 1957 were obtained by using average water levels of the seven wells. Groundwater discharge into Cobb Creek was computed as equivalent to 0.9 inch over the basin, or approximately 15,000 acre-feet per year.

The amount of groundwater discharge by springs is relatively small if the spring occurs near the heads of the tributary stream. Many of these springs do stop flowing when there is a dry year.

> Evaporite deposits, in the Permian age red beds in the study area, have been leached out and have accumulated in the flood plain alluvium. In some areas the seasonal fluctuation of the water table has created areas of highly saline soil and water. Floodwater retarding structures have increased soil and water salinity in some geologic environments in part due to the seepage merging downstream from the structures which has maintained shallow groundwater conditions over a long period of time. (Naney, 1979)

> Moreover, initial analysis of hydrologic data from the Southern Plains Watershed Research Center shows that significant

> reductions can be expected in runoff peaks and sediment loads from watersheds in the area treated with flood-detention dams. This has been shown with peak flow reductions of 16,400 of 8,300 cfs, 1,370 to 400 cfs, and 1,300 to 950 cfs, for three Sugar Creek storms. (Allen, 1979).

Vegetation Cover and Landuse

Field investigation of the study area shows that the vegetation cover differs between the shaly and sandy areas. In sandy sediments, typified by areas along Buggy Creek and areas along Devil's Canyon and Kickapoo Creeks, where deposition is dominant, growth of vegetation is rapid on the alluvium sediments. In a channel the vegetation grows only on areas where deposition has occurred, just below floodwater structures, since the later has reduced the flow in the downstream direction from these dams. Vegetation may aid deposition once it has begun, but fresh alluvium in the channel seems necessary for growth of vegetation.

In Schumm (1961) there was not found the relationship between vegetation and channel alluvium as follows:

> The establishment of vegetation on the noncohesive highly mobile sediments is much more difficult. These channels could be remarkably free of vegetation even when aggrading, but as woon as a veneer of finer sediment is deposited on the sand, vegetation begins to encroach on the channel and eventually completely covers it. (Schumm, 1961).

Data on channel dimensions of several streams of northern Vermont gathered by Zimmerman et al. (1967) have shown that vegetation strongly influences the form of channels in an area with uniform geology. Therefore, hydraulic geometry variables are not only interrelated to landuse and vegetation but landuse and vegetation factors are first order variables that determine the hydrologic and sedimentologic regimen of the stream and, thus, the morphology. Hadley and Schumm (1961) observed sedimentation along channel banks and the adjacent flood plains in ephemeral channels, which increased roughness conditions and correspondingly reduced overall flow velocities. Rapid growth of vegetation along Sandstone Creek in Western Oklahoma caused by sustained seepage from upstread floodwater impounding structures has also been shown by Bergman and Sullivan (1963) to be responsible for changes in channel shape in the period 1957 to 1962.

In the study area, particularly the watersheds on sandstone, the most delicately balanced variable is the spatial distribution of vegetation. The location and density of vegetation in the drainage basins on the sandstone affect tractive force through runoff and resistance through valley floor cover. Climatic changes, and changes in livestock populations in the study area can alter the distribution of vegetation. Carson and Kirby (1975) indicated that

> Some changes from fill to cut are relatively well documented, especially the period of erosion which began between 1880 and 1910 in southwestern United States, and this evidence shows that small changes in the vegetation cover are sufficient to change the situation from one of fill to one of cut. In the southwestern United States, however, it is not clear whether the crucial factor was the introduction of grazing animals or small changes in the intensity of summer rains. (Leopold, 1951; Thornthwaite et al., 1942)

In the absence of cultural factors, increases in low intensity rains, especially in the growing season, will favor denser vegetation, whereas increases in higher intensity rains may have no significant effect on vegetation density, but will encourage higher erosion rates, so that there is no simple relationship even between rainfall and cut or fill. On the whole, however, periods of decreasing aridity tend to be associated with aggradation, and periods of increasing aridity with downcutting.

> In areas where cut and fill alternate frequently, the amplitude of each if likely to be relatively small because of the limited time available. Each time the streams erode, they do so into a consequent surface of their own alluvial deposits, and are therefore likely to cut down in slightly different positions in each episode of erosion. The underlying bedrock will therefore gradually be cut down at all points to form an undulating bedrock surface at a depth of a few feet below the plain surface. In this way a surface not dissimilar from the classic pediment shape in overall form might be produced not as an equilibrium form

> but by the repeated alternation of cut and fill about a mean position. (Carson and Kirkby, 1975, p. 350)

Allen and Welch (1967) stated that [the study area was settled in the late 19th century and most of it was cultivated until the early 1940's. As accelerated erosion reduced soil productivity, much of the upland area was returned to grassland].

> Landuse in the central segment of the Washita River watershed was documented in 1962, 1967, 1971, and 1974 to record changes that might be associated with observed hydrological effects. In 1962 a complete inventory of the Washita River watershed (1130 square miles) was made by aerial observation. An aerial point sampling procedure was developed, tested, and used in the later surveys (Shockey and DeCoursey, 1969). Each of five classifications of cultivated land including 1) sowed crop (planted seeds for growing), 2) summer sowing crop, 3) alfalfa, 4) row crop, and 5) no crop, were planimetered from areal photographs and the remainder were classified as rangeland. That method gave a classification for the Wishita River watershed of 26 percent row crop and 74 percent rangeland. (Naney, 1979)

The landuse inventory (1962-74) shows a significant decrease was in the timber and timbered pasture categories (Sugar Creek, 19 percent in 1967 to 9 percent in 1974) while the change in pasture was increased (Sugar Creek, 36 percent in 1967 to 51 percent in 1974, Salt Creek, 49 percent in 1967 to 52 percent in 1974). Also, the change in cultivation in sandstone drainage basins was stable (25 percent), and a slight increase in shale drainage basins (Salt Creek 37 percent in 1967 to 41 percent in 1974). Streams cut in sandstone are sensitive to removal of vegetation by cultivation particularly when streams are cleaned and lined and banks are alleviated. Cultivation as a factor affecting riparian vegetation was examined by air photos of 1966 and in the field in the summer of 1980. (Shockey & DeCoursey). The percentage of riparian vegetation, therefore, will be used in the analysis to estimate channel cross-section area.

Geomorphic History of the Study Area and Lithologic Formations

> To gain a better understanding of present-day geomorphology in this area, one must look into the past beyond the Cretaceous seas and beyond the building of the Rocky Mountains. Owing to the events that took place almost 150 million years before the disappearance of the Cretaceous seas, the research area offers an excellent opportunity to examine fluvial development upon varying lithologies (Harlin, 1982).

Fay's (1962) classification will be used (Table 5) with the exception of evaporites in the western part of the area and thin carbonates and gypsum beds of the Marlow Formation, the entire area is covered by sandstones, shales, siltstones.

TABLE 5

GENERALIZED SECTION OF GEOLOGIC FORMATIONS IN THE STUDY AREA*

System	Group	Formation	Thickness (feet)	Lithology and water-bearing properties
Quanternary		Alluvium and terrace deposits	0-80	Gravel, sand, silt, and clay on the present and old flood plains of the Washita River and Pond Creek. Yields hard water. Maximum reported yield about 300 gpm.
		High-level deposits	0-25	Unconcolidated gravel occurring as thin, scattered remnants of formerly extensive deposits; on higher ground slumped along valley slopes. Does not generally yield water.
Permiam		Cloud Chief Formation	0-100	Gypsum and anhydrite, dolomitic at the base; clay shale; and silty or sandy shale, which in places resembles the underlying Rush Springs Sandstone. Base of formation marked by Weatherford Member, which in places grades into a dolomitic purple shale or into gypsum. Yields small quantities of water containing large quantities of dissolved calcium sulfate.
	Whitehorse	Rush Springs Sandstone	0-340	Fine-grained, cross-bedded to even-bedded sandstone. Contains large quantities of water suitable for domestic, irrigation or industrial use. Locally, water is hard and contains large quantities of dissolved calcium sulfate. Maximum reported yield about 1,100 gpm, average yield to irrigation wells about 400 gpm.

TABLE 5 --continued

System	Group	Formation	Thickness (feet)	Lithology and water-bearing properties
Permian	Whitehorse	Marlow Formation	0-125	Mostly even-bedded brick-red clay shale, or sandy shale; more sandy toward the northwest. Verden Sandstone Member occurs near the upper-middle part. Emanuel Dolomite Bed at the top. Yields small quantities of highly mineralized water in the eastern part of the area, and moderate amounts of potable water locally in the western part.
	El Reno	Dog Creek Shale	0-300	Mostly even-bedded dark-red gypsiferous clay shale interbedded with gypsiferous siltstone and very-fine-grained sandstone, grading into pure gypsum locally. Does not generally yield water.
		Blaine Formation	0-150	Mostly interbedded gypsum, red shale, and dolomite. Solution cavities containing large quantities of calcium sulfate may yield some water, although the formation generally does not yield water.
		Flower-pot Shale	0-150	Consists mainly of red to reddish-brown shale, and some gray shale. Generally does not yield water.
		Duncan Sand-stone	0-100	Sandstone with minor amounts of interbedded shale and intraformational siltstone conglomerates. Generally water is hard with high sulfate content. Yields 25 to 50 gpm, although the aquifer is capable of yielding more than 100 gpm. Not generally used as an aquifer in the Caddo County area.

*After: Tanaka & Davis, 1963.

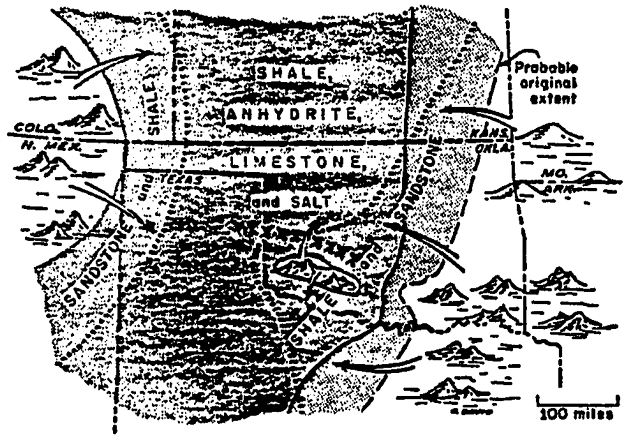


Fig. 3. Early Permian paleogeography of Oklahoma.

From: Johnson, 1971.

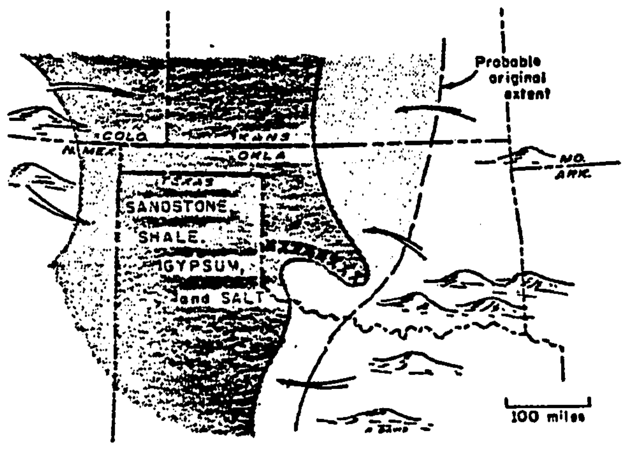


Fig. 4. Late Permian paleogeography of Oklahoma.

From: Johnson, 1971.

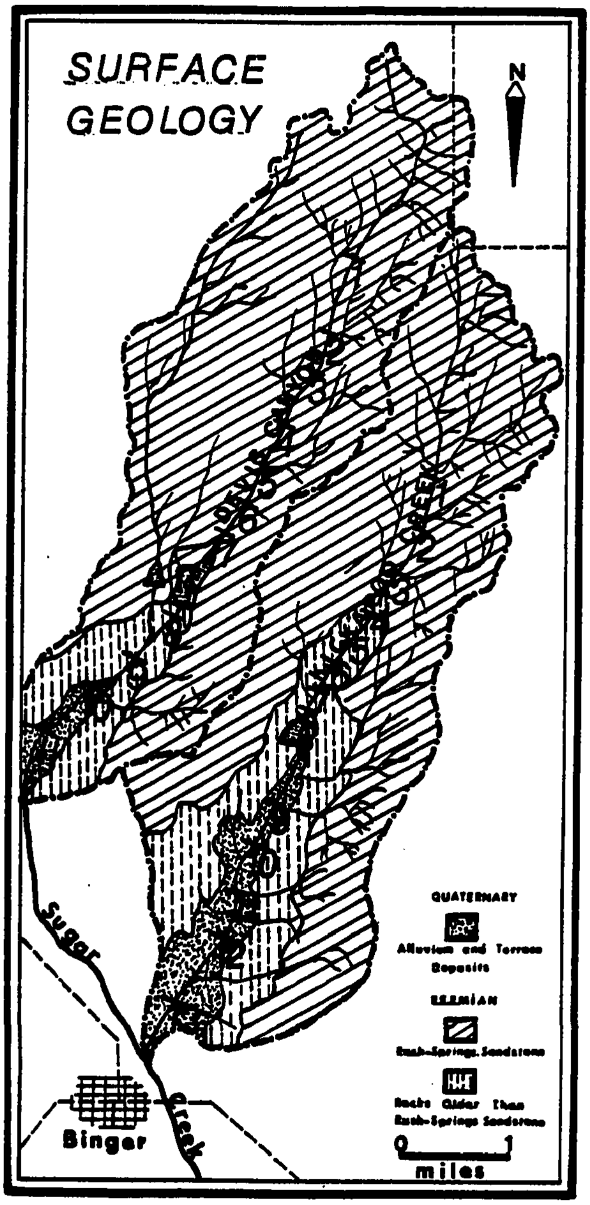


Figure 5

Compiled from: Oklahoma G.S. Circular 61, plate I.

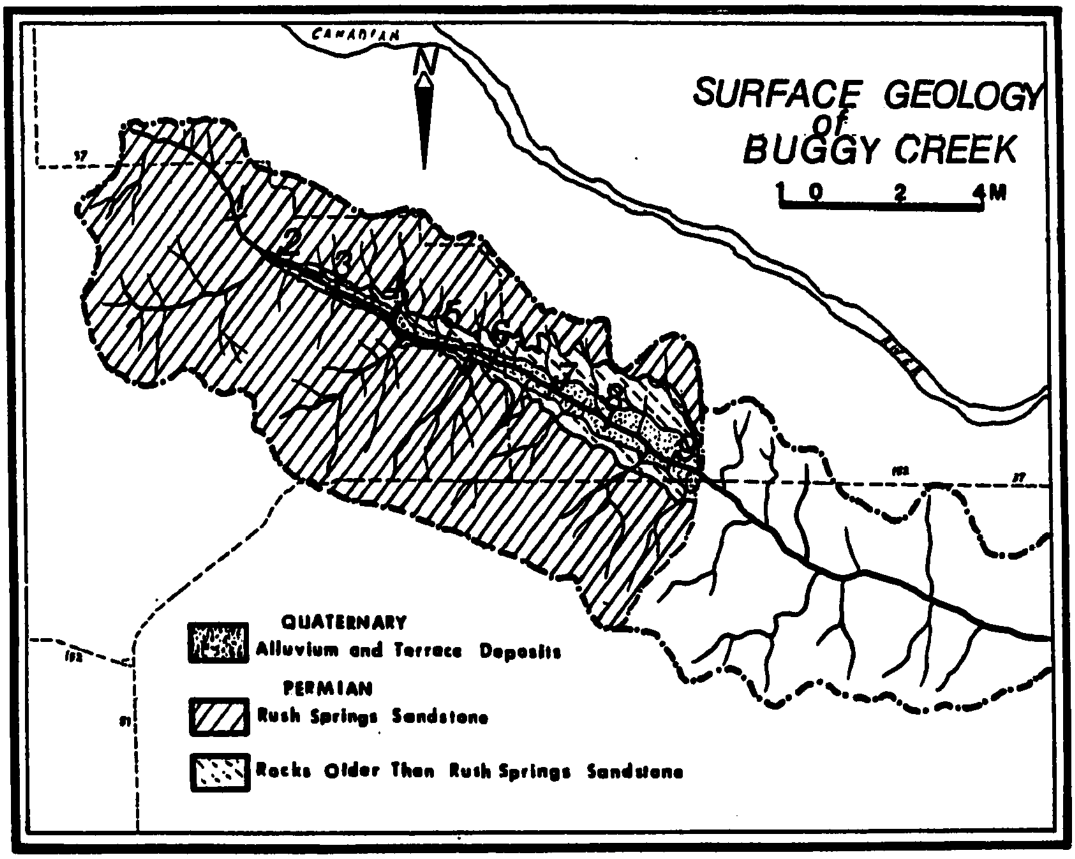


Figure 6

Compiled from: Oklahoma G.S. Circular 61, plate I.

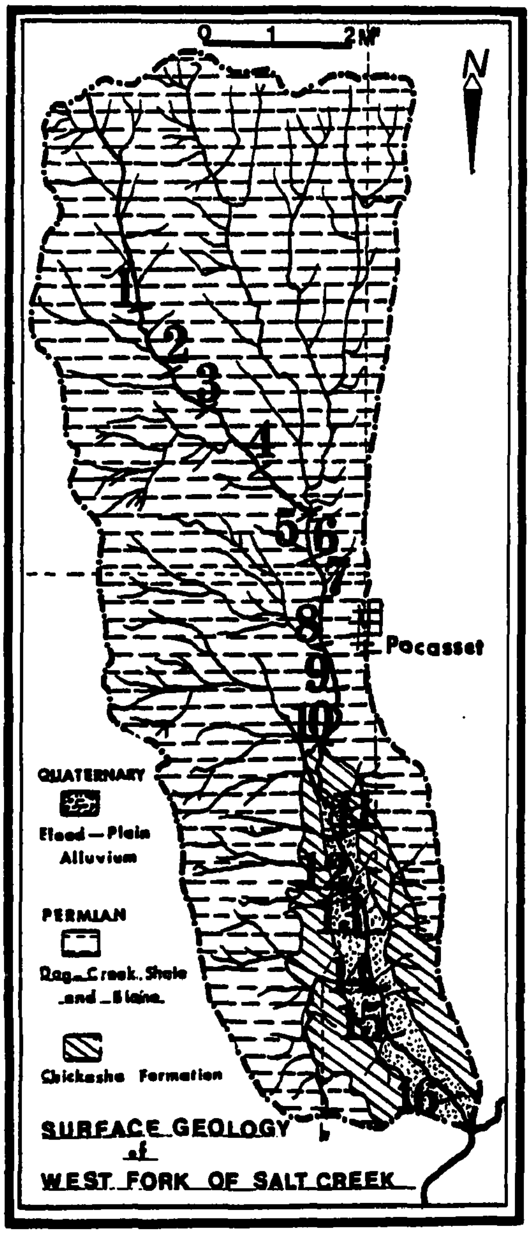


Figure 7

Compiled from: Oklahoma G.S. Bulletin 73, plate I.

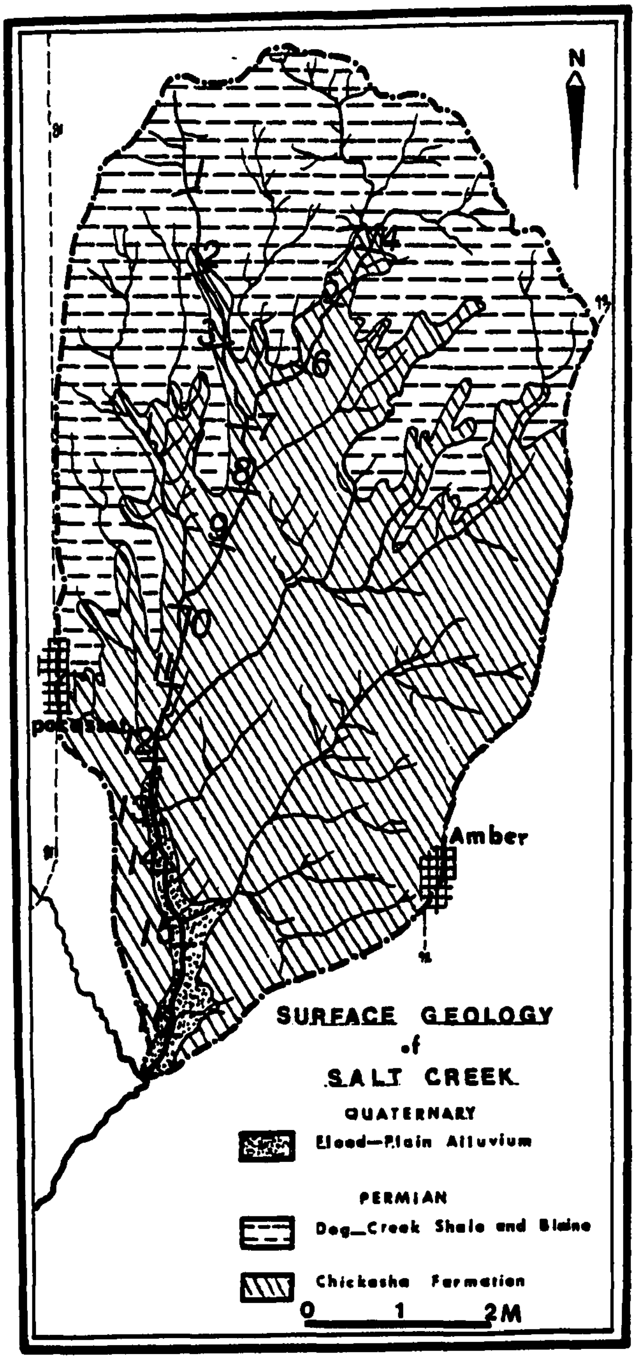


Figure 8

Compiled from: Oklahoma G.S. Bulletin 73, plate I.

The Permian Period: El-Reno and Whitehorse Groups

Thickness of the entire Permian sequence is generally 1,000 to 5,000 feet but reaches 6,000 to 6,500 feet in deeper parts of the Anadarko basin (Freie, 1930, and Tanaka and Davis, 1963).

The Whitehorse Group includes the Upper Permian rocks above the El-Reno Group and below the Cloud Chief Formation. The restriction of the group is in accord with current usage of the Oklahoma Geological Survey and that published on the geologic map of Oklahoma (Miser, 1954). This group is divided into two formations: the Marlow Formation at the bottom, and the Rush Springs Sandstones at the top. The contact between these formations is now widely accepted as being the top of the Upper Relay Creek Dolomite Member of the Marlow (Evans, 1931).

The Marlow Formation includes the sandstone, siltstones, and shales above the Dog Creek Shale and below the Rush Springs Sandstone. The Marlow has a typical moderate reddish-orange color and possesses varying amounts of calcite and gypsum throughout the entire area. Streams have cut through the Rush Springs Sandstone exposing the entire Marlow Formation over much of the selected drainage basins for the study.

The Marlow Formation consists of evenbedded, fine-grained, silty sandstones and shales having a typical moderate reddish orange color. Typical Marlow sandstones are fine-grained, silty, and orange red with individual grains of iron-oxide-stained quartz. The Dog Creek shale is mostly evenbedded dark red gypsiferous clay shale interbedded with gypsiferous siltstone and very fine-grained sandstone grading into pure gypsum locally. Cementing materials consist of calcite, gypsum, iron oxide, and smaller amounts of clay and possibly manganese oxide. The Marlow-Dog Creek sandstones and siltstones are more resistant where they are rich in gypsum or calcite as cementing material. In the area of this study, the Marlow is generally a moderate reddish-orange ranging to a reddish-tan while the Dog Creek is dark red. These formations are composed of approximately equal amounts of even-beeded sandstones, shales, and siltstones. The sandstone members are composed of fine-grained sand with varying amounts of silt and clay. Each individual grain is coated with a veneer of iron oxide, a condition which gives the formation its gross color (Evans, 1931).

The Rush Springs Sandstone conformably overlies the Marlow Formation. The Rush Springs sandstone in Caddo County consists chiefly of reddish-brown very fine filty sandstone. A few calcareous sandy beds occur at random but they are more common in the lower part of the formation. The sand grains composing the Rush Springs in Caddo County are subangular to subround, are remarkably homogeneous in lithologic character, and loosely cemented with iron oxide and calcite. These sandstones probably were laid down along the eastern side of a shallow embayment that was at times cut off, or severely restricted, from the main Permian Sea, because westward from the study area (west of T.11 N.) the sandstones grade laterally into a hydrite and gypsum and gypsiferous silty clay in what must have been a disiccation basin in western Oklahoma during the time of Rush Springs deposition (Tanaka and Davis, 1963).

Howery (1960) identified the characteristics of the Rush Springs in eastern Caddo County as: [the lower Rush Springs contains small 1 mm-sized "sand balls." They consist of white spherical masses of fine sand grains. The white cement uniting these grains appears to be a clay material whose exact identity is unknown. These "sand balls" stand out as pimples on a weathered surface of the Rush Springs Sandstone.]

The Rush Springs Sandstone is highly porous, loosely cemented, and easily eroded. Dense growth of scrub-oak and juniper grow in the red sandy soil derived from its weathering. The Rush Springs is a good ground water aquifer (Tanak & Davis, 1963, p. 21).

Triassic and Jurassic Periods

> Most of Oklahoma was apparently above sea level during the Triassic and Jurassic Periods [Figure 9]. Sandstones and shales in the Panhandle and adjacent areas were deposited mainly in rivers and lakes draining the hills of central Colorado, although some of the sand and mud must have come from lowlands of central and western Oklahoma where the recently deposited Permian sediments cropped out. There is no evidence of the Triassic and Jurassic sediments in the study area. (Johnson, 1971)

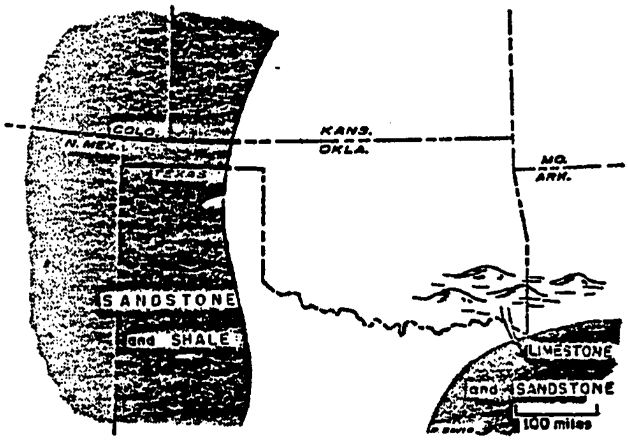


Fig. 9. Triassic-Jurassic paleogeography of Oklahoma.

From: Johnson, 1971.

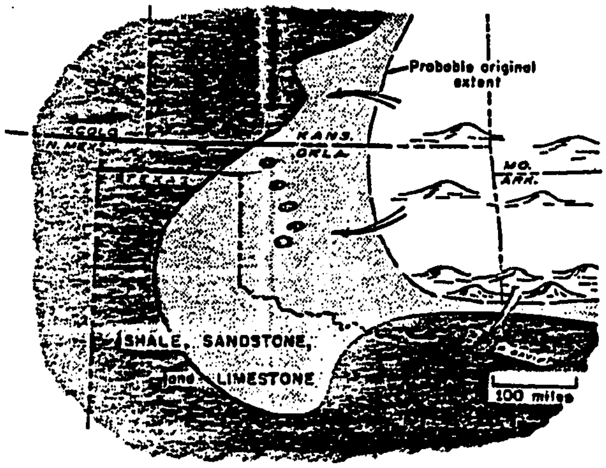


Fig. 10. Cretaceous paleogeography of Oklahoma.

From: Johnson, 1971.

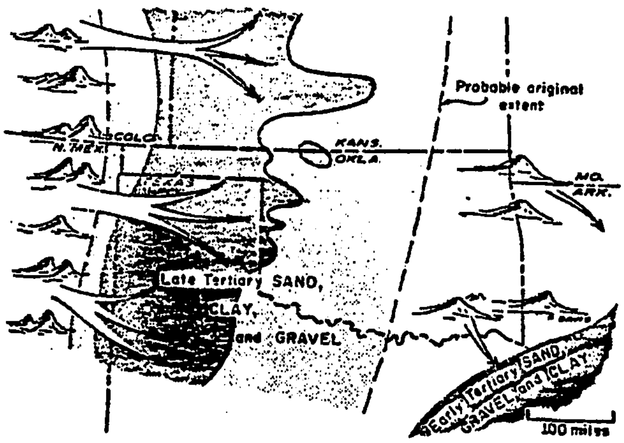


Fig. 11. Tertiary paleogeography of Oklahoma.

From: Johnson, 1971.

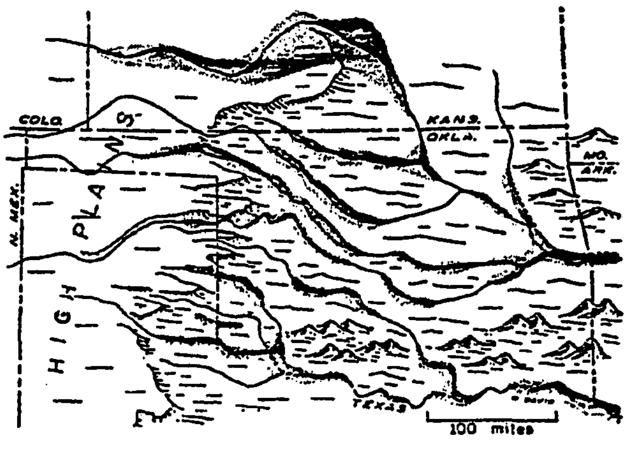


Fig. 12. Major rivers of Oklahoma and principal deposits of Quaternary age.

From: Johnson, 1971.

Cretaceous and Tertiary Periods

Cretaceous seas covered all but northeastern and eastern Oklahoma [Figure 10]. The ancestor of the Gulf of Mexico extended up to and across the southern part of the state, and shallow seas spread northward in the last great inundation of Oklahoma and the western interior of the United States.

The state's pattern of east-flowing drainage had its beginning in the Tertiary Period. The precursor of the Gulf of Mexico extended almost to the southeast corner of Oklahoma in Early Tertiary time [Figure 11], and the shoreline gradually retreated southward through the remainder of the period. Sediments in the southeast include marine and marine sand, gravel, and clay. In late Tertiary time, a thick blanket of sand, clay, and gravel eroded from the Rocky Mountains and was laid down across the High Plains and farther east by a system of coalescing major rivers. These deposits, principally the Ogallala and Laverne Formations, are generally 200 to 600 feet thick in western Oklahoma and originally may have extended across central Oklahoma. (Johnson, 1971)

The Quaternary Period; The Alluvial Deposits

The Quaternary Period is divided into the older Pleistocene Epoch (the "Great Ice Age") and the Holocene or Recent Epoch that we live in today. The boundary between these epochs is set at about 10,000 years ago at the end of the last episode of continental glaciation. While continental glaciers extended southward from Canada as far as northeastern Kansas, Oklahoma's surface was being sculptured by major rivers fed by meltwater from Rocky Mountain glaciers and by the increased precipitation associated with glaciation with glaciation [Evans, 1970). Major drainage systems of today were initiated during the Pleistocene [Figure 12]. The shifting early positions of these rivers are marked by old alluvial deposits that have been left as terraces above the present flood plain. Three distinct alluvial

> terraces, 2 to 5 miles wide, flank the Washita River forming broad flat plains across the area. Correlative terraces are found along most of the major tributary streams and also along the Canadian River at the north edge of Grady County. (Davis, 1955; Johnson, 1971)

The Quaternary Period is characterized as a time of erosion. Rocks and loose sediments at the surface were weathered to soil, and the soil particles are then carried away to streams and rivers. In this way, hills and mountain areas were worn down, and sediment transported to the sea or temporarily deposited on the banks and in the bottoms of rivers and lakes. Sand, silt, clay, and gravel deposits of Pleistocene and Holocene rivers and lakes are unconsolidated and are typically 25 to 100 feet thick. Pleistocene terraces 100 feet to more than 300 feet above modern flood plains attests to the great amount of erosion and downcutting performed by major rivers in the past one million years.

> Many of the major streams' valleys and larger canyons have been filled with a silty sand deposit. At present the streams are flowing at an elevation below the top of these sediments, indicating that the streams are again cutting down the previous levels. The thickness of these alluvial deposits is unknown but several of the streams are flowing on these deposits 20 feet below their upper limits. (O'Brien, 1963)

The Washita River flows across a broad floodplain composed of unconsolidated sands, silts, and clays. No gravels were seen on the floodplain. The thickness of these deposits is unknown but in several places along the river, the river bed is 30 feet below the adjacent floodplain.

> Alluviation has at least at one time filled much of the existing canyon system in the area south of the Canadian River in northern Caddo County. Compaction of the fill took place and controls the drainage in the incompletely excavated canyons. Various stages of re-excavation are present throughout the area. Remnants of semi-indurated canyon fill sediments can be seen along the bases of some of the vertical walls of the canyons. (Howery, 1960)

> Recent alluvium deposits are present along most of the stream valleys of this area. Low level terraces of the canyon fill sediments and the stream alluvium are mapped under the general heading of recent alluvium. (Howery, 1963)

> Alluvium is the unconsolidated gravel, sand, silt, and clay underlying the present flood plains of the existing streams. The principal constituents are derived from nearby Permian exposures and Quaternary terraces. As much as 50 feet of thickness is present along the main stream valleys. (Howery, 1960)

Summary

> The channel of Oklahoma's larger rivers are aggrading today. Periodic flooding sweeps segments of channel only to bring additional fill from upstream and from numerous upland tributaries. Stumps buried in as much as 100 feet of canyon fill at the mouth of a Washita tributary in Caddo County, Oklahoma, have been dated (Hall, 1982) at 3,200 years before the present and attest to the degree of adjustment since the last glacial melt. However, enough relief remains from the Cretaceous uplift so that upland surfaces are now undergoing channel initiation and development much as they have for perhaps 100 million years. (Harlin, 1982).

An important aspect of the characteristics of the alluvium in the valley floor is the difference in general appearance in the valley between where the channel is completely filled and grassed and where it is entrenched. In the flat-floored valley, typical of Devil's Canyon on sandstone, the alluvium may contain much water, thereby supporting heavy vegetation. In addition, a veneer of fine sediment generally overlies the coarser sediment. In predominantly sandy areas the valley floor will be an area of much finer sediment. Upon renewed trenching, however, the coarser sediment will be exposed and a wide shallow channel will form, which will appear inconsistent with the fine surface sediment. Thus, in studying erosion and deposition in the study streams with watersheds composed of coarser sediment, soil profiles in the alluvium should be investigated.

Soil is the product of the interaction of the five major factors of soil formation--climate, parent material, plants and animals (especially vegetation), relief, and time. If in one area the parent material is different from that in another area, but the other factors are the same, the soil formed in one area differs from that formed in the other area. The physical characteristics of soils and vegetation cover have a great effect on channel stability. During the past 30 years man has altered the natural soil-forming processes in much of the study area. In clearing the land and cultivating the soils, man has caused severe loss of soils through sheet and gully erosion, typical of the sandstone area.

Cook and Reeves (1976), p. 173) characterize the controversy around the effect of soil and vegetation cover on the production of runoff as:

> The foundation of most regional hypotheses--that vegetation cover and soil conditions regulate the production of runoff from slopes during storms --is rather surprisingly one of the weakest links in the rainfall-erosion argument. Two observations emphasize this weakness. First, the overwhelming strength of rational argument that both decreased vegetation cover and disturbance of soil should act to increase runoff (Luff, 1964; Storey, Hobbá, and Rosa, 1964), conflicts sharply with the results of some empirical field studies on the hydrological effects of removing vegetation from various watersheds in California. Secondly, if one assumes that rational arguments are correct (that runoff is controlled by vegetation and soils), it is difficult to arrive at estimates of how much additional runoff might have accompanied the various secular or short-term vegetation changes that are known to have occurred. It is quite possible, for example, that increases in discharge resulting from such changes in watershed conditions were insignificant relative to the hydraulic effects of localized modifications of valley floors.

Also the type of the alluvium composing the stream channels in the watershed will affect the hydrologic regime of the drainage basin. In basins developed on the sandstone, the materials composing the valley floor are coarse and non-cohesive. A decrease of vegetation cover accelerated the erosion but meanwhile there is no increase generated in runoff to transport the eroded materials out of the system. On the other hand, cohesive and fine materials composing the valley floors of the basins developed on shale, generate more runoff and at the same time accelerated erosion does not occur.

Therefore, the differential erosiveness and cohesiveness of the sediment composing the watershed can have a great effect on the amount of runoff and in turn affect the hydraulic factors in the basin's stream channel. Then the aspect of the relationship between erosion and deposition mechanics in stream channel and lithology and in turn their relation to channel form is the focus of this research.

CHAPTER IV

DEFINITION AND DERIVATION OF THE CONTROLLING FACTORS

To identify the focal point of the study, I used the term "geologic-lithologic-variation in west-central Oklahoma" to denote the differences in dynamics and morphology of streams in the area. The spatial distributions of soils, vegetation, and morphometric parameters of the drainage basins are designated as independent variables so as to take into account their probable predictive value of channel morphology (width-depth ratio and/or channel cross-section area). It is on this set of variables, therefore, that this chapter will focus.

The Morphometric Parameters of the Drainage Basins and Their Streams

Morphometric parameters may refer to slope gradients, slope lengths and drainage texture. Once a drainage basin and its stream had been defined, measurements are made of its morphometric parameters.

Channel Form

At all cross-sections width, depth, and slope of the channel were measured. The lowest part of the channel from the edge of the first surface or bank above the channel floor is the maximum depth of the stream to represent a bankfull discharge condition. In all cases, the channel width was measured from the edge of the bank to the corresponding elevation on the opposite side of the channel by a tape measure. The depth of the channel was measured by a pole from a horizontal level between the banks to the channel bottom (Schumm, 1960, 1961). The gradient of the channel at a cross-section was measured with a hand-level. To confirm field measurements of gradient in its relation to channel form (Park, 1976), similar measurements of gradient were measured from topographic maps (1:24,000).

The shape of the cross-section expressed as a dimensionless width-depth ratio was obtained from the measured data of width and depth of the channel in order to plot it against the weighted mean percent of silt-clay (M). Moreover, the cross-section area was calculated to relate channel capacity (C) as a dependent variable to the total stream length (ΣL) above a particular channel cross-section.

Tractive Force

DuBoys equation for tractive force was used to estimate the force conditions at a cross-section ($\tau_o = \gamma ds$; where τ_o is the tractive force condition, γ is the specific weight of one cubic foot of clear water; 62.4 pounds; d is the depth of channel at a bankfull discharge condition; and s is the slope of the channel at cross-section). There are some threshold values of valley floor gradient, which if exceeded, result in channel entrenchment (Schumm, 1973; Schumm and Hadley, 1957; and Patton and Schumm, 1965). The tractive force condition explains why gradient is not the only variable with a significant threshold (Graf, 1979a), for example similar thresholds exist for the resisting vegetation (Zimmerman, et al., 1967).

The Hypsometric Function: The Gradient Aspect of the Drainage Basin

The simplest relatively effective index of basin gradient is the relief ratio (Schumm, 1956). This, however, combines two rather uncertain measurements. Basin length is subject to a variety of alternative definitions (Goudie, 1981); thus Gardiner (1975, p. 18) lists nine alternative definitions based on the basin outline, relief distribution or stream net, and Cannon (1976) offers further definitions and compares their effect upon the values of morphometric variables. Basin relief may be greatly influenced by one isolated peak within the basin and may therefore be unrepresentative. The hyposometric integral (Strahler, 1952) offers a conceptually attractive alternative for expressing overall relief aspects of the basin, despite criticisms recently made of the index (Harlin, 1978). The hypsometric function is an approach to show the percentage of an area that lies above and/or below a certain elevation.

There are two main difficulties with the hypsometric curve as a generalized slope profile: 1) The hypsometric integral settles down to a value of about 50 percent as a basin attains "maturity" and thereafter changes little. 2) If the hypsometric curve is considered as a frequency distribution, it is clearly remarkable that the form of the curve is similar to that for a normal distribution, even allowing for the high degree of auto-correlation among the elevations of points in a basin. From this viewpoint, the curve cannot be expected to look at all like a single slope profile, since the individual basin profile are all over different elevation ranges (Carson & Kirkby, 1975, p. 401). It appears preferable, except in the rare cases (e.g., Schumm, 1956a) where the "original" basin relief is known, to define the shape of the curve by its skewness; high hypsometric integrals being associated with negative skewness and low integrals with positive skewness (Evans, 1970). See Appendix 2; hypsometric curves of the study's drainage basins.

The true potential of the hypsometric method in quantitative landform analysis cannot be realized through applications of the integral value alone. A second expression of the hypsometric curve form is the curve's density function. This function can be expressed by the statistical moments of the curve. The calculation of statistical moments can monitor changes along the hypsometric function (Harlin, 1978). The hypsometric curve is, by definition, monotonic and can be represented by a continuous polynomial function of the form: $f(x) = b_o + b_1x + b_2x^2 + . . . + b_nx^n$. The polynomial function of this equation was obtained by the least squares format.

Hypsometric curves have an advantage over other profiles when the regression format is employed, in that the curves are monotonic. The relatively simple forms were fitted to low order (3rd) polynomial which were highly significant.

The procedure to generate the energy slope factor at the measurement sites (statistical moments of the curve for the cross-section sites) is to mark these sites on the curve and calculate the $\frac{dy}{dx}$ values. The derivative $(\frac{dy}{dx})$ was obtained as the derivative of the regression equations calculated for the study drainage basins. The derivation of the regression function is in the form; $\frac{dy}{dx} = b_1 + 2b_2x + 3b_3x^2 + \ldots + nb_nx^{n-1}$. The channel cross-section area (C) will be plotted against the hypsometric function $(\frac{dy}{dx})$ in order to determine the effect of the distribution of elevation over the area on channel capacity.

Channel Length and Drainage Density Above a Cross Section-Site

Drainage area was measured by a planimeter from topographic maps (1:24,000). Stream lengths were measured from the same topographic maps with an opisometer. Drainage density was then calculated from the ratio between stream length and drainage area ($Dd = \frac{\Sigma L}{A}$; where ΣL is the total stream length and A is the drainage area). To obtain an index of network volume, it is necessary to relate the length of network to the capacity of the channel (channel cross-section area) to which the network is tributary. Within each drainage basin, measurements of total stream length above each cross-section were made. These measurements serve as a basis for an index of stream network volume. Several workers have related stream channel capacity to drainage area (Harvey, 1969; Gregory et al., 1974) or stream channel width to drainage area (Day, 1972), and stream capacity to stream network volume (Gregory, 1977). As stream lengths increase channel capacity increases in the downstream direction. However stream channels below both abstracting and regulatory reservoirs are liable to alter in size and possibly channel pattern.

Subsequent studies of streams below small reservoirs controlling catchments of up to 50 km^2 have shown complex patterns of channel change which are reviewed by Petts (1979). Channels below reservoirs may passively accommodate reduced peak flows or they may adjust in one of three ways: clearwater degradation, aggradation, or pattern metamorphosis. The last is not reported frequently in the literature but clear water erosion is indicated by channel capacities more than 30 percent greater than expected immediately below two of the 14 reservoirs investigated by Lewin (1981) in British rivers. Inferred reductions in capacity of at least 30 percent were found immediately below six dams and at one or more sections further downstream on all 14 rivers investigated by Lewin (1981), the maximum reduction in capacity being 75 percent which is about the same as the maximum reduction in peak flows (Petts and Lewin, 1979). Gregory and Park (1974) showed that the reduction in capacity is substantial immediately below the dam, but becomes progressively less marked downstream as the fraction of the catchment controleld by the reservoir decreases.

Sediment Properties

Samples of sediment were taken from the stream channel bed and banks. The samples were taken from at least three points along the cross-section depending upon channel width. Samples representing channel banks sediments were taken from the edge of the channel to avoid sampling from transported soils from the adjacent fields. Only the properties of sediment pertinent to sediment transport (particle size analysis) will be discussed in this section.

Particle Size and Sieve Analyses

Grain size is the most important property affecting transportability of sediment. At one extreme of the spectrum of sizes are large boulders which are rolled only by violent mountain streams, and at the other, fine clay, which once suspended, requires days to settle. For clay, silt, and fine sand it is practical to determine the sedimentation diameter by settling analysis methods such as 1) pipette, 2) bottom withdrawal tube, 3) hydrometer and 4) visual accumulation tube. For sand and gravel sieve analysis is convenient, and for gravel boulders direct measurements can be made by calipers or by volumetric displacement. It should be noted that terms such as wash load, suspended load, etc., refer to method of transport in any particular situation and do not correspond to any fixed size graduations.

Some recommendations need to be considered when performing sieve analyses. If the particle is smaller than the opening or the sieve, it will pass easily, of course, once it gains access to an opening. However, two factors can reduce the probability of access, namely: 1) overloading, i.e., so much material on the sieve that the particle never gets shaken down near an opening, and 2) clogging, which may occur when there is a large proportion of the sample just a little too large to pass the screen. The effect of clogging was reduced by reducing the size of the sample.

The American Society for Testing Materials (ASTM Standards 1955, Part 3, C 136-46) requires, for material 95 percent of which is smaller than 2.0 mm (10 mesh), that the sample be approximately 100 grams. However, their purpose is to specify a standard test for aggregates, where the desired degree of accuracy is not as high as in research on sediment transportation. In particular, for well-sorted fine sands, 100 grams could lead to considerable overloading. As a general rule Shergood (1946) and ASTM (1972) found that for river sands, the sample should be limited to between 100 and 150 grams for coarse sand and between 40 and 60 grams for fine sands. The lithology composing streams in the study area ranges from coarse sand and fine sandy silty to clay material, so that a 100 grams of a dry and ground sample was sufficient to load on the number 10 mesh sieve. A sieving time of 5 minutes was adequate and was the period after which the greatest sieve fraction began to lose a constant amount for each succeeding minute (ASTM, 1972; also see Miller, 1970).

Silt-Clay Percent in the Channel Perimeter

The information about the proportion of silt and clay in sediment samples were obtained from the hydrometer analysis. The determination of percent silt and clay in a sediment alluvium is usually referred to as a mechanical analysis. The Buoyoucos method of mechanical analysis is based on differential settling rates of the separation; that is, particles settle differently in water depending upon the amount of the surface area per unit volume. Clay particles, for example, have a high amount of surface area per unit volume and settle slowly, while sand particles settle rapidly because of their low specific surface area. A hydrometer can be used to measure the density of a sediment suspension; density being a function of the amount of particles in suspension (Day 1979). (See Appendix 3.)

Vegetation Retardance Index

The Manning equation expresses velocity as a function of hydraulic radius and slope,

$$V = \frac{1.49}{n} R^{0.6} S^{0.5}$$

where V is the mean velocity, R is the hydraulic radius, S is the slope, and n is the roughness factor. Anything which affects the roughness of the channel changes n, including the size and shape of grains on the bed, sinuosity, obstructions in the channel such as vegetation, logs, piers, and sandbars, and any irregularity in channel section (Marrisawa, 1968). Table 6 compiled with data adapted from Chow (1959, pp. 112-113) illustrates the change of the value of n by changing the roughness condition in the channel.

TABLE 6
VALUES FOR ROUGHNESS COEFFICIENT n FOR NATURAL STREAMS

Type of Channel	n
a - Streams on Plain (top width at flood stage < 30 m)	
i - clean, straight, no deep pools	0.025-0.033
ii - clean, winding, some pools and shoals	0.033-0.045
iii - sluggish reaches, weedy, deep pools	0.050-0.080
b - Mountain Streams	
i - bottom: gravels, cobbles, and few boulders	0.030-0.050
ii - bottom: cobbles, large boulders	0.040-0.070
c - Flood Plains Streams	
i - pasture, short grass	0.025-0.035
ii - cultivated, no crop	0.020-0.040
iii - cultivated, mature crops	0.030-0.040
iv - medium to dense brush in winter	0.045-0.110
v - medium to dense brush in summer	0.070-0.160

TABLE 6--continued

Type of Channel	n
Flood Plains (Morisawa, 1968, p. 38)	
i - pasture, no brush, short grass	0.030
ii - pasture, no brush, high grass	0.035
iii - brush, scattered to dense	0.050-0.100
iv - trees, dense to cleared, with stumps	0.150-0.040

From the table, it is clear that the main roughness element in a valley floor's streams, that is, the principle resistance to flow, is vegetation. Removal of riparian vegetation could lead to a reduction of surface roughness and thus increase the erosiveness of valley-floor flows. Stability of channels, therefore, is attributed to roots of grass and shrub vegetation, which provide a riprap-like protection to silt banks from fluvial erosion (Smith, 1973). Moreover, the expanding roots of trees may detach soil or may also bind or stabilize (Pitty, 1971). The difference of retardance to flow between grassy channels and tree lined channels depends, in part, on the hydraulic variables in the Manning equation (Chow, 1959, p. 179). Melton (1965), and Simons & Richardson (1962) found higher roughness (n) for sandy grassy streams. Cooke and Reeves (1976) reported that Melton (1965), in an exploration of the potential of precise hydraulic arguments in explaining entrenchment in southern Arizona, recognized that the Manning equation is not applicable under conditions in which a large retardance coefficient is provided by high, dense cienega grass.

Melton argued "prior to entrenchment, reasonable figures for hydraulic radius (R) would be 0.5 feet and for the slope (S) 0.02, in the upstream reaches. Then taking VR from Chow (1959, p. 182, which is the product of the mean velocity of flow V and the hydraulic radius R), the following table can be constructed of velocity (V) in relation to roughness (n), which applies to grassy channels.

Roughness n	VR	Velocity (V) feet/second
0.08	6.8	13.8
0.10	4.2	8.4
0.15	2.5	5.0
0.20	1.6	3.2
0.40	0.6	1.2

Note that the gradient does not enter into these figures. For easily eroded soils such as the cienega deposits, a maximum stream velocity of about 6 feet per second is permissible, and, as the actual values were probably much below that, the roughness was necessarily greater than 0.1. If we assume that under new conditions immediately prior to entrenchment, the roughness coefficient was reduced to a value comparable to modern sandy arroyos, for example, 0.02 and the hydraulic radius increased to 1.5 feet, then Manning's formula becomes applicable:

$$\bar{V} = \frac{(1.49)(1.5)^{0.667}(0.02)^{0.5}}{0.02} = 13.7 \text{ ft/sc.}$$

This would put the velocity above the critical value"(Melton, 1965, p. 33).

A point of general importance emerges from this discussion. The degree of effect of vegetation on stream channel is considered as; a) turf grasses where the average depth of flow is less than one-half the height of vegetation, b) growing season—bushy willows, intergrown with weeds in full foliage along side slopes with depth of flow up to 10 feet, and c) growing season—trees intergrown with weeds and brush all in the full foliage, with any value of hydraulic radius up to 10 or 15 feet (Chow, 1959, p. 108).

This study examined the force-resistance interplay in the channel. The measurement of force was estimated by calculating the tractive force condition in the channel. Then, the measurement of resistance to erosive flows is made by estimating the resistance index of vegetation cover within the channel perimeter, and also, by determining the silt-clay percentage in the sediment composing the channel perimeter. The study of the land form-vegetation relationship, therefore, may be best conducted with historical data for both variables. In many ways the problem of explaining vegetation change is similar to that of explaining channel cutting or widening. The present patterns can be explained in a variety of ways and conflicting hypotheses have rarely been conclusively evaluated (Cooke & Reeves, 1976).

> Vegetation can be quantified in many ways. Most methods relevant to the geomorphologist will consist of measurements of vegetation density and

> species composition. Floristic analysis usually relies on sampling techniques, and the most well known are the transect method (line transects or belt transects) and the 'quadrant' method. In the former method, every specimen that touches or shades a straight line or belt is recorded, together with the relative distances along the line. This method is recommended for rather homogeneous communities. The latter method consists of sampling in a delineated area which can have any form. Goudie, 1981, p. 269.

Measurements of riparian vegetation cover were made on a picture taken in the field for the channel cross-section. A quadrat was defined on the picture covering an area of approximately 2,000 square feet. The quadrat consists of the adjacent areas of the flood plain on both sides of the channel, and approximately 100 feet upstream and downstream from the cross-section site. Then, four lines were drawn transecting the channel in the quadrat. Areas along the line covered with trees were measured relative to the length of the line. The same measurement was made for areas along the line covered with grass. The distortion in the picture was treated mathematically (Harold Malde, U.S.G.S.) Journal of Research, vol. 1, 1973, p. 193). See Appendix 4.

The percentage of each line transects covered with trees was multiplied by a constant (ρ_1), and the percentage of the line covered with grass was multiplied by another constant (ρ_2). The vegetal retardance index of riparian vegetation, therefore, was obtained by the following equation calculated for each line transects:

$$V = \rho_1(T) + \rho_2(G) \tag{25}$$

where V is vegetation resistence index, T is the percent cover with trees, G is the percent cover with grass, ρ_1 is a constant representing the partial weight of trees coverage to resist the erosive flow, and ρ_2 is a constant representing the partial weight of grass coverage. ρ_1 and ρ_2 are unknown constants in equation (25). In order to determine these constants, a trial and error method was applied using the following values:

TABLE 7 POSSIBLE VALUES FOR THE CORRECTION FACTORS IN EQUATION 25

ρ_1	ρ_2
0.1	0.9
0.2	0.8
0.3	0.7
0.4	0.6
0.5	0.5
0.6	0.4
0.7	0.3
0.8	0.2
0.9	0.1

In Chapter I a relationship between channel cross-section area and the ratio between tractive force and vegetation retardance index was hypothesized as follows:

$$C = a + b \frac{\tau_o}{V} \tag{26}$$

where C is the cross-section area, a and b are the regression constants, τ_o is the tractive force value, and V is the vegetation retardance index. Channel capacity is also a function of several primary factors affecting channel form such as:

$$C = f(G, \frac{\tau_o}{V}, M, \Sigma L) \tag{27}$$

where G is the type of lithology composing the channel perimeter, M is the percent silt-clay in the sediment in the channel bed and banks, and ΣL is the total stream length above the section site. However, statistically a significant relationship between channel cross-section area and the ratio between tractive force and vegetation, as in equation (26), can be established. Equation (26) can be written for each observation of a group of points having equal tractive force value.

$$\begin{aligned} C_1 &= a + b\,\frac{\tau_o}{V_1} \\ C_2 &= a + b\,\frac{\tau_o}{V_2} \\ C_3 &= a + b\,\frac{\tau_o}{V_3} \\ &\vdots \\ C_n &= a + b\,\frac{\tau_o}{V_n} \end{aligned}$$

Subtracting every two consecutive observations from each other or:

$$C_1 - C_2 = b\,\tau_o\left(\frac{1}{V_1} - \frac{1}{V_2}\right) \tag{28}$$

$$C_2 - C_3 = b\,\tau_o\left(\frac{1}{V_2} - \frac{1}{V_3}\right) \tag{29}$$

dividing equation (28) by equation (29) as

$$\frac{C_1 - C_2}{C_2 - C_3} = \frac{(1/V_1 - 1/V_2)}{(1/V_2 - 1/V_3)}$$

$$= \frac{(V_2 - V_1) V_2 V_3}{V_1 V_2 (V_3 - V_2)}$$

$$= \frac{(V_2 - V_1)\ V_3}{V_1 (V_3 - V_2)} \tag{30}$$

Substituting equation (25) for V in equation (30)

$$\frac{C_1 - C_2}{C_2 - C_3} = \frac{[(\rho_1(T_2)+\rho_2(G_2)) - (\rho_1(T_1)+\rho_2(G_1))]\,[\rho_1(T_3)+\rho_2(G_3)]}{[\rho_1(T_1)+\rho_2(G_1)]\ \ [\rho_1(T_3)+\rho_2(G_3))-(\rho_1(T_2)+\rho_2(G_2))]} \tag{31}$$

Data for C, T, and G were measured. To determine the values of ρ_1 and ρ_2 one must solve equation (31) using the values of ρ_1 and ρ_2 in Table 7 in an iterative way. The mathematical solution of equation (31) may not exist since, in fact, channel cross-section area is a function of several factors as in equation (27). Fortunately, the channel cross-section area holds a certain relationship with the ratio between tractive force and vegetal retardance index. This relationship is characteristic of the vegetation and practically independent of silt-clay percent in the sediment composing the channel, and total stream length above the cross-section site. As a result, therefore, equation (31) is solved statistically by the least square method (see Appendix 5).

Subsequently, the independent variables which were chosen to predict channel cross-section area (channel capacity) and/or width-depth ratio are a result of the review of the literature presented in the theoretical framework in Chapter II. These variables were defined and derived in this chapter. The independent and dependent variables are listed in Table 8.

TABLE 8

LIST OF THE DEPENDENT AND INDEPENDENT VARIABLES

Variable	Symbol	Units	How derived or defined	Source
1. Geology (lithology)	G	dummy; takes the value of 1 for the shale formation and the value of 2 for the sandstone formations	Mapped from geologic maps of Caddo, Canadian and Grady Counties, Oklahoma, in Oklahoma geological survey, Norman, Bulletins 73, 87, and Circular 61	Oklahoma Geological Survey Bulletins 73, 87 and Circular 61
<u>Channel Morphology</u>				
2. Channel width	W	feet	measured in the field	
3. Channel depth	D	feet	measured in the field	
4. Channel cross-section area or channel capacity	C	square feet	channel width x channel depth	Gregory (1977)
5. Width-depth ratio	F	ratio	channel width ÷ channel depth	Schumm (1960)
6. Tractive force	τ_o	pound/square foot	$\tau_o = (\gamma)(d)(\theta_c)$ where: τ_o = tractive force γ = specific weight of water (62.4 pounds) d = depth of water (bank-full conditions) θ_c = channel slope at the cross-section site	Graf (1979a)

TABLE 8--continued

Variable	Symbol	Units		Source
<u>Drainage Basin Morphometry</u>				
7. Statistical moments of the hypsometric curve or energy slope factor at the cross-section site	S_h	dimensionless	$\frac{dy}{db} = b_1+2b_2X+3b_3X^2+\ldots+nb_nX^{n-1}$ The above equation is the derivative for the polynomial function for the hypsometric curve using the regression format	Harlin (1978)
8. Total stream length above a cross-section site	ΣL	miles	Measured from large scale topographic maps (1:24,000) with an opisometer	Gregory (1977)
9. Drainage basin density	Dd	miles/sq. mile	Calculated dividing total stream length by area of the drainage basin above a cross-section site	Emmett (1975)
<u>Physical Properties of Sediment Composing the Channel Perimeter</u>				
10. Silt-clay percent in the channel bed	S_c	percentage	Determined by grain-size analysis; the hydrometer analysis	Schumm (1960)

TABLE 8--Continued

Variable	Symbol	Units	How derived or defined	Source
11. Silt-clay percent in channel banks	S_b	Percentage	Determined same as #10	Schumm (1960)
12. Silt-clay percent in channel perimeter	M	percentage	$M = \frac{S_c x w + S_b x 2d}{w + 2d}$ where: S_c = silt-clay percent in the channel bed S_b = silt-clay percent in the channel banks d = channel depth w = channel width M = silt-clay percent in channel perimeter	Schumm (1960)
13. Vegetation retardance index	V		$V = \rho_1(T) + \rho_2(G)$ where T is the tree cover, G is the grass cover, and ρ_1 & ρ_2 are correction factors and are defined in Appendix 5.	

CHAPTER V

METHOD OF STUDY

This chapter focuses on sampling procedure and the statistical methods which will be used in the analysis.

If, as stated in the general hypothesis, geology (lithology) creates patterns of environmental characteristics, this should be reflected in the spatial distribution of the morphometric variables and sediment properties in the area. The relationship betwen lithology and channel characteristics may be brought out by numerical analysis through comparing channel form characteristics on different lithologies. Analysis of covariance may be used to test the significant difference between channel form on varying lithology by using geology as an indicator factor.

Sampling Method

The study focuses on the spatial variation of channel morphology, and the change of channel form along streams in west central Oklahoma. Therefore, geographic sampling procedure in collecting the data is important. Berry et al. (1968, p. 91) identify three types of geographic sampling procedures; 1) random sampling, 2) systematic sampling, and 3) stratified sampling. Because the nature of the distribution of the morphometric characteristics and sediment properties are unknown within each geologic formation, a choice of an optimal sampling was difficult. However, it appears that a stratified systematic sampling scheme was desirable and efficient.

With respect to geology (lithology), it is desired to restrict the choice of drainage basins to a study area subdivided into two stratas; one strata represents sandstone and the other represents shale. Some watersheds of the central segment of the Washita River and the South Canadian River were selected according to geology. Devil's Canyon Creek (10.16 square miles), Kickapoo Creek (12.7 square miles), and Buggy Creek (51.2 square miles) were selected on sandstone. Within the Anadarko-Alex reach of the Washita River, West Fork of Salt Creek (26.65 square miles), and Salt Creek (41.50 square miles) were selected on shale.

The choice of Buggy Creek was necessary, because it is not treated by floodwater retarding structure. Many of the streams in the area, particularly those on sandstone, are treated with these structures. It was found that the seepage from reservoirs on these streams has accelerated the growth of riparian vegetation, which has led to a change in channel geometry and to a reduction in channel conveyance (Gilbert, 1959; Sullivan and Bergman, 1963; and Kennon, 1966). Devil's Canyon Creek and Kickapoo Creek, as streams treated with the flood retarding structures, along with Buggy Creek, as a natural stream, explain the effect of man-induced changes on stream channel morphology. Also, Buggy Creek provides a valid comparison between streams on sandstone and streams on shale, since the latter are not treated with flood retarding structures. However, Buggy Creek is modified by another type of man-induced change in natural streams. This type of intervention is known as channelization.

Field investigation of streams in areas of sandstone indicated that accelerated meandering, such as changing the course of the stream, and/or channel straightening occurred, partially man-induced. Some landowners moved the stream channel from their fields and aligned the stream to the road.

Measurement sites along the study streams were located at a distance of approximately 200 feet upstream direction from the county and/or private roads crossing the stream channel. That is the cross-sections were determined along each stream at approximately equal intervals. However, these intervals may differ from one stream to another depending upon the stream length. Eleven cross-sections were surveyed on Devil's Canyon Creek, and thirteen on Kickapoo Creek at intervals of approximately two-thirds of a mile. Nine cross-sections were surveyed on Buggy Creek at intervals of two miles. Sixteen cross-sections were surveyed on West Fork of Salt Creek at intervals of two-thirds of a mile, and sixteen on Salt Creek with a distance of one mile between the cross-sections. A total of

65 cross-sections were surveyed. Thereby measurement sites along each stream were systematically located.

Statistical Methods and Model Construction

The first stage of analysis was concerned with identifying the number of groups, and with characterizing them through testing the three hypotheses identified in Chapter I. In this stage of analysis, the patterns of spatial variation were analyzed in terms of the identified groups (which in this case are the sandstone group and the shale group) by calculating a regression equation for the number of observations surveyed on each stream.

Plotting all the 65 cross-sections may result in a pattern that shows a significant relationship between the dependent and independent variables. However, when individual streams are plotted a group of lines may result. The position of the lines representative of these streams identifies the groups. In this case, the number of groups may be two groups. One group represents streams flowing on sandstone and the other streams flowing on shale.

In the second stage of analysis, covariate technique is used in the development of the model, to be established, to reveal that channel form represented by the cross-section area as a dependent variable is linked to an independent variable (geology) by another set of variables. The aim of using the regression format, therefore, is to discover the direct and indirect effects of these variables on channel form.

The final product of the dissertation, therefore, is the building of a general model as the sum of the three testable hypotheses introduced in Chapter I. The following is a step-by-step procedure for building such a model.

$$C = a_1 + b_1 \frac{\gamma d S}{V} \quad \ldots\ldots\ldots\ldots \quad (32)$$

$$w/d = a_2 - b_2 M \quad \ldots\ldots\ldots\ldots \quad (33)$$

$$wd = a_3 + b_3 \Sigma L \quad \ldots\ldots\ldots\ldots \quad (34)$$

where,

b_1, b_2, b_3 . . . are the slopes of the regression lines

a_1, a_2, a_3 . . . are the intercepts of the regression lines

γ is a specific weight of water (constant; 62.4 pounds of a cubic foot of clear water)

d is the depth of flow or the depth of the channel assuming a bankfull discharge condition

S is the slope of the channel

M is the weighted mean percent silt and clay in the sediment composing the channel perimeter

ΣL is the total stream length above a cross-section

w/d is the width/depth ratio

wd = C = cross-section area (channel capacity at a cross-section point) and V is the vegetation cover index.

Rearranging equation 33.

$$d = \frac{W}{a_2 - b_2 M} \quad \ldots\ldots\ldots\ldots \quad (35)$$

substitute equation 35 for d in equation 32

$$C = \frac{(a_1 + b_1 \frac{S\gamma}{V})\ W}{a_2 - b_2 M} \quad \ldots\ldots\ldots\ldots \quad (36)$$

dividing equation 34 by d

$$W = \frac{a_3 + b_3 \Sigma L}{d} \quad \ldots\ldots\ldots\ldots \quad (37)$$

substituting equation 37 for W in equation 36

$$C = \frac{(a_1 + b_1 \frac{S\gamma}{V})(a_3 + b_3 \Sigma L)}{d(a_2 - b_2 M)} \quad \ldots\ldots\ldots \quad (38)$$

when a dummy variable (i.e., geology; lithology) is included as a separating factor for the groups of sandstone formation and shale formation, a new format will be defined for equations 32, 33, 34, 35, 36, 37 and 38.

Dummy variable (geology) D_1 = 0 if shale

1 if sandstone

$$C = a_1 + Z_1 D_1 + b_1 \frac{\gamma dS}{V} + Z_2 D_1 \frac{\gamma dS}{V}$$

$$= (a_1 + Z_1 D_1) + (b_1 + Z_2 D_1) \frac{\gamma dS}{V} \quad \ldots\ldots \quad (39)$$

$$w/d = a_2 + Z_3D_1 - b_2M + Z_4D_1M$$

$$= (a_2 + Z_3D_1) + (-b_2 + Z_4D_1)M \quad \ldots\ldots\ldots\ldots \quad (40)$$

$$wd = a_3 + Z_5D_1 + b_3\Sigma L + Z_6D_1\Sigma L$$

$$= (a_3 + Z_5D_1) + (b_3 + Z_6D_1)\Sigma L \quad \ldots\ldots\ldots\ldots \quad (41)$$

Rearranging equation (40)

$$d = \frac{W}{(a_2+Z_3D_1) + (-b_2-Z_4D_1)M} \quad \ldots\ldots\ldots\ldots \quad (42)$$

Dividing equation (41) by d

$$w = \frac{(a_3+Z_5D_1) + (b_3+Z_6D_1)\Sigma L}{d} \quad \ldots\ldots\ldots\ldots \quad (43)$$

substituting (43) in (42) for w

$$d = \frac{(a_3+Z_5D_1) + (b_3+Z_6D_1)\Sigma L}{d[(a_2+Z_3D_1)+(-b_2+Z_4D_1)M]} \quad \ldots\ldots\ldots\ldots \quad (44)$$

substituting (44) in (39) for d

$$C = \frac{(a_1+Z_1D_1)+(b_1+Z_2D_1)\frac{YS}{V}[(a_3+Z_5D_1)+(b_3+Z_6D_1)\Sigma L]}{d[(a_2+Z_3D_1)+(-b_2+Z_4D_1)M]} \quad \ldots\ldots \quad (45)$$

Equation (45) is the general model that this study attempts to establish based on the data and when $D_1 = 1$. If $D_1 = 0$, the model is reduced and becomes equation (38).

Procedure for Testing the Hypotheses

A sample size of 65 observations was obtained. It is assumed to be large enough to detect any small differences, investigate associations, and to map out the main relationship in the study area.

A linear relationship in equations 32, 33, and 34 is considered by using the least squared method. It was hypothesized that the distribution of f(X) for the different values of X in the three distributions are normal. In dealing with a normally distributed population with unknown variance, the t distribution is encountered. Therefore, the same is true in the linear regression model (Winkler, 1975). In equation 32 the t test can be used as:

$$t = \frac{b-B}{\text{est}\sigma_b}$$

and the same can be applied for equations 33 and 34.

In equation 38, the model involves more than one variable; a multiple regression problem. In order to make inferences in this equation, the assumption is that the error term in the linear regression equation is normally distributed with mean zero and variance σ_e^2, where σ_e^2 is not known. Under this assumption, the marginal sampling distributions of individual regression coefficients are t distributions. For a given value of b_i, where i can take on any value from 1 to K variables

$$t = \frac{b_i - B_i}{S_e\sqrt{n/(n-K)}\sqrt{a_{ii}}}$$

Equations 32, 33 and 34 have dependent variables (channel cross-section area and/or width depth ratio), and independent variables for $n_s = 33$ observations for streams on sandstone, and $n_{sh} = 32$ observations for streams on shale. Two questions of interest when analyzing such data are as follows:

1. Is the linear relationship between the dependent variable and the independent variables in equations 32, 33 and 34 the same for sandstone as for shale?

2. Does mean cross-sectional area differ significantly for streams on sandstone from streams on shale?

Question 1 focuses on a comparison between the slopes of regression lines (parallelism) in equations 32, 33, and 34 for streams on sandstone and shale, whereas question 2 focuses on a comparison of the mean cross-section (the intercepts of the regression lines) of streams of the two groups.

The test of parallelism is made before proceeding with analysis of covariance. If this test supports the conclusion that the regression lines (sandstone, and shale groups) were not truly parallel, the question arises as to what should be done about adjustment. It is recommended in this case that no adjustment at all be made to the sample means, since any such adjustment would be misleading. The model describing this general situation is written in equation 45. To determine whether the mean scores for the dummy variable used in equation 45 are significantly different from one another, the following null hypothesis can be tested.

$$H_o : Z_2, Z_4, \text{ and } Z_6 = 0 \text{ (parallelism)}$$

where Z's are constants for the interaction terms in equation 45 between geology and the other independent variables as to take into account their probable effects on predicting channel cross-section area. Subsequently, the approach used in the study is the product of different theories to explain the fluvial systems adjustment, such as dynamic equilibrium, hydraulic geometry, complex responses, thresholds, allometry, and catastrophe theory. Therefore, adapting the three models (equations 32, 33 and 34) can synthesize the findings in mapping the general relationships in area.

CHAPTER VI

TEST AND INTERPRETATION OF THE RESEARCH HYPOTHESES

In Chapter I, three hypotheses were formulated. These hypotheses stated the expected relationship between cross-section areas and/or width-depth ratio, and three controlling factors. The controlling factors are the ratio between tractive force and vegetation resistance index percent silt-clay in channel perimeter, and total stream length above cross-section site. In this chapter, the coefficients of correlations and the coefficients of regression functions will be presented as an indication of relationships between the dependent and independent variables. These statistical results will help substantiate the construction of a general model.

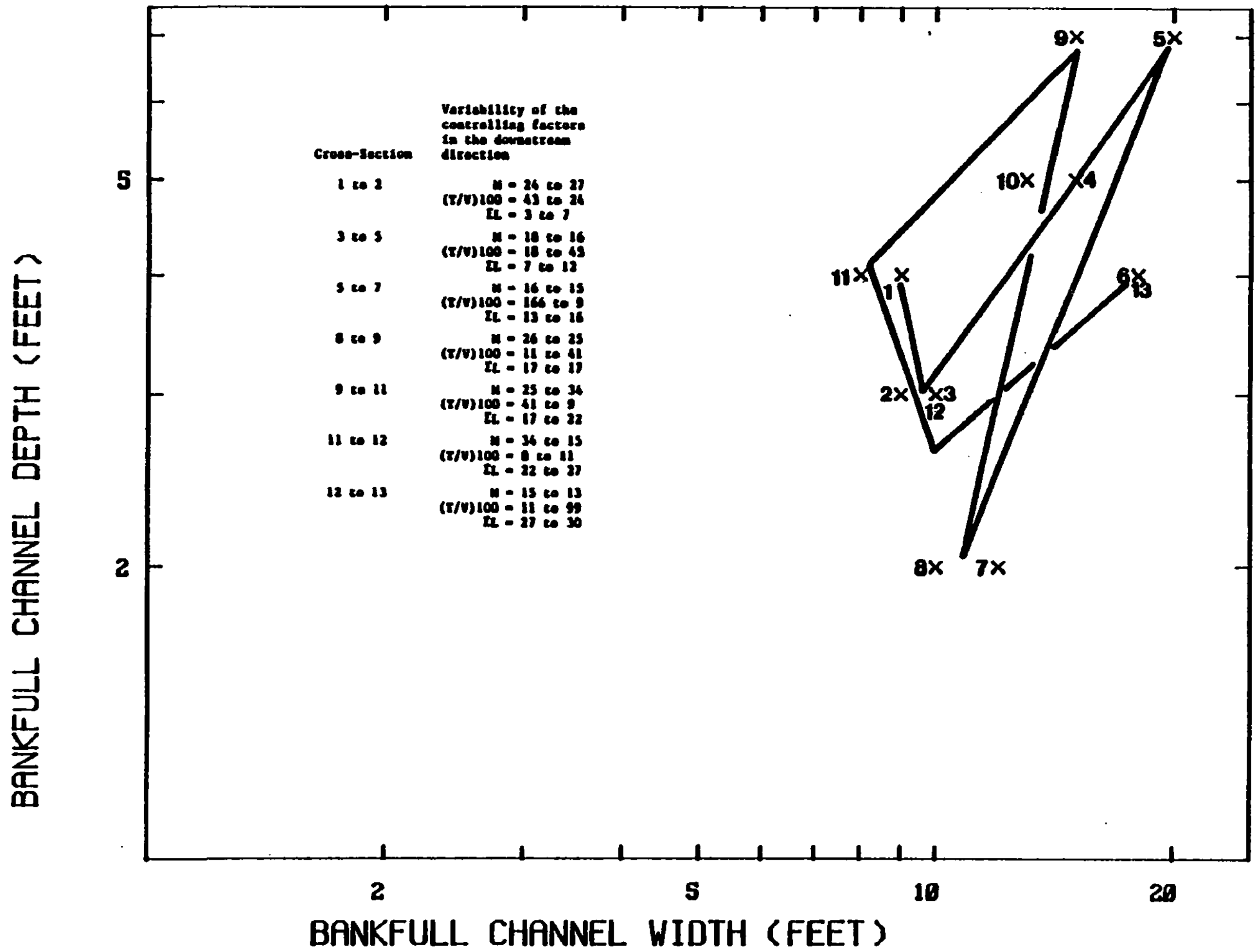
Introduction

Spatial Variation of Channel Form

The two basic dimensions, that always change at different rates, are channel width and depth. Width is controlled by mass movement processes, especially soil fall and slumping, while depth is primarily controlled by fluvial processes on the channel floor. If the rates of change for width and depth are related to a third variable (amount of material excavated, discharge, or tractive force), then the dimensions are related to each other by a power function; width = a $(\text{depth})^b$ (Graf, 1979). The change of channel dimensions in the downstream direction on West Fork and Salt Creek confirm this power function relationship between width and depth (figures 16 and 17). However, the change of channel dimensions in the downstream direction on Kickapoo Creek and Devil's Canyon results in an indecipherable pattern (figures 13 and 14). Note that the lines in figures 13 through 17 were drawn through consecutive cross-section points of the stream system on the diagram (figures 13, 14, and 15).

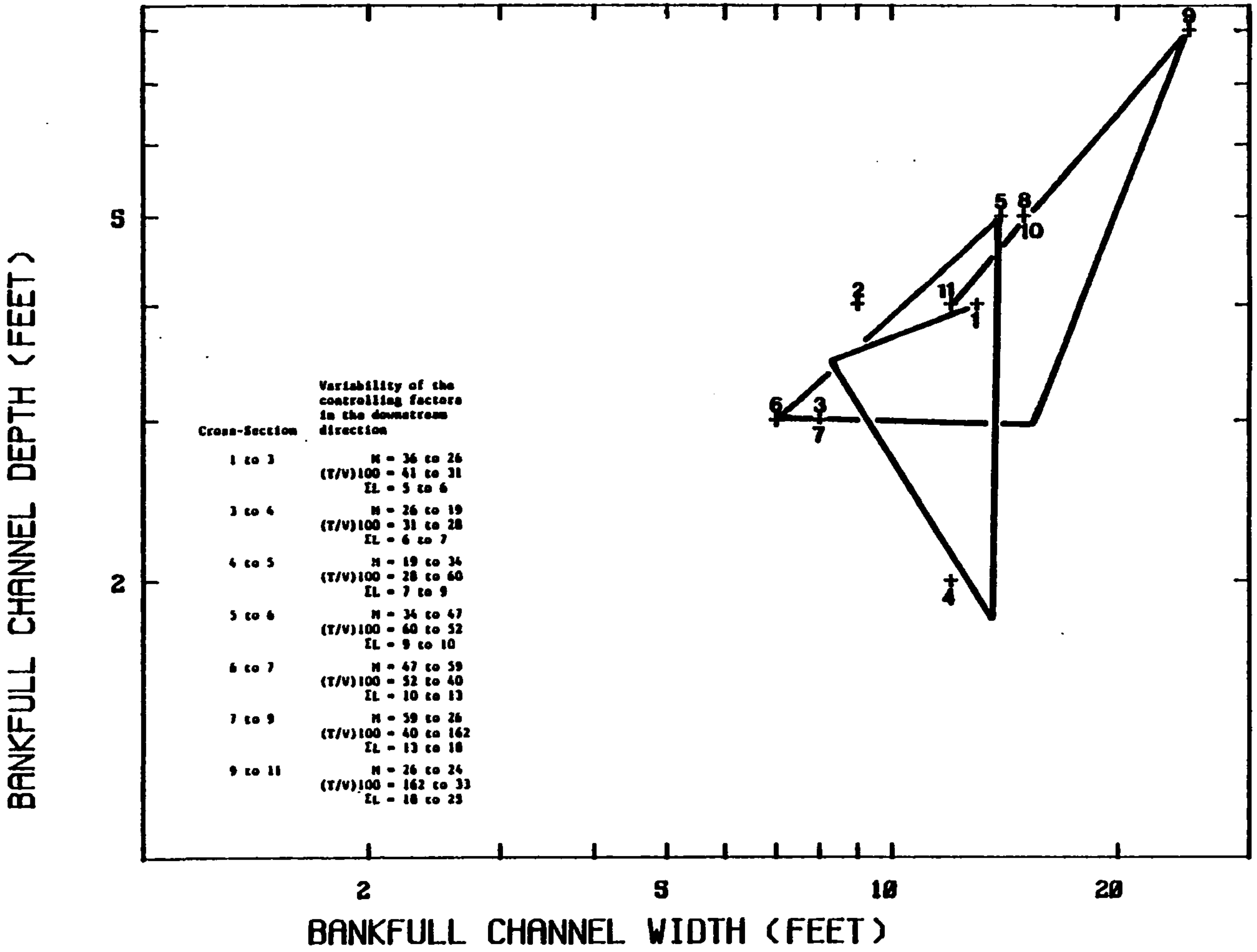
Throughout the literature review in Chapter II, investigators sought to explain areas of erosion and deposition among segments of a stream channel and among streams in a region, as a consequence of 'human' or 'climatic' impact. But few have attempted to explain degradation and aggradation along streams according to environmental variations (i.e., differential geologic formations, unusual flood or rainfall events, differential flows through the drainage network, and inherent disadjustments in stream profiles). The failure of adopting a single approach in the study, to explain the spatial variation of the downstream channel form change, led to adopt three approaches for different streams on different lithology to predict changes in channel form.

In this chapter, three approaches will be examined. Firstly, the relationship between channel form and the force-



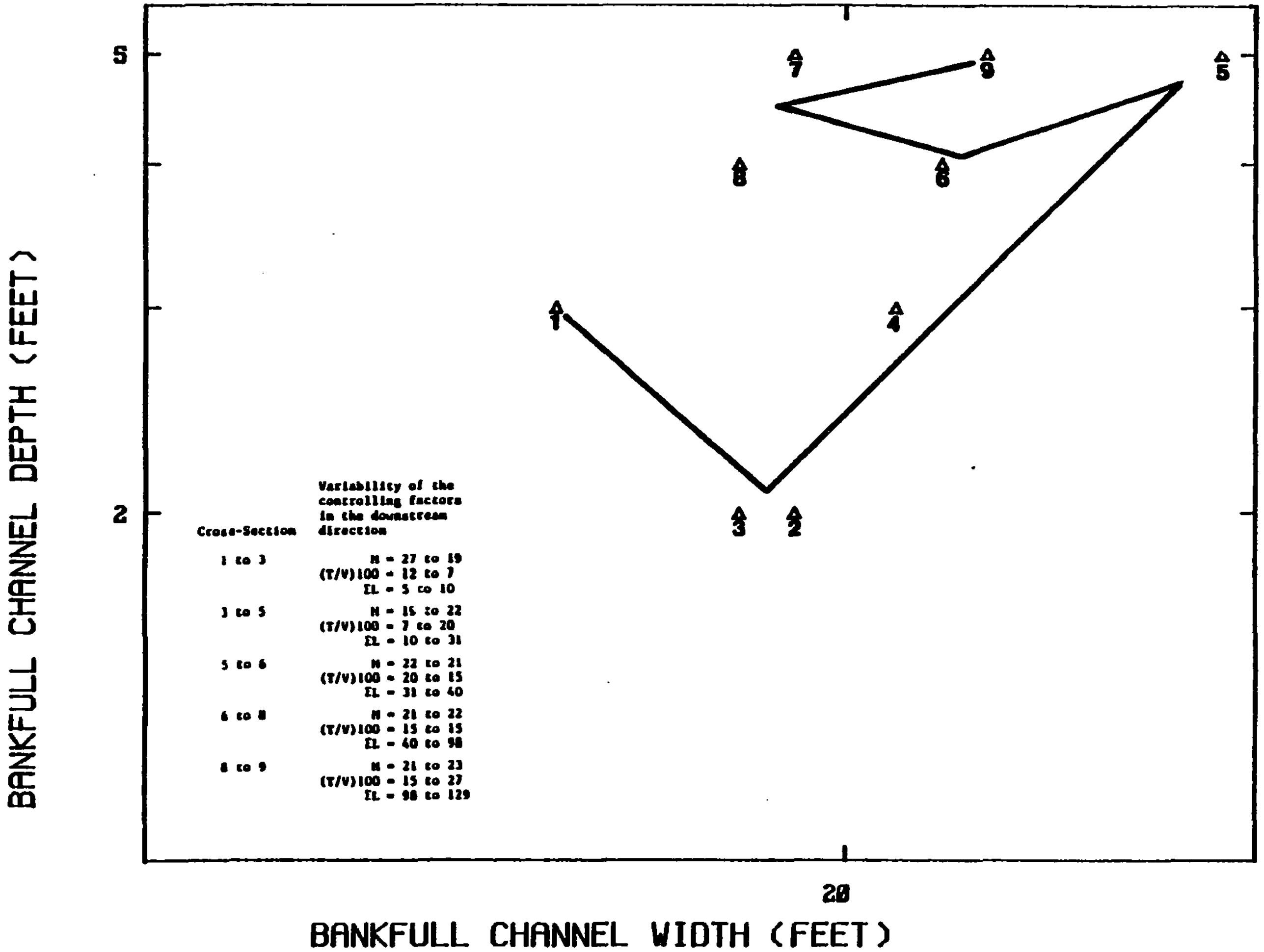
CHANNEL DEPTH VS CHANNEL WIDTH OF KICKAPOO CREEK

Figure 13



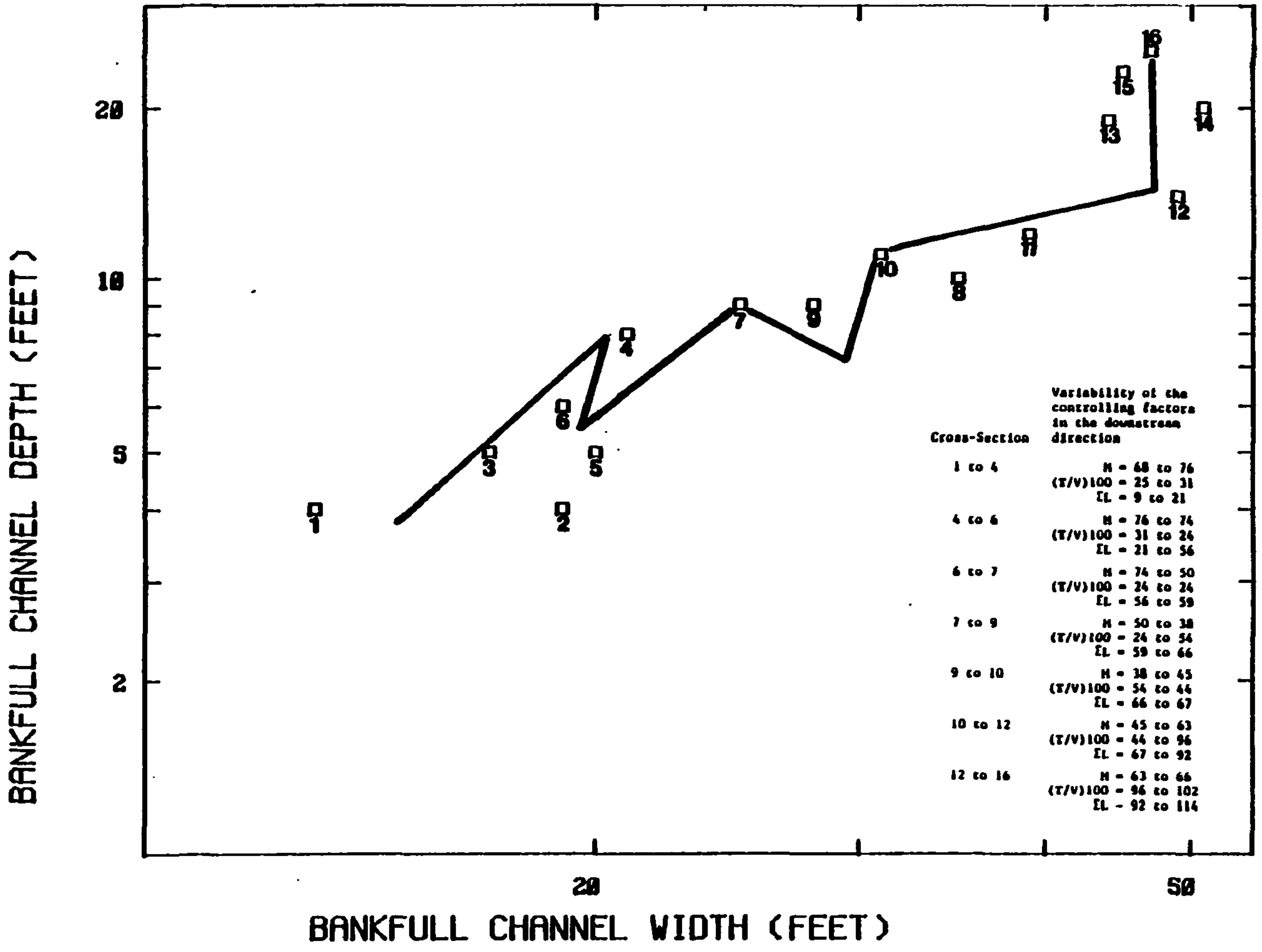
CHANNEL DEPTH VS CHANNEL WIDTH OF DEVILS CANYON CREEK

Figure 14



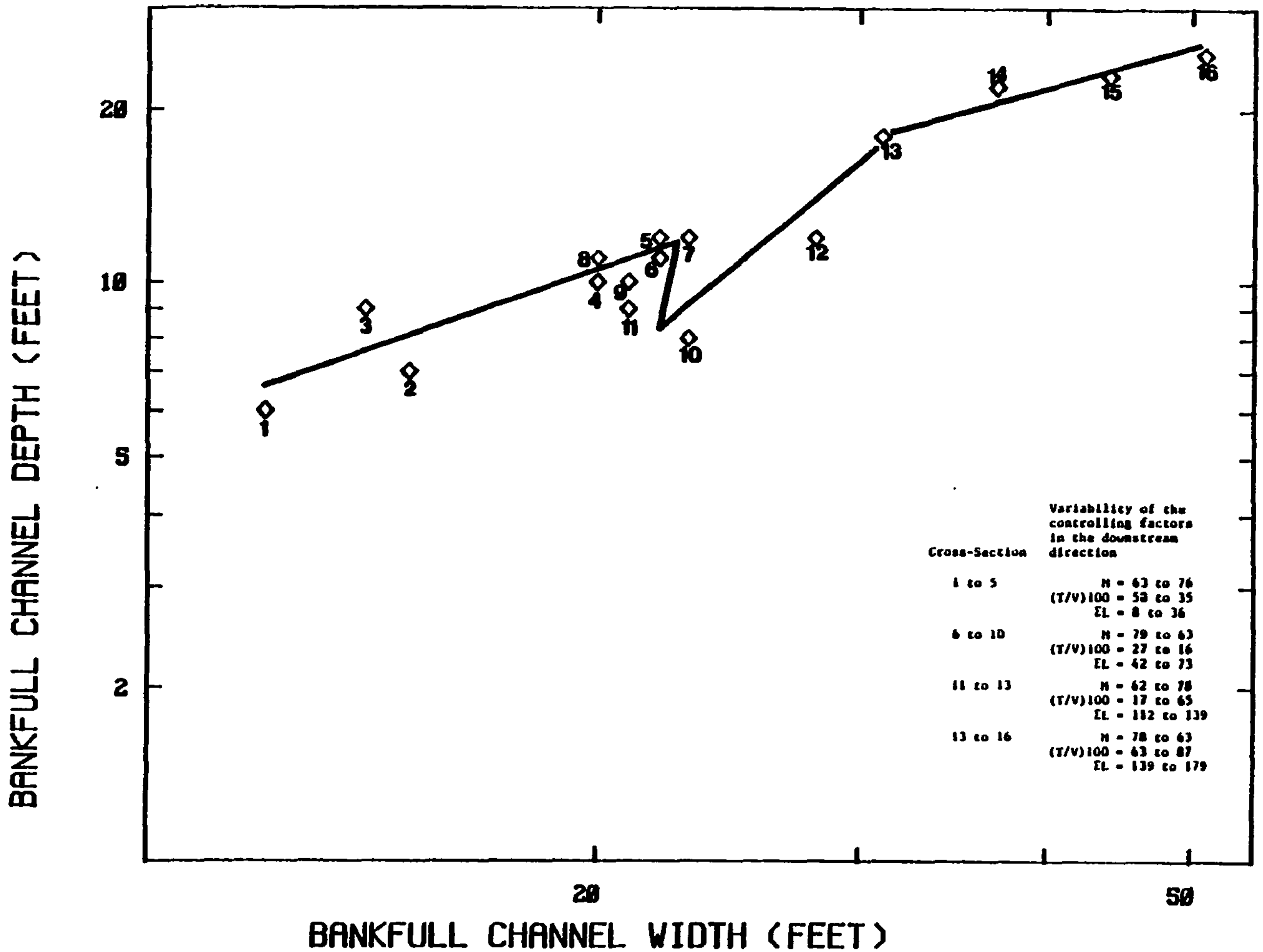
CHANNEL DEPTH VS CHANNEL WIDTH OF BUGGY CREEK

Figure 15



CHANNEL DEPTH VS CHANNEL WIDTH OF WEST FORK

Figure 16



CHANNEL DEPTH VS CHANNEL WIDTH OF SALT CREEK

Figure 17

resistance index is an approach that emphasizes channel form change as the fluvial system adjusts to internal or external stimuli (i.e., changing stream channel slope and/or alteration of vegetation cover). Secondly, the relationship between cross-section area and network volume is a means of characterizing the variations of channel form along individual streams. Thirdly, the relationship between width-depth ratio (cross-sectional form) and sediment characteristics is a means to describe the type of geomorphic thresholds in the system, whether they are thresholds controlled by external factors or are thresholds controlled by internal factors.

Testing the Hypotheses

The Relationship Between Channel Form and Force-Resistance Index

The variability of vegetation cover along the stream channel may determine the rate of increase of the tractive force condition, thereby determining the rate of change of channel form in the downstream direction (Cooke and Reeves, 1976; Graf, 1979a). Vegetation cover in the study area is highly variable and subject to manipulation by human activities. The combined algorithm and comparison between force and resistance demonstrate why channel cutting is sensitive to vegetation changes throughout the basin and results in an increase in force, which might initiate cutting. On the other hand, alteration of vegetation in the valley floor also might lower resistance below the critical value, thereby resulting in channel cutting (Graf, 1979a).

Regression equations for logarithms of vegetation retardance index (V) and tractive force (τ_o) were derived (Table 9, Figure 18).

TABLE 9

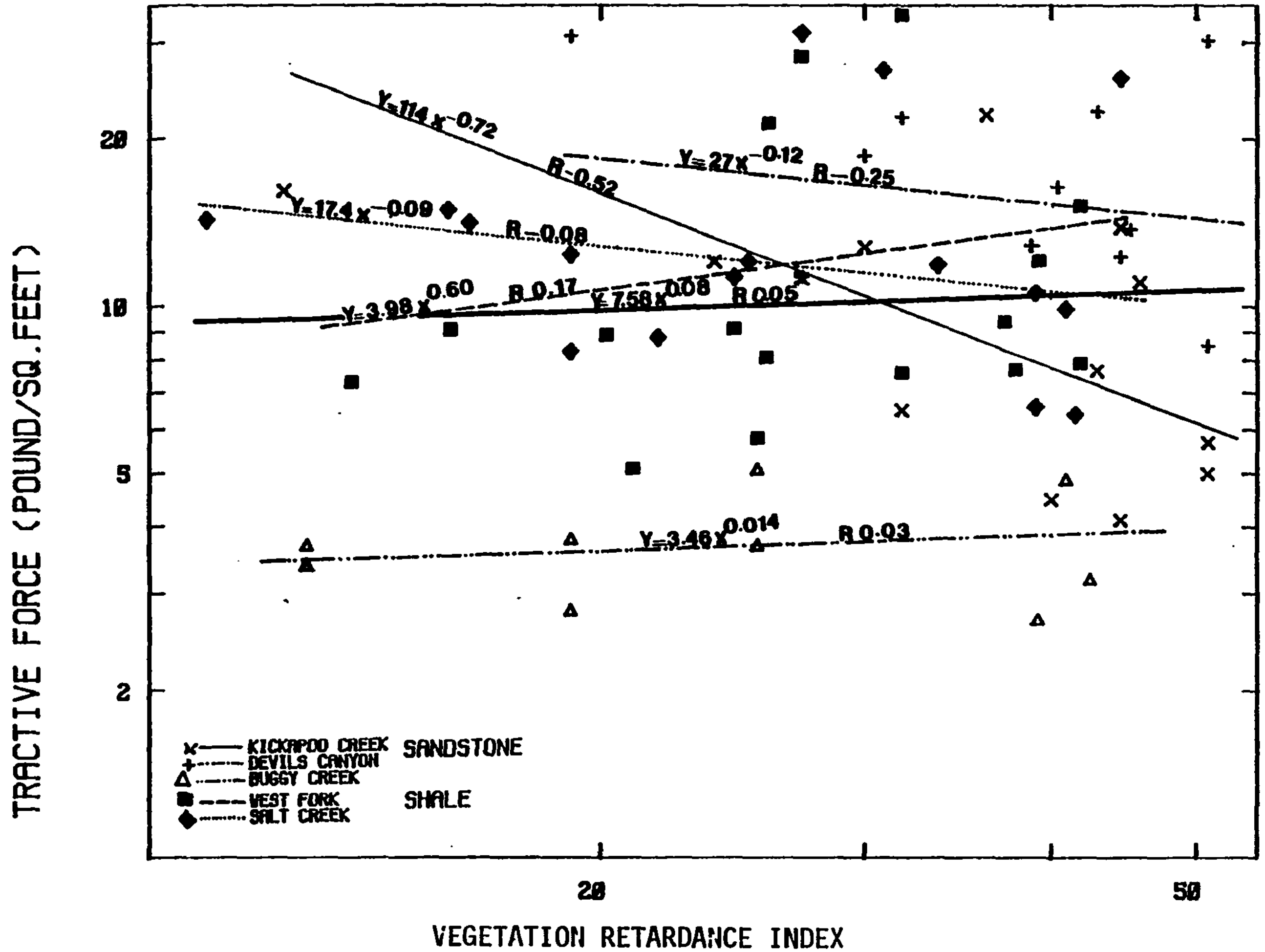
THE RELATIONSHIP BETWEEN TRACTIVE FORCE AND VEGETATION RETARDENCE INDEX

Formation	Stream	Equation	r	R^2
Sandstone	Kickapoo Creek	$\tau_o = 114\ V^{-0.72}$	-0.52*	0.27
	Devil's Canyon	$\tau_o = 27\ V^{-0.12}$	-0.25	0.06
	Buggy Creek	$\tau_o = 3.46\ V^{0.01}$	0.03	0.0009
Shale	West Fork	$\tau_o = 3.98\ V^{0.60}$	0.17	0.029
	Shale	$\tau_o = 17.4\ V^{-0.09}$	-0.08	0.006
Total		$\tau_o = 7.58\ V^{0.08}$	0.05	0.003

*Significant at 0.05 level.

In Kickapoo Creek, the relationship between τ_o and V is significantly different from zero at 0.05 level. It indicates that there is a relationship between vegetation and channels involving a positive feedback loop, whereby decreased vegetation in the valley floor leads to channel cutting, which lowers water tables and in turn accelerates further decline of vegetation (i.e., cross-sections 4 and 5; Figure 15).

The relationship between τ_o and v in the other study streams is not significant and is positive in Buggy Creek and West Fork of Salt Creek. The average value of tractive force in Buggy Creek is low (3.73 pounds/sq. foot) compared with average value in all the streams (12.02 in streams on shale, 8.51 in streams on sandstone, and 10.16



TRACTIVE FORCE VS RESISTANCE

Figure 18

in all the streams). Tractive force conditions along Buggy Creek also do not vary over space because cross-sections 6, 7, and 8 were surveyed in channelized reaches, and cross-sections 2 through 5 were surveyed in reaches where fresh alluvium is deposited in the channel. Also, forest cover is dominant in cross-section 1 (see Appendix 6). The slope of the channel in all Buggy Creek cross-sections is relatively high while average depth of the channel is low (3.66 feet) compared with average depth in other streams (10.47 feet in streams on shale, 3.71 feet in streams on sandstone, and 6.26 feet in all the streams). The low channel depth in Buggy Creek lowered tractive force conditions because the depth of flow at bankfull discharge condition is lowered to a critical value that channel erosion does not prevail (Figure 15). This is evident in high value of width-depth ratio (6.52 average value). On the other hand, tractive force conditions in West Fork Creek increase as bankfull depth increases. Increasing discharge as a function of larger drainage area increases tractive force condition in the downstream direction.

However, the relationships between cross-section area (C) and tractive force (τ_o), and between (C) and vegetation retardance index (V) were calculated in order to determine the direct effect of τ_o and V on channel form. Tables 10 and 11 show the rate of increase of channel cross-section area (C) given a unit of tractive force (τ_o) or one unit of vegetation retardance index (V). (See Figures 19 and 20.)

TABLE 10

THE RELATIONSHIP BETWEEN CHANNEL CAPACITY AND TRACTIVE FORCE CONDITION

Formation	Stream	Equation	r	R^2	F Value
Sandstone	Kickapoo Creek	$C = 7.76\ \tau_o^{0.81}$	0.70*	0.49	10.74
	Devil's Canyon	$C = 3.63\ \tau_o^{0.91}$	0.54	0.29	3.72
	Buggy Creek	$C = 18.62\ \tau_o^{1.01}$	0.43	0.18	1.69
Shale	West Fork Creek	$C = 9.77\ \tau_o^{1.37}$	0.81*	0.66	27.19
	Salt Creek	$C = 741.31\ \tau_o^{-0.39}$	-0.22	0.05	0.75
Total		$C = 28.18\ \tau_o^{0.61}$	0.35	0.12	8.81

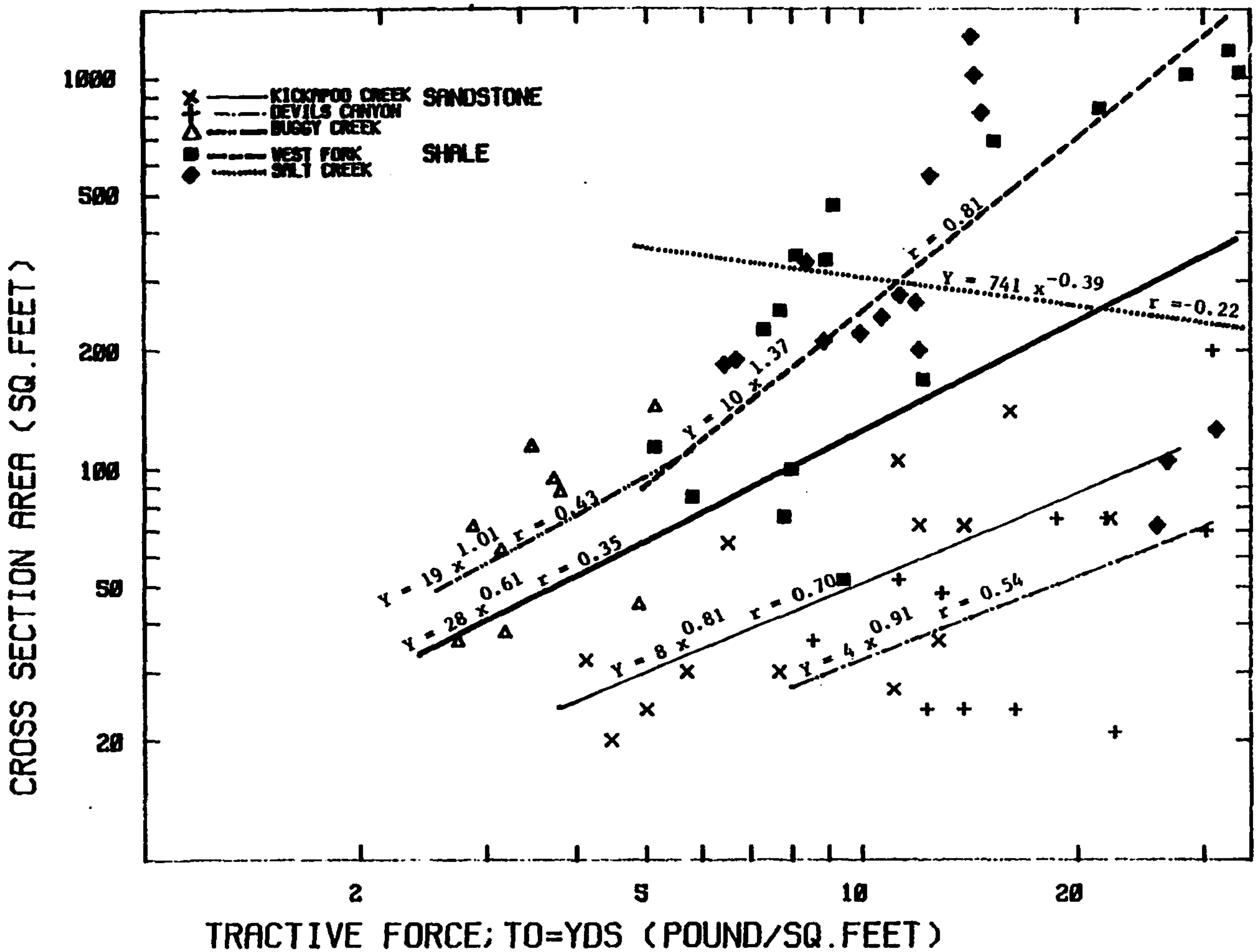
* significant at 0.005

TABLE 11

THE RELATIONSHIP BETWEEN CHANNEL CAPACITY AND PERCENT RIPARIAN VEGETATION COVER

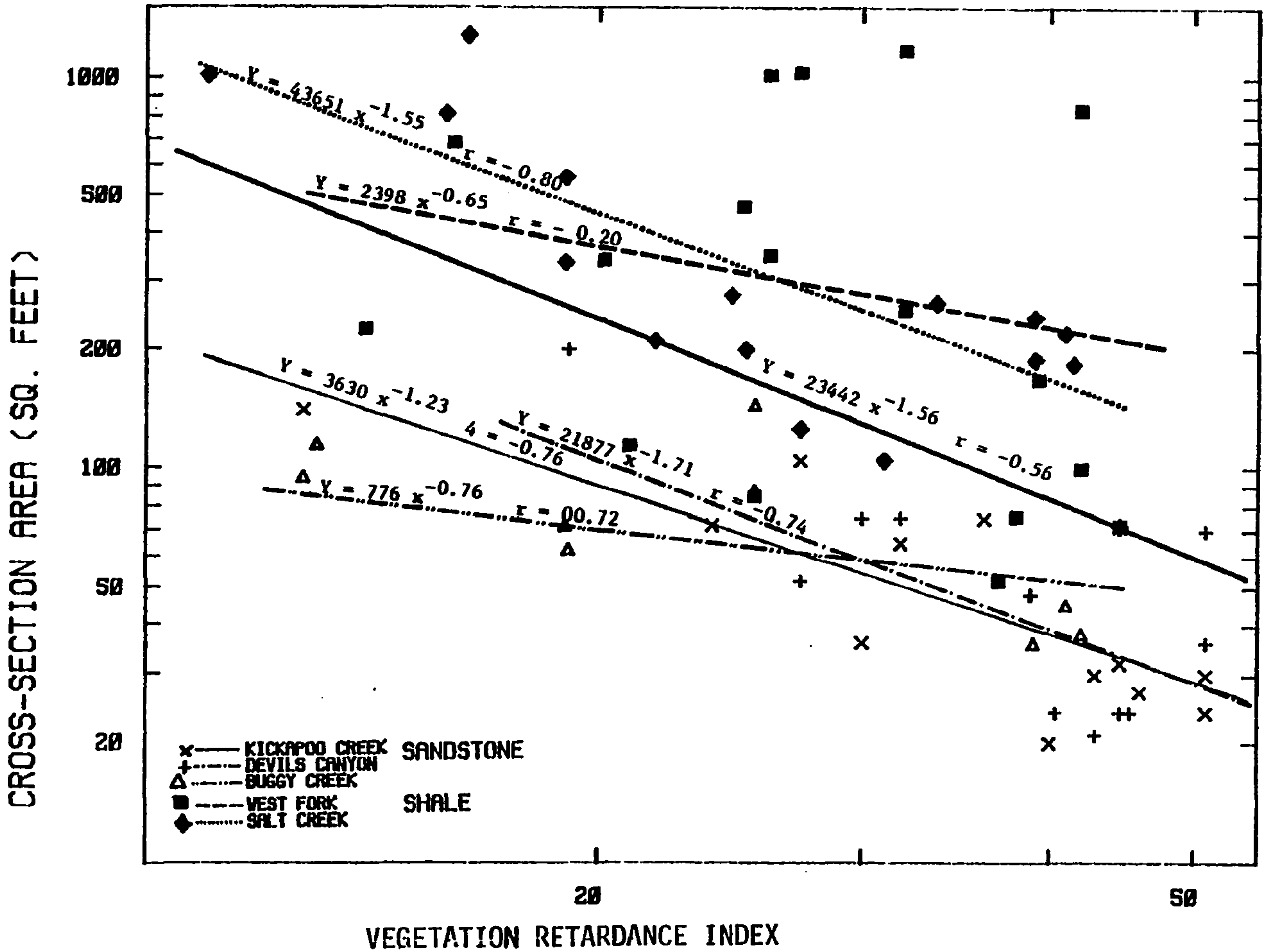
Formation	Stream	Equation	r	R^2	F Value
Sandstone	Kickapoo Creek	$C = 3630\ V^{-1.22}$	-0.76*	0.58	15.53
	Devil's Canyon	$C = 21877\ V^{-1.71}$	-0.75*	0.56	11.27
	Buggy Creek	$C = 776\ V^{-0.76}$	-0.72*	0.52	7.79
Shale	West Fork Creek	$C = 2398\ V^{-0.65}$	-0.21	0.04	0.67
	Salt Creek	$C = 43651\ V^{-1.55}$	-0.81*	0.66	26.24
Total		$C = 23442\ V^{-1.56}$	-0.55	0.30	27.73

*significant at 0.005 level



CHANNEL CAPACITY VS TRACTIVE FORCE

Figure 19



CHANNEL CAPACITY VS PERCENT VEGETATION COVER

Figure 20

The relationships between (C) and τ_o or V (Tables 10 and 11, Figures 19 and 20) are not significant at 0.05 level for all the observations which represent all the study streams. However, the relationships in individual streams are significant at 0.005 significant level. These results will substantiate the need for an index of force exerted by flow and resistance offered by vegetation. An index of force-resistance was calculated by obtaining the ratio between tractive force and vegetation retardance index. This ratio may indicate that when vegetation cover along the channel is dense, channel cutting does not occur. On the other hand, if tractive force conditions exerted by flow in the channel exceed the resistance offered by vegetation, channel deepening occurs. In the later case, channel banks tend to stablize because of the establishment of vegetation by deep rooted trees (e.g., West Fork Creek). Also, the force-resistance index is a good indicator of human landuse activities along the valley floor and its impact on stream channel form (e.g., the relationship between τ_o and V in Kickapoo Creek is r -0.52 which is significant at 0.005 level).

It was hypothesized that the relationship between channel capacity (C) and force-resistance index (TV) should be positively significant. Table 10 shows the rate of increase of channel capacity (C) given one unit of force-resistance index (TV) in each stream (Figure 21).

TABLE 12

THE RELATIONSHIP BETWEEN CHANNEL CAPACITY AND FORCE-RESISTANCE INDEX

Formation	Stream	Equation	r	R^2	F Value
Sandstone	Kickapoo Creek	$C = 7.2\ TV^{0.56}$	0.83*	0.69	22.45
	Devil's Canyon	$C = 1.69\ TV^{0.87}$	0.75*	0.56	12.79
	Buggy Creek	$C = 7.8\ TV^{0.81}$	0.85*	0.72	14.45
Shale	West Fork	$C = 1.4\ TV^{1.4}$	0.87*	0.76	45.63
	Salt Creek	$C = 50\ TV^{0.44}$	0.35	0.12	1.95
Total		$C = 5.58\ TV^{0.86}$	0.58*	0.34	28.60

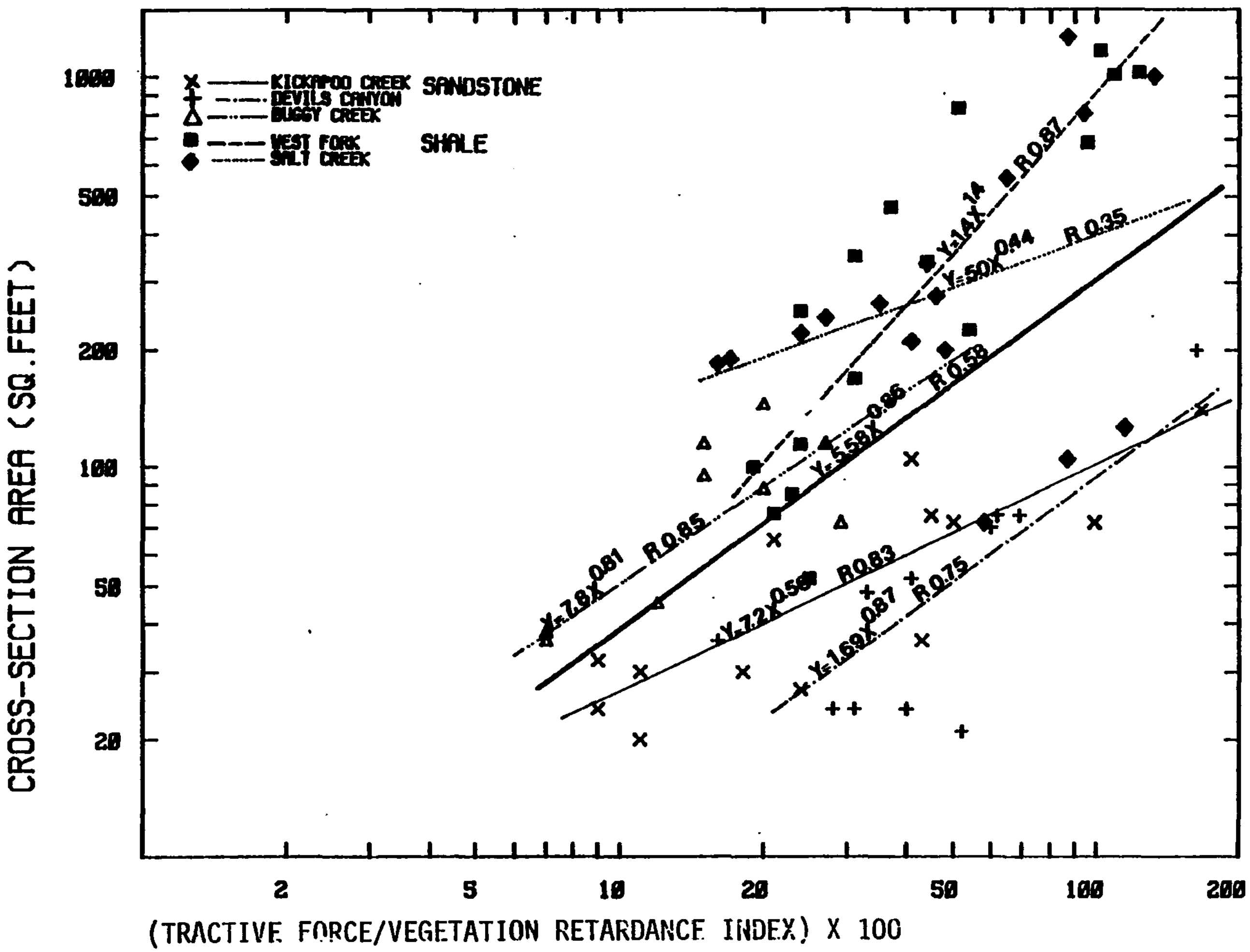
*significant at 0.005 level

The significance test of regression coefficients were derived by comparing the statistical t and t_{df}, 0.995 is shown in Table 13.

TABLE 13

RESULTS OF TESTING THE HYPOTHESIS OF THE RELATIONSHIP BETWEEN CHANNEL CAPACITY AND FORCE-RESISTANCE INDEX

Stream	Calculated t	Hypothesis		d_f	t_{df}, 0.995	
Kickapoo Creek	3.50	$H_o = \hat{\beta} = 0$;	$H_A = \hat{\beta} > 0$	12	3.05	reject H_o
Devil's Canyon	4.35	"	"	10	3.16	"
Buggy Creek	6.23	"	"	8	3.25	"
West Fork	6.36	"	"	15	2.94	"
Salt Creek	1.23	"	"	15	2.94	accept H_o
Total	3.14	"	"	64	2.66	reject H_o



CHANNEL CAPACITY VS(FORCE/RESISTANCE)×100

Figure 21

The slope of Salt Creek regression line is not significant at 0.005 level. Extreme cases of increased vegetal retardance index while headcut by stream, occur in the upper part of the basin, influenced the correlation and regression coefficients between C and TV. The formation composing this part of the basin is interbedded with silty material which provides low cohesiveness for the shaly Dog Creek Formation. Nevertheless, the position of the regression lines for the individual stream (Figure 21) is forming two groups of regression lines. Kickapoo Creek and Devil's Canyon Creek (sandstone) regression lines occupying the lower side of Figure 21, while West Fork Creek and Salt Creek (shale) regression lines occupying the upper part of the graph. Therefore, two groups of regression lines can be separated.

a. Comparing the two groups' regression lines slopes:

$H_O \hat{\beta}_1 = \hat{\beta}_2$ (the two groups regression lines are parallel)

$H_A \hat{\beta}_1 \neq \hat{\beta}_2$

where $\hat{\beta}_1$ is the sum of regression lines slopes for streams on sandstone,

and $\hat{\beta}_2$ is the sum of regression lines slopes for streams on shale.

$$t = \frac{\hat{\beta}_1 - \hat{\beta}_2}{S_{\hat{\beta}_1 - \hat{\beta}_2}}$$

where $S_{\hat{\beta}_1 - \hat{\beta}_2} = \sqrt{S^2_{P}, y/x \left[\frac{1}{(n_1-1)S^2_{x_1}} + \frac{1}{(n_2-1)S^2_{x_2}}\right]}$

$$\text{and } S^2_{p}, \; y/x = \frac{(n_1-2)S^2_{x_1} + (n_2-2)S^2_{x_2}}{(n_1-2) + (n_2-2)}$$

$$t = \frac{1.73}{0.2559} = 6.76$$

$$t_{60}, \; 0.995 = 2.66$$

Therefore, the H_A is accepted at 0.005 significant level, and there is a strong evidence the two groups regression lines are not parallel.

b. Comparing the two groups' regression lines intercepts.

$H_o \beta_1 = \beta_2$

$H_A \beta_1 \neq \beta_2$

$$t = \frac{-34.71}{0.2559} = -135.63$$

$$t_{60}, \; 0.995 = 2.66$$

Therefore, the H_A is accepted at 0.005 significant level, and there is a strong evidence the two intercepts of the two groups are not the same.

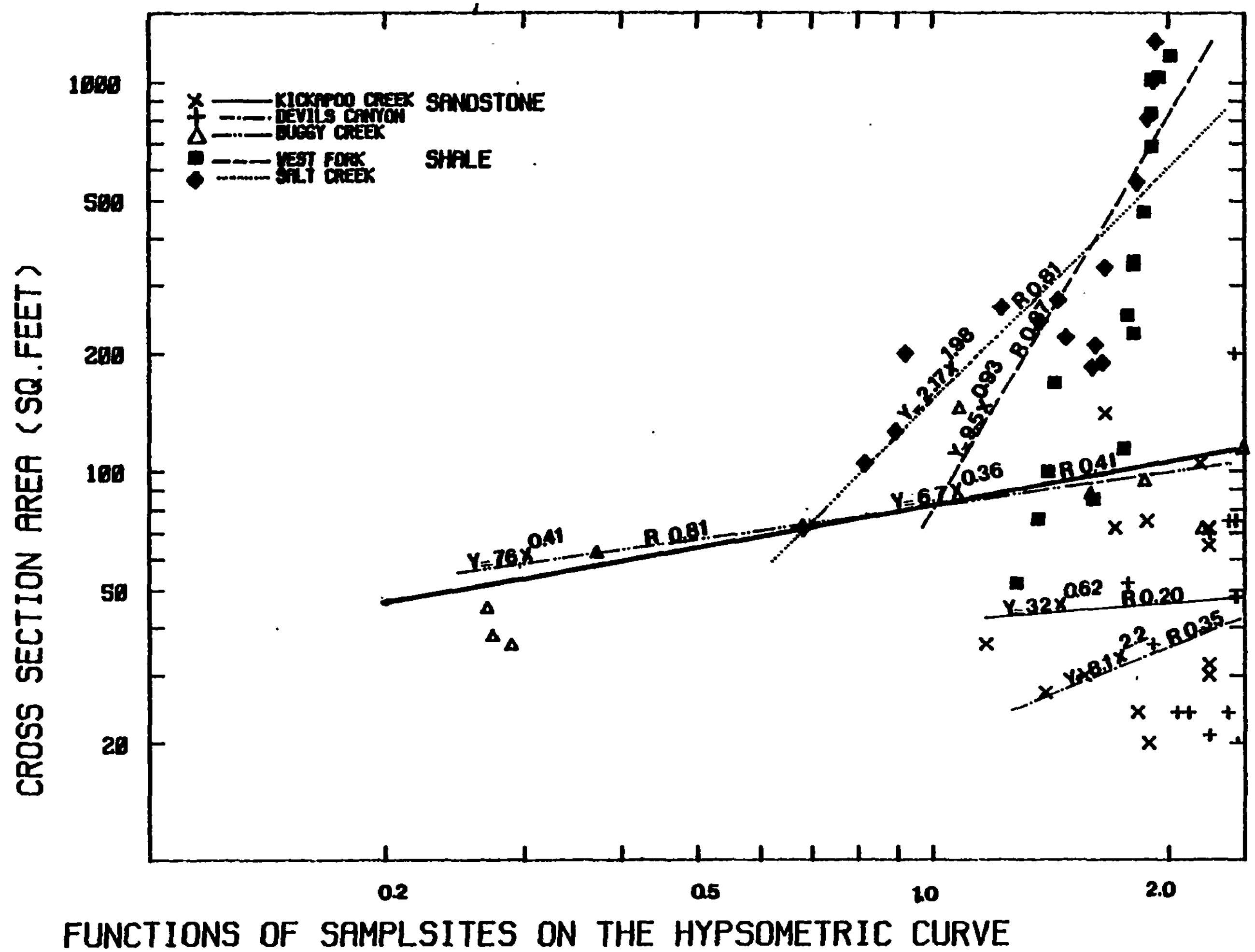
The average cross-section area for streams on sandstone is 51 square feet, while the average cross-section area for streams on shale is 275 square feet. The differences in the two means may be the function of bank-full discharge condition which streams on shale have a higher volume of discharge than streams on sandstone. On sandstone various valley floor changes introduced by man influenced the stream channel form where some cross-sections were entrenched as a result of overgrazing while other cross-sections were filled because of the increasing vegetation cover. Channel entrenchment on sandstone is affected by the erosiveness of flows independently of increasing discharge (i.e., increasing stream channel gradient, thereby, increased tractive force condition in the channel). Also, human activities act to increase the erodibility of valley-floor sediment (i.e., removal of vegetation cover lowers the resistance offered by vegetation).

The nature of Kickapoo Creek and Devil's Canyon valley floor vegetation is not well recorded, but it is certainly variable. This is evident in cultivated fields. When the cultivated fields remain relatively vulnerable to erosion throughout most of the rainy season, valley-floor discharge accelerates channel cutting. Graf (1979a) hypothesized in his study of the development of Montane arroyos and gullies that there is a great effect of biomass of vegetation on tractive force condition in the channel. It was found true in Kickapoo Creek and Devil's Canyon (F values are 22.45 and 12.79 respectively and they are significant at 0.005 level). On the other hand, Graf's hypothesis is difficult to prove in streams on shale. Increasing discharge in the downstream direction in West Fork Stream, which increases depth of flow, therefore, tractive force condition is increased. In this case, vegetation in the channel bottom cannot withstand the erosiveness of flow. The rate of change of channel depth in the downstream direction, therefore, is greater than channel width.

The Relationship Between Channel Form and Network Volume of the Drainage Basin

A greater understanding of the relationships between channel form and drainage basin characteristics is required in fluvial geomorphology studies. Hypsometric curves were constructed for the drainage basins included in the study to express the overall slope of the basin above the cross-section site. The hypsometric curve is an expression of relative basin height to relative basin area, and the discrete points employed in obtaining these curves relate to contour lines within the watershed. Total potential energy derived from the hypsometric integral is a useful parameter. Converting the energy potential to discharge at the mouth of the basin will require the identification of other basin parameters as the hypsometric integral is not unique to any particular basin; there may exist a multitude of hypsometric curve forms for any one specific integral (Harlin, 1978, p. 180).

The form of the hypsometric curve demonstrates how basin mass is distributed from the mouth of the basin to the top. The amount of headward erosion attained by the stream within the basin can, under certain circumstances, be monitored by hypsometric skewness, but much more critical to hydrophysical



CHANNEL CAPACITY VS HYPSOMETRIC FUNCTIONS

Figure 22

research are moments and centers of gravity along the curve's density function. The derivative of the cumulative curve indicates how much, and where watershed slope change is occurring (Harlin, 1982).

The interpretation of the hypsometric curves for the study streams (see Appendix 2) reveals that negative skewness characterize the curves of basins on sandstone. Headward development of the mainstream and associated tributaries is the early stage of shifting mass within the basin as well as remove material through the basin mouth. Correlation coefficients between channel capacity and hypsometric functions (Figure 22) are not significant (Kickapoo Creek, r = 0.20; Devil's Canyon, r = 0.35). In contrast, the positive skewness of the hypsometric curves for the shale basins shows a headward development as well as a decrease in the mid-basin slope. Streams are efficient and associated with an accelerated increase of the channel cross-section area in the downstream direction. Correlation coefficients between channel capacity and hypsometric function (West Fork, r = 0.88; Salt Creek, r = 0.82) are significant at the 0.005 level). Buggy Creek (sandstone) shows a strong correlation between channel capacity and hypsometric function (r = 0.81), while the regression coefficients indicate a moderate rate of change of channel capacity given one unit of hypsometric value ($Y = 75x^{0.81}$). The variability of channel cross-section area of Buggy Creek is relatively low along the stream channel. Local change of channel form is a result of human intervention. Cross-sections 5 through 8 in Buggy Creek have been channelized and the channel banks are alleviated. Although the energy slopes of Kickapoo Creek and Devil's Canyon are high, the relationship between channel capacity and hypsometric function is not significant at 0.05 level. Floodwater retarding structures built on these streams have decreased the potential of generating erosive discharges, therefore channel capacity has not been increased. Fluctuation of the water table along the stream has caused local cut and fill of the stream channel.

TABLE 14

THE RELATIONSHIP BETWEEN CHANNEL CAPACITY (CROSS-SECTION AREA) AND HYPSOMETRIC FUNCTION

Formation	Stream	Equation	r	R^2	F Value
Sandstone Formation	Kickapoo Creek	$C = 32\ H^{0.62}$	0.20	0.04	0.49
	Devil's Canyon	$C = 8.12\ H^{2.19}$	0.35	0.12	1.22
	Buggy Creek	$C = 75\ H^{0.81}$	0.81*	0.66	13.75
Shale Formation	West Fork	$C = 9.5\ H^{0.93}$	0.87*	0.76	46.97
	Salt Creek	$C = 2.17\ H^{1.98}$	0.81*	0.66	28.08
Total		$C = 6.7\ H^{0.36}$	0.41	0.17	1.70

*significant at 0.005 level

The basins used in the study ranges in size from 12 to 50 square miles. As drainage basin area increases, discharge increases and then channel capacity increases (Park, 1978) despite the fact that large basins always become less efficient in translating total rainfall input to discharge (Harlin, 1982b). The data of channel capacity of large basins indicates that channel capacity increases dramatically given one unit of hypsometric function (Figure 22). In addition, the data for streams on sandstone agree with Schumm's findings (1956) of a strong linear relationship between the hypsometric integral and relief ratio and stream gradient, where, as stream gradient increases, stream width-depth ratio increases.

Harlin (1982b) stated

> Drainage density is an important parameter relating hydrograph characteristics to geomorphology, however, channel widening in the central reaches of the watershed as Strahler basin order increases and deposition of upland material are often concomitant with increasing drainage density. This mid-basin response will produce negative feedback in the rainfall runoff system by increasing surface storage and increasing lagtime. Drainage density thus becomes region specific and does not transfer well across changing climatic and geologic environments. Combining stream density and mid-basin slope (hypsometric density Kurtosis) projects a geomorphic parameter sensitive to stream flow and overland flow.

It was desirable to define an independent variable that provides a surrogate for discharge and a measure of spatial variation. A volumetric index of stream network was chosen particularly because of its applicability on drainage basins impounded by flood-water structures. Also, the index is applicable to basin streams where channel cross-section area increases relative drainage area.

The linear regression between the logarithms of drainage density and total stream length above the cross-section area were employed as a basis for the comparison of the two group networks, i.e., the sandstone group and the shale group (Figure 23) (Gregory and Gardner, 1975). The two groups regression relationships indicate that the rate of change of drainage density (Dd) with stream length above the cross-section area (ΣL) is greater in sandstone basins than in shale basin (Table 15).

TABLE 15

THE RELATIONSHIP BETWEEN DRAINAGE DENSITY AND TOTAL STREAM LENGTH ABOVE THE CROSS-SECTION AREA

Formation	Stream	Equation	r	R^2
Sandstone	Kickapoo Creek	$Dd = 1.44\ \Sigma L^{0.21}$	0.66*	0.44
	Devil's Canyon	$Dd = 1.02\ \Sigma L^{0.22}$	0.49*	0.24
	Buggy Creek	$Dd = 1.43\ \Sigma L^{0.09}$	0.62*	0.38
Shale	West Fork	$Dd = 4.27\ \Sigma L^{-0.02}$	-0.07	0.005
	Salt Creek	$Dd = 3.22\ \Sigma L^{0.02}$	0.16	0.03

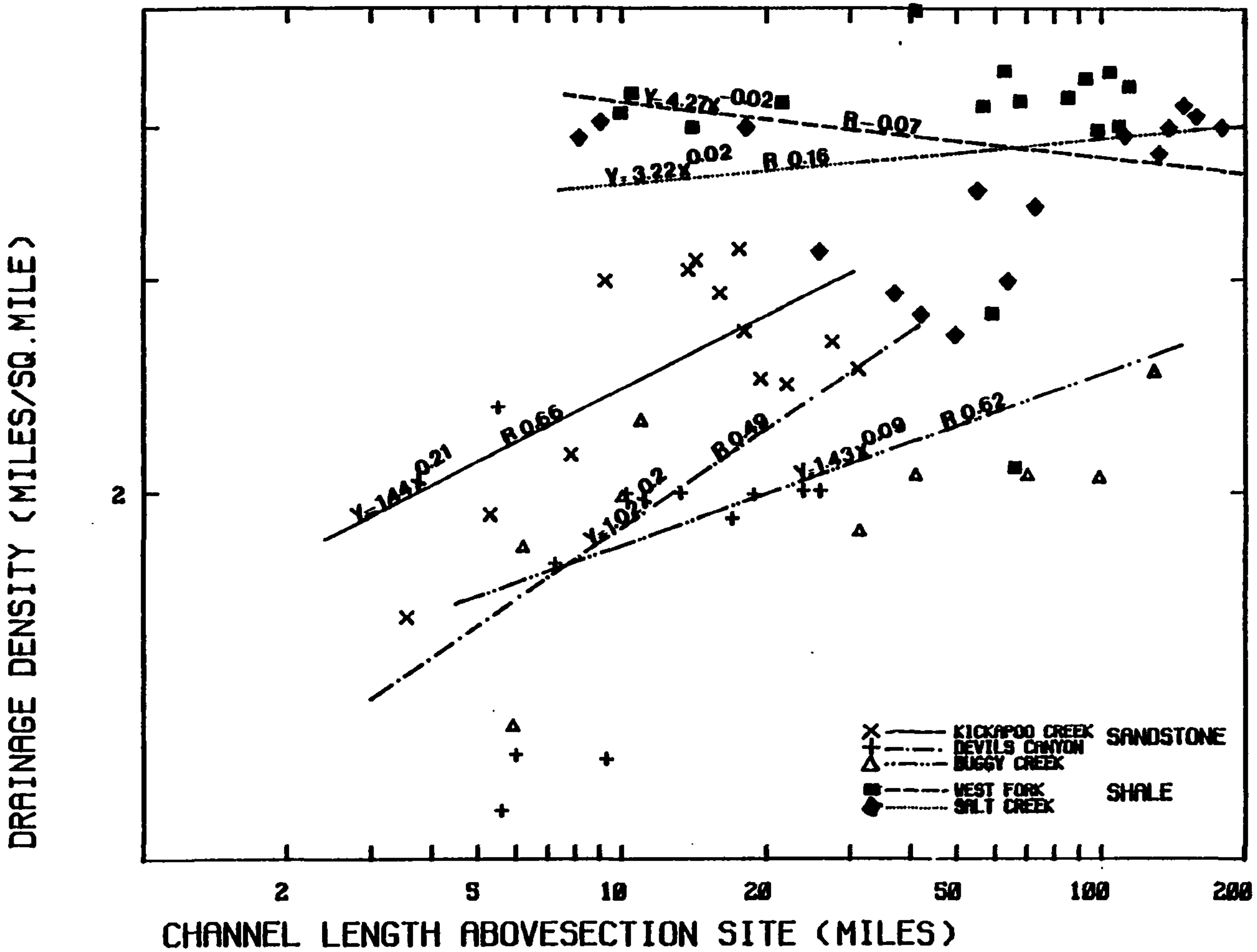
*significant at 0.05 level

a - Comparing the two grouped (sandstone group, and shale group) regression slopes; are the regression lines for the two groups parallel?

$H_O\ \hat{\beta}_1 = \hat{\beta}_2$ (the two group regression lines are parallel)

$H_A\ \hat{\beta}_1 > \hat{\beta}_2$

where β_1 is the sum of regression lines slope values for streams on sandstone,



DRAINAGE DENSITY VS CHANNEL LENGTH

Figure 23

and β_2 is the sum of regression lines slope values for streams on shale.

$$t = \frac{\hat{\beta}_1 - \hat{\beta}_2}{S_{\hat{\beta}_1 - \hat{\beta}_2}} = \frac{(0.22+0.21+0.09)-(-0.02+0.02)}{S_{\hat{\beta}_1 - \hat{\beta}_2}} = \frac{0.52 - 0}{S_{\hat{\beta}_1 - \hat{\beta}_2}}$$

where $S_{\hat{\beta}_1 - \hat{\beta}_2} = \sqrt{S^2_{P,Y/X}\left[\frac{1}{(n_1-1)S^2_{x_1}} + \frac{1}{(n_2-1)S^2_{x_2}}\right]}$

and $S^2_{P,Y/X} = \frac{(n_1-2)S^2_{x_1}+(n_2-2)S^2_{x_2}}{(n_1-2)+(n_2-2)}$

$t = \frac{0.52}{0.12} = 4.33$

$t_{60}, 0.995 = 2.66$

Therefore, the H_o is rejected at 0.005 significant level, and there is a strong evidence the two groups regression lines are not parallel.

b - Comparing the two groups regression lines intercepts:

$H_o\ \beta_1 = \beta_2$

$H_A\ \beta_1 < \beta_2$

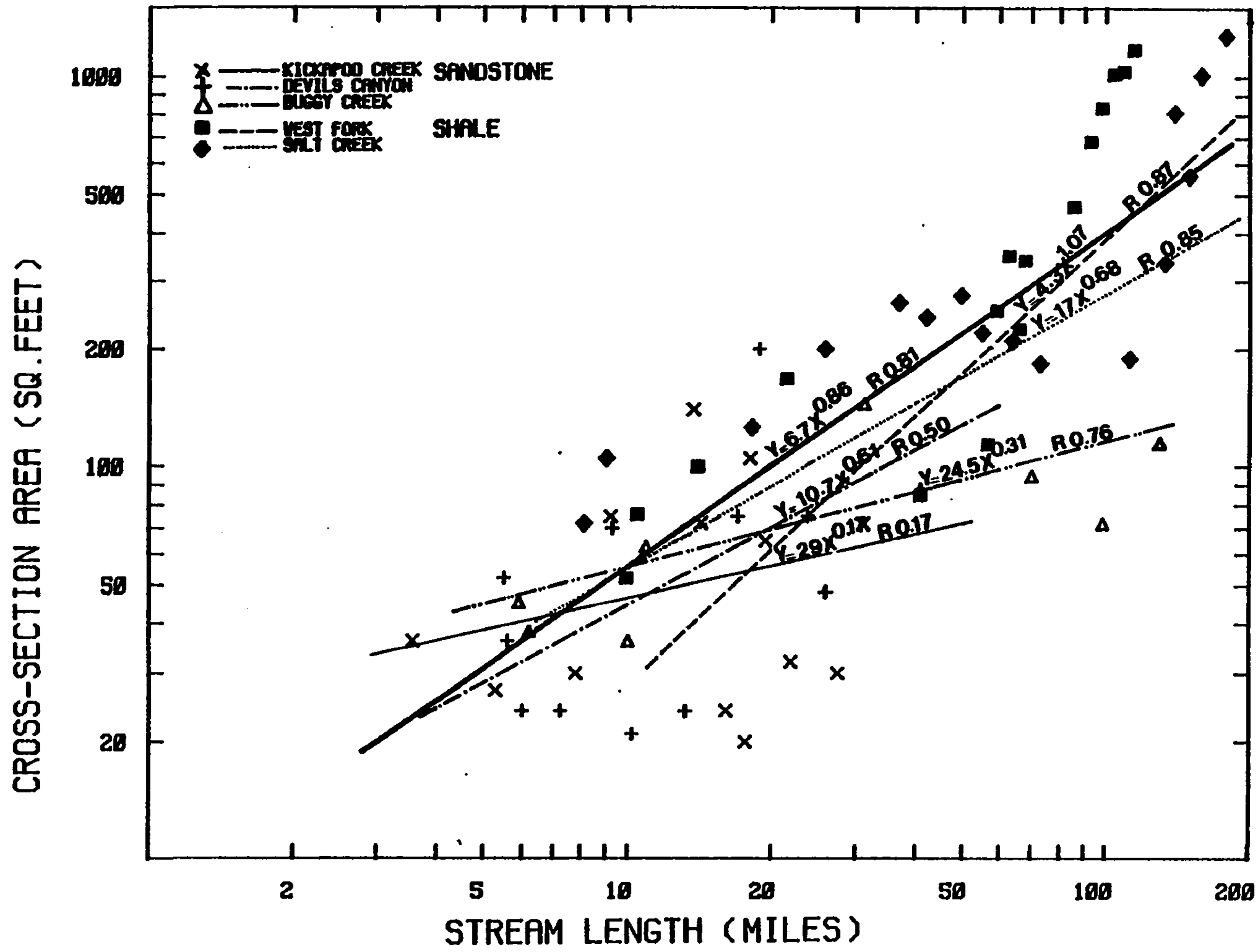
$t = \frac{3.89 - 7.49}{0.121} = -29.75$

$t_{60}, 0.995 = 2.66$

Therefore, the H_o is rejected at 0.005 significant level, and there is a strong evidence the two groups regression lines intercepts are not the same.

Drainage density in the sandstone basins was found to be higher than in basins on shale as a result of the large area in sandstone basins converted from rangeland to cropland (e.g., removal of vegetation increases rills and gully cutting). The sandstone is a highly erodible material once vegetation cover is removed. The shale, on the other hand, is more resistant to erosion by headward cutting by streams. However, as indicated above, a comparison of drainage densities is a purely spatial comparison and one which ignores the volume of the channels constituting the network. Regression relationships were then established between the channel capacity (C) and total stream length (ΣL) so that a volumetric comparison could be made between the two groups of streams (sandstone group and shale group). For each basin in the two groups a regression equation (Figure 24) was derived. These regression equations are included in Table 16.

The values of total stream length for each drainage basin provided the upper limits over which the regression equations could be integrated. The integration is illustrated for one of the five calculations. The five volumes calculated are included in Table 16. For West Fork basin the regression equation for the bankfull stage was $C_{Bf} = 4.3\Sigma L^{1.07}$ and this equation has to be integrated between limits



CHANNEL CAPACITY VS STREAM LENGTH ABOVE THE CROSS-SECTION AREA

Figure 24

TABLE 16

THE RELATIONSHIP BETWEEN CHANNEL CAPACITY AND TOTAL STREAM LENGTH ABOVE THE CROSS-SECTION AREA

Formation	Stream	Function	r	R^2	F
Sandstone	Kickapoo Creek	$C = 29\ \Sigma L^{0.17}$	0.17	0.03	00.36
	Devil's Canyon	$C = 10.7\ \Sigma L^{0.61}$	0.50	0.25	03.11
	Buggy Creek	$C = 24.5\ \Sigma L^{0.31}$	0.76*	0.58	10.06
Shale	West Fork	$C = 4.3\ \Sigma L^{1.07}$	0.87*	0.76	43.99
	Salt Creek	$C = 17\ \Sigma L^{0.68}$	0.85*	0.72	36.94
Total		$C = 7.2\ \Sigma L^{0.82}$	0.81*	0.66	119.56

*significant at 0.005 level

$\Sigma L = 0$ and $\Sigma L = 114.59$ miles (total stream length in the drainage basin), thus:

$$\int C_{Bf} = 4.3 \int_0^{114.59} \Sigma L^{1.07}\, d\Sigma L$$

$$= 4.3\ [\Sigma L^{1.07}]_0^{114.59}$$

$$= 4.3\ [159.69 - 0]$$

$$= 686.66 \text{ square feet/mile}$$

In the calculation, ΣL was expressed in miles, but to obtain the volume in cubic feet, it is necessary to multiply the value by 5280. The volume of the drainage network above a catchment area of 26 square miles (total drainage area of West Fork basin), and up to bankfull stage is 3,625,670.40 cubic feet or 134,224.09 cubic yards. Table 17 shows the bankfull stage channel capacity and calculated network volumes.

TABLE 17

THE NETWORK VOLUMES FOR THE STUDY STREAMS

Formation		Equation	Bank Full Stage Channel Capacity Sq. Feet/Mile	Network Volume cubic ft./Mile
Sandstone	Kickapoo Creek	$C_{Bf} = 29\Sigma L^{0.17}$	51.98	274,483.19
	Devil's Canyon	$C_{Bf} = 10.7\Sigma L^{0.61}$	77.89	411,273.83
	Buggy Creek	$C_{Bf} = 24.5\Sigma L^{0.31}$	83.35	440,049.86
Shale	West Fork	$C_{Bf} = 4.3\Sigma L^{1.07}$	686.68	3,625,670.40
	Salt Creek	$C_{Bf}\ \ 17\Sigma L^{0.68}$	229.68	1,212,713.38

It is apparent that the network volumes at the bankfull stage are not comparable between sandstone basins and shale basins, so that the flood-producing potential of the two groups networks is not the same.

A number of studies have analyzed the relationship between stream discharges and channel capacity (Gregory, 1976b), and it is known that the cross sectional area of a stream channel is related to the frequency of occurrence of higher discharge (Gregory, 1977). If a change in the frequency of these higher discharges occurs, then changes in channel capacity would be expected. Two ways in which changes in the frequency of peak discharges may arise include dam construction, which may reduce the frequency of peak flows; and urbanization, which leads to an increase in frequency of peak flows (Wolman, 1967; Hammer, 1972). The first situation (dam construction) was investigated in Kickapoo Creek and Devil's Canyon Creek where these two streams were treated by floodwater retarding structures in 1962. The method of investigation was to employ the volume index of channel morphometry to evaluate the effect of floodwater retarding structures on channel morphometry. In the upper reaches of Kickapoo Creek and Devil's Canyon Creek, data on channel cross-section areas (channel capacity) were obtained of 15 cross-sections (8 on Kickapoo Creek and 7 on Devil's Canyon). The area of each cross section (C) was related to total stream length (ΣL) above each site. Comparable information from nine additional cross-sections was obtained at sites below the reservoirs. Two regression lines were then obtained: one to express the relationship between C and ΣL above the reservoirs and one pertaining to the data below the reservoirs. The regression equation for 15 cross-section area sites above the reservoirs is:

$$\begin{aligned} \int C_{Bf} &= 29.53 \int_0^{30.78} \Sigma L^{0.109} \, d\Sigma L \\ &= 29.53 \, [\Sigma L^{0.109}]_0^{30.78} \\ &= 29.53 \, [1.453 - 0] \\ &= 42.90 \text{ sq. feet/miles} \end{aligned}$$

then the network volume = 42.90 x 5280 = 226,549.44 cubic feet. The regression equation for 9 cross-section area sites below the reservoirs:

$$\int c_{Bf} = 5399.85 \int_{30.78}^{56.88} \Sigma L^{-1.41}\, d\Sigma L$$

$$= 5399.85\ [\Sigma L^{-1.41}]_{30.78}^{56.88}$$

$$= 5399.85\ [0.0033 - 0.0079]$$

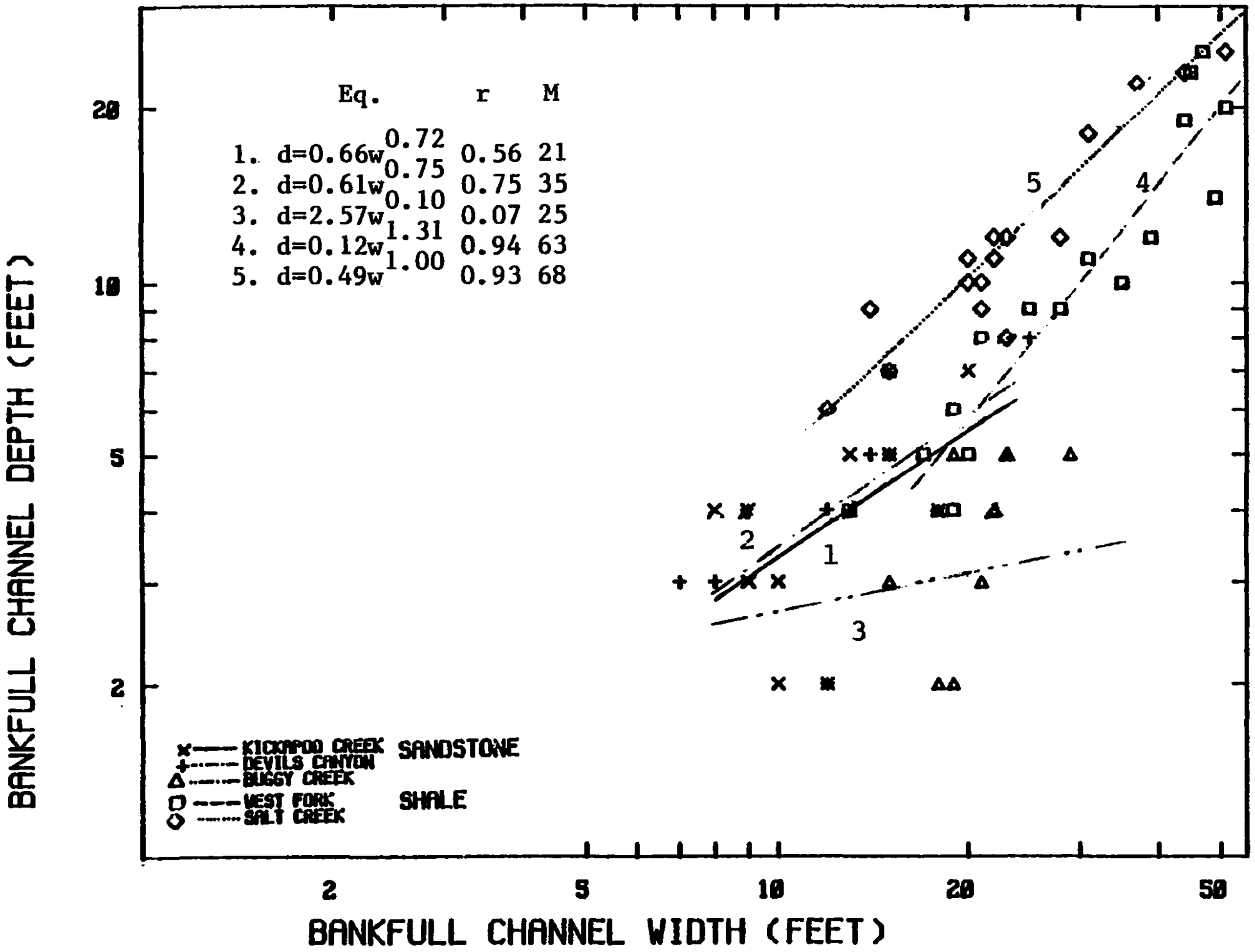
$$= 24.83 \text{ sq. feet/mile}$$

yielding a network volume of 131,151,56 cubic feet. Network volume above the reservoirs is twice the network volume below the reservoirs. The results, therefore, agree with Allen's (1979) findings that small channels immediately below floodwater retarding structures tend to aggrade, which is just the opposite of the effect of large structures (Leopold, et al., 1964). Aggradation occurs because peak flows are greatly reduced, reducing even further the sediment transporting capacity. Also, prior to structures most of these channels were ephemeral. After structures were built these channels had prolonged base flow periods that encouraged a lush growth of aquatic plants which in turn increased sedimentation.

The Relationship Between Channel Form and Sediment Properties

As Shumm (1960) noted, the reason for the relationship between channel shape and percentage of silt and clay is found in studies made by hydraulic engineers. Lane (1937, p. 124) states, ". . . the greater the width-depth ratio the greater will be the ratio of the velocity acting on the bottom to that acting on the sides of a channel." Where a narrow trench is cut in an alluvial valley, the tractive forces acting on the channel sides will be great, causing widening of the channel. Widening will continue until the resistance of the banks to scour prevents it. If the material in which the channel cut is highly cohesive (has a high percent of silt-clay) the channel will be narrow, but if the alluvium lacks cohesion (has a small percent of silt-clay) the channel will widen to a greater extent.

Also, Schumm (1960) showed that the relationship between M (weighted mean percent silt-clay) and channel form in a downstream direction is important. Most studies of fluvial geomorphology deal with relations between discharge and channel dimensions along a single river (Leopold, et al., 1964). The width and depth of a group of streams increases in an orderly manner with discharge. But the intercept or position of the regression line of width to depth for each stream could be determined by grain size (Figure 25). The position of the lines representative of all the streams in the area are apparently determined by M. The following are correlation coefficients between width and depth for individual streams and the correspondent M. values. Intercepts of streams on shale lie in a higher position in the diagram than those of streams on sandstone. The regression line for Buggy Creek lies in a flat position. This was reflected on a higher width-depth ratio (mean 6.07 and standard deviation of 2.016) compared to Kickapoo Creek (mean 3.39 and standard deviation 1.22). Although the



CHANNEL DEPTH VS CHANNEL WIDTH

Figure 25

TABLE 18

THE RELATIONSHIP BETWEEN CHANNEL WIDTH AND DEPTH WITH THE CORRESPONDENT M VALUE

Formation	Stream	Equation	r	R^2	Weighted Mean Silt-Clay Percent M
Sandstone	Kickapoo Creek	$depth = 0.66\ width^{0.72}$	0.56*	0.31	21
(group 1)	Devil's Canyon	$depth = 0.61\ width^{0.75}$	0.75*	0.56	35
	Buggy Creek	$depth = 2.57\ width^{0.10}$	0.07	0.0047	25
Shale	West Fork	$depth = 0.12\ width^{1.31}$	0.94*	0.88	63
(group 2)	Salt Creek	$depth = 0.49\ width^{1.006}$	0.93*	0.86	68

*significant at 0.05 level

intercepts of regression lines for Kickapoo Creek and Devil's Canyon (M is 21 and 35 respectively, Figure 25) indicate that these streams have lower depth, the slopes of the regression lines indicate that depth is related to width by a power function. The variation of channel dimensions of Kickapoo Creek and Devil's Canyon could be the result of several factors: 1) decreasing water table in the toe of the deposited alluvium, 2) higher channel gradient in areas above the reservoir of Kickapoo Creek which resulted in entrenching the channel (M = 21). In streams on shale (West Fork and Salt Creeks), the value of M is higher than streams on sandstone but does not change greatly from one cross section to another in the downstream direction. However, the regression lines of streams on shale in Figure 25 indicate an increase of channel depth in relation to width. This suggests, if M remains constant in the downstream direction then the width and depth of the channel will increase at a uniform rate with discharge, excluding the influence of other variables, and the width-depth ratio will remain constant.

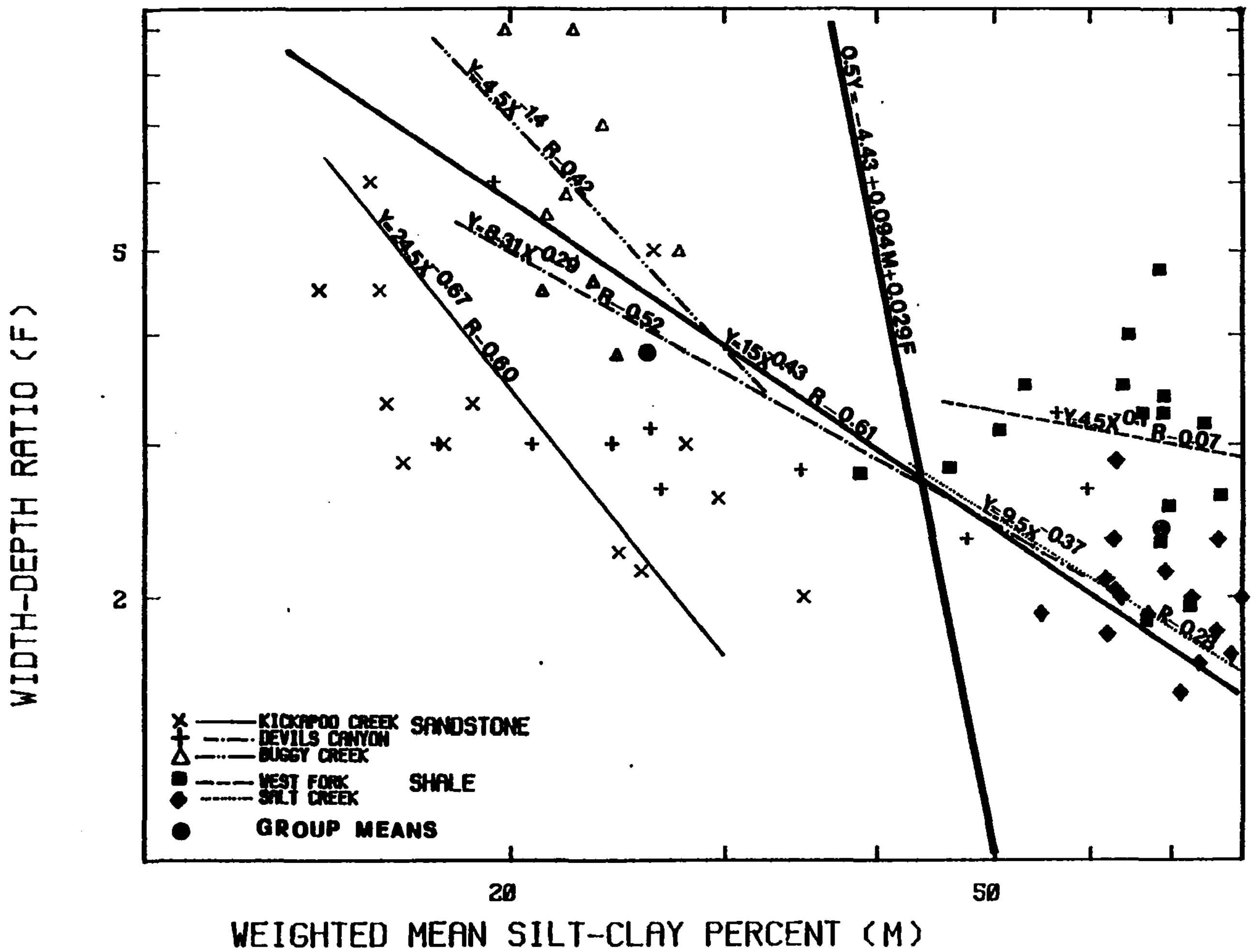
In Figure 26 plotting width-depth ratio (F) against silt-clay percent in the channel perimeter (M) for all the 65 cross-sections results in a linear pattern with a negative relationship. The pattern of the points confirm Schumm's relationship between F and M. But when data for individual streams are used to compute regression lines for each stream two groups of lines results. Moreover, the two groups of sandstone and shale formations are separated by a discriminant line. Table 19 shows the difference between regression lines of the groups (sandstone vs. shale).

TABLE 19

THE RELATIONSHIP BETWEEN WIDTH-DEPTH RATIO AND WEIGHTED MEAN SILT-CLAY PERCENT

Formation	Stream	Equation	r	R^2
Sandstone (group 1)	Kickapoo Creek	$F = 24.5\ M^{-0.67}$	-0.60*	0.36
	Devil's Canyon	$F = 8.31\ M^{-0.29}$	-0.52*	0.27
	Buggy Creek	$F = 4.5\ M^{-1.4}$	-0.42*	0.18
Shale (group 2)	West Fork	$F = 4.5\ M^{-0.1}$	-0.07	0.005
	Salt Creek	$F = 9.5\ M^{-0.37}$	-0.28	0.08
Total		$F = 15\ M^{-0.43}$	-0.61*	0.37

*significant at 0.05 level



WIDTH-DEPTH RATIO VS WEIGHTED MEAN SILT-CLAY PERCENT

Figure 26

a. Comparing the two groups slopes

$H_O\ \hat{\beta}_1 = \hat{\beta}_2$

$H_A\ \beta_1 > \beta_2$

$$t = \frac{\hat{\beta}_1 - \hat{\beta}_2}{S_{\hat{\beta}_1 - \hat{\beta}_2}} \quad \text{where } S_{\hat{\beta}_1 - \hat{\beta}_2} = \sqrt{S^2_{P,Y/X}\left[\frac{1}{(n-1)S^2_{x_1}} + \frac{1}{(n_2-1)S^2_{x_2}}\right]}$$

$$S_{\hat{\beta}_1 - \hat{\beta}_2} = \sqrt{0.26\left[\frac{1}{(32)0.35} + \frac{1}{(31)0.17}\right]}$$

$$= \sqrt{0.26(0.99)}$$

$$= \sqrt{0.257} \qquad = 0.5072$$

Then $t = \dfrac{1.89}{0.5072} = 3.72$

and tabulated $t_{60,}0.995 = 2.66$

therefore the H_A is accepted at 0.005 significant level.

Then, the two groups' regression lines are not parallel.

b. Comparing the two groups intercepts

$H_O\ A_1 = A_2$

$H_A\ A_1 > A_2$

$$t = \frac{2.66}{0.5072} = 5.24$$

and tabulated t_{60}, 0.995 is 2.66

Therefore, the H_A is accepted at 0.005 significant level. The intercept of the sandstone group is greater than that of the shale group.

A discriminant analysis was made to distinguish between the two groups based on geology (which were given the value of 1 for sandstone and 0 for shale) for observation of F and M. The group means are significantly different at 0.001 level.

Geology	F	M
0	2.53	66
1	4.02	26

On groups defined by geology, Wilk's Lambda was minimized to 0.77 for M and 0.22 for F. The standardized canonical discriminant function coefficients are;

	Function
M	1.012
F	0.043

and chi-square = 93.029 which is significant at 0.00001 level. Finally, the unstandardized discriminant function is

$$\text{Geol} = -4.43 + 0.094M + 0.029F$$

Schumm (1960) stated in his study that the relation of the unstable channels suggests that the position of points in relation to a regression line (above or below the regression line) may indicate that aggradation or degradation is occurring within a channel. Aggrading channels will have a higher width-depth ratio than indicated by M; whereas degrading channels will have a lower width-depth ratio than indicated by M.

The correlation coefficient for all the streams is -0.61, while the correlation coefficient of Schumm's relationship (F and M relationship) was -0.91. The Fisher Z transformation for testing the correlation between F and M is significantly lower than Schumm's correlation.

$\rho_o = 0.91$

$H_o\ \rho = 0.91$ using $r = -0.61$, $\rho_o = -0.91$

$H_A\ \rho \neq 0.91$ and $n = 65$

$$1/2\ \log_e \frac{1+\rho_1}{1-\rho_1} = 1/2\ \log_e \frac{1+0.91}{1-0.91} = 1.5275$$

$$1/2\ \log_e \frac{1+r}{1-r} = 1/2\ \log_e \frac{1+0.61}{1-0.61} = 0.7089$$

$$Z = \frac{0.7089-1.5075}{1/\sqrt{65-3}} = \frac{-0.9675}{0.1828} = -5.2927$$

for $\alpha = 0.05$, the critical region is

$$|Z| \geq Z_{0.095} = 1.96$$

Since $|Z| = 5.2927$ exceeds 1.96 the hypotheses $H_o \rho_o = -0.91$ is rejected at the 0.05 significance level. This result indicates that F values for individual streams may not primary be a function of M. This result confirms Schumm's idea that low correlation may indicate an unstable channel. Streams on sandstone are unstable streams and width-depth ratio may not decrease by increasing M. The decline of correlation between F and M for streams on shale does not confirm that as M increases downstream along a given stream, the depth increases more rapidly and the width less rapidly with discharge than if M were constant downstream and width-depth rates decrease.

Again, plotting F against M for all the 65 points (Figure 26) of the two groups streams (sandstone group and shale group) results in a pattern that confirms Schumm's (1960) relationship between F and M. Although r for the overall regression line is -0.61 is not significant compared with Schumm's r value of -0.91, the relation between F and M in the study area is significantly different from zero at 0.05 significant level. The overall F value is 36.73 and the PR > F is 0.0001. However, when individual streams are plotted, a grouping of lines results. The positions of the lines representative of these streams are not apparently determined by M.

The General Model

Emerging from the developments in channel form studies, subsequent to Leopold and Maddock's (1953) paper, the focus in this thesis is in three approaches to explain how geology (alluvium derived from sandstone and shale formations) controls the differential changes of channel form. These approaches are; first, to examine the relationship between channel capacity and force -- resistance index, second, to examine the relationship between channel capacity and network morphology; and third, to examine the relationship between channel form and sediment characteristics. The establishment of a causal relationship between geology and the independent variables to take into account their probable predictive value of channel form now can be achieved. The results presented in this chapter indicate a significant relationship between channel form and each controlling factor examined in the three approaches earlier. A covariance analysis was made to examine how geology influences channel form. The constants were substituted in equation (38) in Chapter V as:

$$TC = \frac{0.74 + 0.85\ T(\frac{S\gamma}{V})[0.82 + 0.86T\Sigma L]}{Td(1.17 - 0.43\ TM]} \quad \ldots\ldots\ldots\ldots \quad (46)$$

where, $TC = \log_{10}$ cross-section area (bankfull width x bankfull depth)

$T(\frac{S\gamma}{V}) = \log_{10}$ slope of the channel x (62.4) ÷ vegetation resistance index

$Td = \log_{10}$ bankfull depth

$T\Sigma L = \log_{10}$ total stream length above cross-section site

$TM = \log_{10}$ weighted mean percent silt-clay in channel perimeter and, the numbers are regression line constants derived from separate regression lines between:

1. TC and $T(\frac{dS\gamma}{V})$
2. TC and $T\Sigma L$
3. TF and TM

where TF is $\log_{10}$ width-depth ratio, and the rest of the terms are defined above.

The correlation and regression coefficients of the first regression equation are:

$$TC = 0.74 + 0.85\ T(\frac{dS\gamma}{V})$$

$r = 0.58$ $r^2 = 0.34$ $F = 32.32$ which is significant at 0.005 level

The correlation and regression coefficients of the second regression equation are:

$$TC = 0.83 + 0.86\ T\Sigma L$$

$r = 0.81$ $r^2 = 0.65$

$F = 119.56$ which is significant at 0.005 level

and the correlation and regression coefficients of the third regression equation are:

$$TF = 1.17 - 0.44\ TM$$

$r = -0.61$ $r^2 = 0.37$

$F = 36.73$ which is significant at 0.005 level.

However, when the dummy variable (D_1) is included, in order to test the influence of geology on channel form as it interacts with the controlling factors, a new covariance regression equation was obtained. The regression coefficients were substituted in equation (45) as follows:

$$TC = \frac{(0.95+0.22D_1)+(0.9-0.52D_1)T(\frac{S\gamma}{V})[(0.99+0.29D_1)+(0.84-0.48D_1)T\Sigma L]}{Td[(1.28-0.07D_1)+(-0.49+0.03D_1)TM]} \quad . . . . (47)$$

The terms were defined before, and numbers are regression model constants derived from separate regression lines between:

1. <u>TC</u> and D_1, $T(\frac{dS\gamma}{V})$, and $D_1[T(\frac{dS\gamma}{V})]$

2. <u>TC</u> and D_1, $T\Sigma L$, and $D_1(T\Sigma L)$

3. <u>TF</u> and D_1, TM, and D_1(TM)

The terms were defined before.

The correlation and regression coefficients of the first regression line are:

$$TC = 0.95+0.22D_1+0.90T(\frac{Sd\gamma}{V}) - 0.52D_1[T(\frac{dS\gamma}{V})]$$

$R = 0.84 \quad R^2 = 0.71 \quad F = 49.18$

The correlation and regression coefficients of the second regression equation are:

$$TC = 0.99 + 0.29\,D_1 + 0.84\ T\Sigma L - 0.48\,D_1\,(T\Sigma L)$$

$$R = 0.89 \qquad R^2 = 0.80$$

$$F = 80.41$$

and, the correlation and regression coefficients of the third regression equation are;

$$TF = 1.28 - 0.07\,D_1 - 0.49\,TM + 0.03\,D_1\,(TM)$$

$$R = 0.61 \qquad R^2 = 0.37$$

$$F = 11.89$$

A reliability test was made for equation (47) by plotting the logarithms of the observed values of cross-section areas (TC) against the logarithms of the predicted values of cross-sections (T17) by equation (47). The results were found insignificant and do not represent the real world.

$$TC = 3.12 - 0.107\ (T17) \tag{48}$$

$$R^2 = 0.28 \quad \text{and} \quad F = 25.42$$

The regression coefficient (slope) was found negative (-0.107) and $R^2 = 0.28$ was found insignificant. The reason was that the relationship between equations 32, 33, and 34 was not defined when constructing equations 38 and 45.

An attempt was made to write equation 38 and equation 45 in a new format when the result did not confirm the reliability of equations 38 and 45. The multiple regression equation for equation 38 became equation 49 and was calculated using the properties of the three approaches discussed in this chapter.

$$TC = -0.42 + 0.50\ T\left(\frac{ds\gamma}{V}\right) + 0.49\ TM + 0.65\ T\Sigma L \qquad \ldots\ldots\ldots\ldots (49)$$

$R = 0.91 \qquad R^2 = 0.83 \qquad F = 103.38$

And, the new covariance regression equation for equation 45 became equation 50 and was calculated using the same properties in the three approaches and geology as an indicator factor:

$$TC = -0.082 + 1.41\ D_1 + 0.58\ T\left(\frac{ds\gamma}{V}\right) + 0.17TM + 0.71\ T\Sigma L - 0.38D_1(T\Sigma L) - 0.21\ D_1\left[T\left(\frac{ds\gamma}{V}\right)\right] - 0.38D_1(TM) \qquad \ldots\ldots\ldots\ldots (50)$$

This equation can be written as,

$$TC = (-0.082+1.41D_1)+\left[(0.58-0.21D_1)T\left(\frac{ds\gamma}{V}\right)\right] + [(0.17-0.38D_1)TM]+[(0.71-0.38D_1)T\Sigma L] \quad \ldots (51)$$

$R = 0.94 \qquad R^2 = 0.88 \qquad F = 63.74$

The correlation matrix between the variables in equation 51 is:

	TC	D_1	$T(\frac{dS\gamma}{V})$	TM	TΣL	D_1 $[T(\frac{dS\gamma}{V})]$	D_1 (TM)	D_1 (TΣL)
TC	1.00	0.74	0.58	0.59	0.81	-0.63	-0.74	-0.60
D_1		1.00	-0.31	-0.86	-0.58	0.94	0.98	0.91
$T(\frac{dS\gamma}{V})$			1.00	0.27	0.29	-0.05	-0.31	-0.29
TM				1.00	0.42	-0.82	-0.78	-0.84
TΣL					1.00	-0.55	-0.60	-0.30
$D_1[T(\frac{dS\gamma}{V})]$						1.00	0.93	0.86
D_1(TM)							1.00	0.88
D_1(TΣL)								1.00

The reliability test for equation 49 was made (Figure 27).

$$TC = 0.00091 + 1.00021(T15) \quad . . . . (52)$$

$r = 0.91$, $r^2 = 0.83$ and F value = 300.28

where TC = observed cross-section area

and T15 = predicted cross-section area by equation 49

Also, the reliability test for equation 51 was made (Figure 28)

$$TC = 0.02629 + 0.99974(T9) \quad . . . . (54)$$

$r = 0.94$, $r^2 = 0.88$ and F value = 320.74

where TC = observed cross-section area

and T9 = predicted cross-section area predicted by equation 51.

Now, a comparison between streams on sandstone and streams on shale with regard to the regression line (using the dummy variable (D_1); 1 = sandstone, 0 = shale) of TC(Y) on $T(\frac{dS\gamma}{V})$; (X_1), TM; (X_2), TΣL; (X_3), D_1(TΣL); D_1X_3, $D_1(\frac{dS\gamma}{V}))$; D_1X_1, and

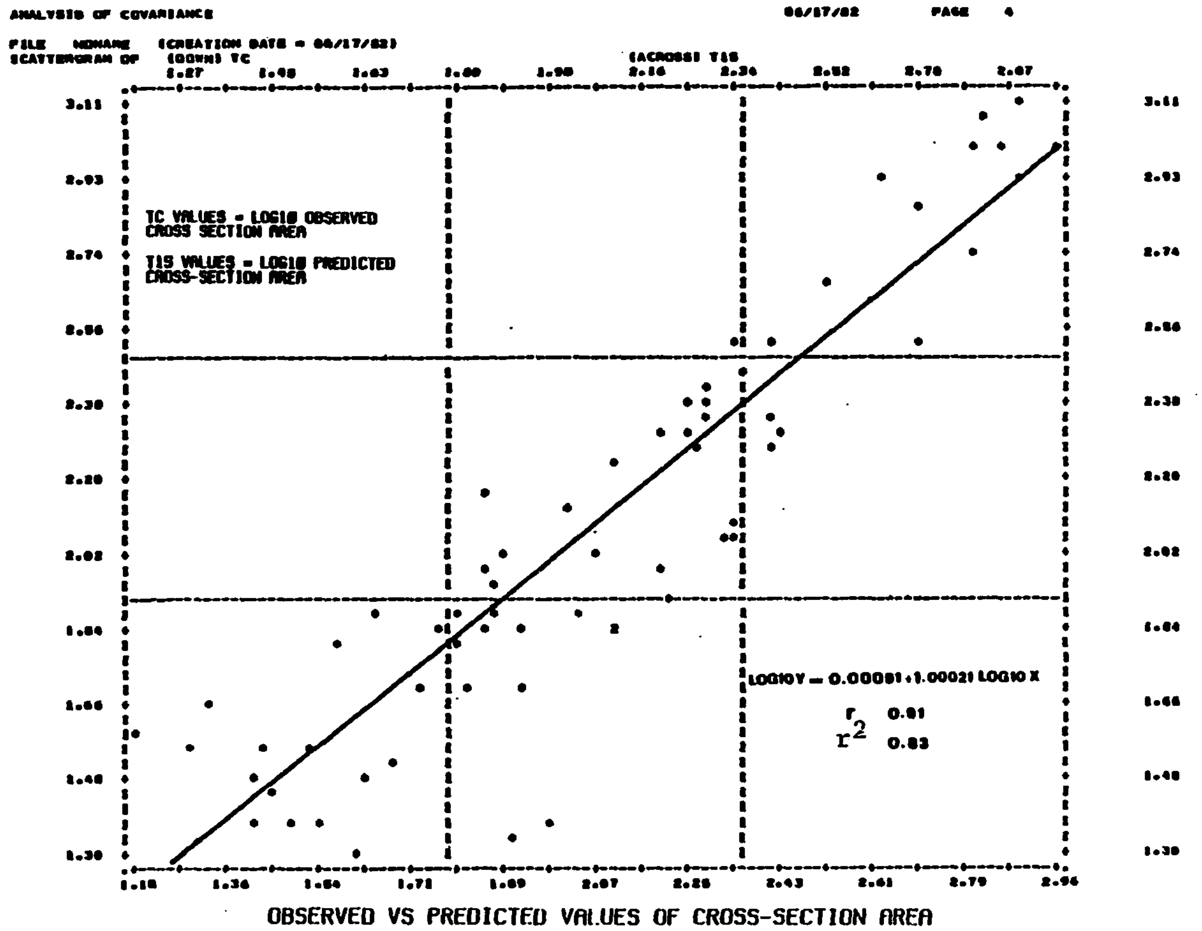

ANALYSIS OF COVARIANCE
TC VALUES = LOG10 OBSERVED CROSS SECTION AREA
TIS VALUES = LOG10 PREDICTED CROSS-SECTION AREA
LOG10 Y = 0.00081 + 1.00021 LOG10 X
r 0.91
r2 0.83
OBSERVED VS PREDICTED VALUES OF CROSS-SECTION AREA

Figure 27

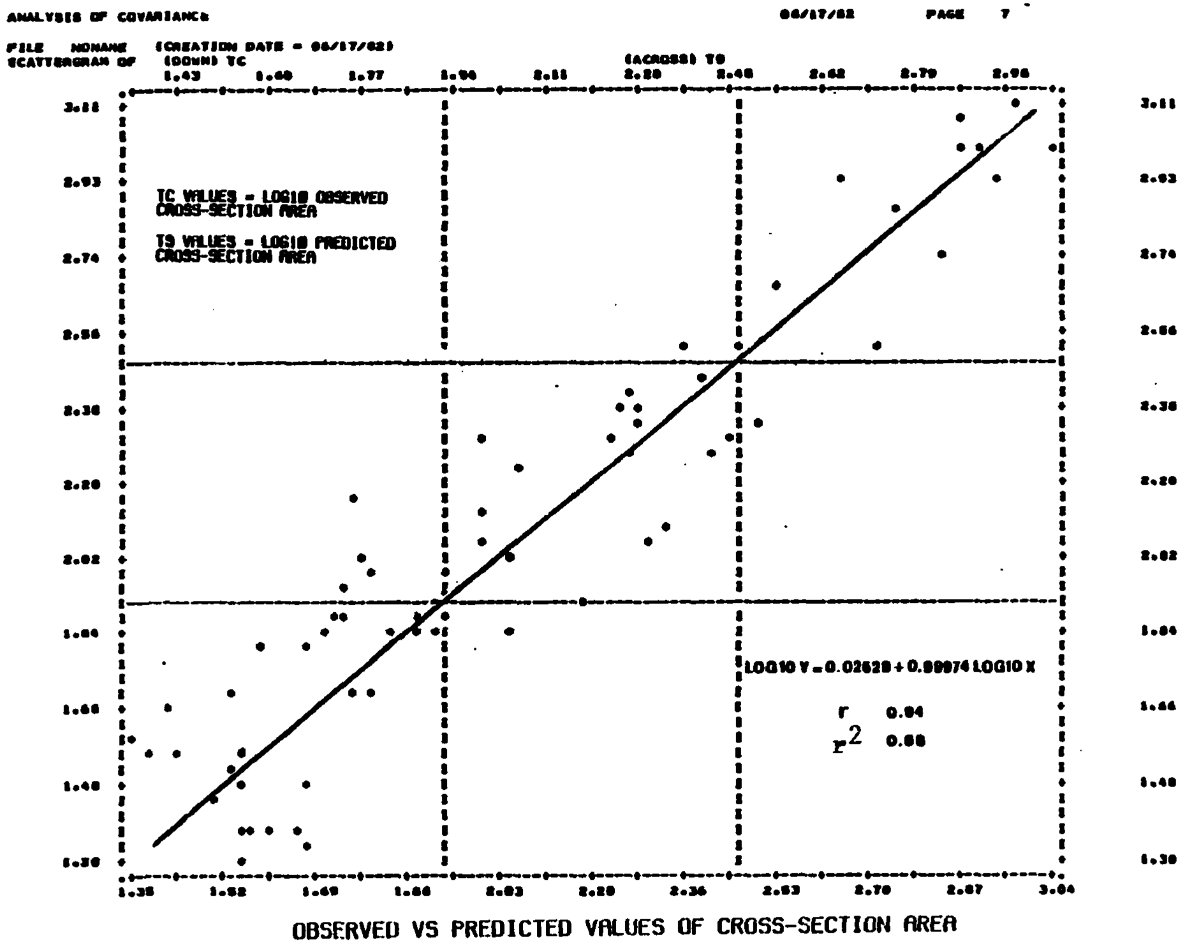

TC VALUES = LOG10 OBSERVED CROSS-SECTION AREA
TS VALUES = LOG10 PREDICTED CROSS-SECTION AREA
LOG10 Y = 0.02628 + 0.99974 LOG10 X
r 0.94
r2 0.88
OBSERVED VS PREDICTED VALUES OF CROSS-SECTION AREA

Figure 28

$D_1(TM); D_1X_2$, should be made.

Equation (51) fits the combined data for sandstone and shale as;

$$Y = B_0 + Z_1 + B_1X_1 + B_2X_2 + B_3X_3 + Z_2DX_3 + Z_3D_1 X_1 + Z_4D_1X_2 \quad \ldots \quad (55)$$

which yields the following two equations for the two values of D_1:

$$\text{Shale} \quad D = 0 = Y = B_0+B_1X_1+B_2X_2+B_3X_3 \quad \ldots \quad (56)$$

$$\text{Sandstone} \quad D = 1 = Y = (B_0+Z_1)+(B_1+Z_2)X_1+(B_2+Z_3)X_2+ (B_3+Z_4)X_3 \quad \ldots \quad (57)$$

For equation (55) the following two hypotheses are especially relevant.

1. The two equations (51 & 57) of $D = 0$ and $D \neq 1$ are parallel, $H_o:Z_2,Z_3$, and $Z_4 = 0$. When Z_2, Z_3, and $Z_4 = 0$, the slope for streams on sandstone (B_1+Z_2), (B_2+Z_3), and (B_3+Z_4) is simply equal to B_1, B_2, and B_3. The test statistics concerning $H_o:Z_2$, Z_3, and $Z_4 = 0$ is the usual partial F test for the significance of the addition of the variables; D_1X_3, D_1X_1, and D_1X_2 to an equation already containing X_1, X_2, X_3 and D_1.

Source	df	SS	MS	F
Regression $(X_1,X_2,X_3$	3	13.16580	4.3886	103.39
Residual	61	2.58943	0.04245	
Regression (D,X_1,X_2,X_3)	4	13.52893	3.38223	91.15
Residual	60	2.22630	0.03711	
Regression $(D,X_1,X_2,X_3,$ $DX_1,DX_2,DX_3)$	7	13.97069	1.99581	63.74
Residual	57	1.78455	0.03131	

Using these results, the following test is:

$H_0: Z_2,\ Z_3,$ and $Z_4 = 0$ (parallelism)

$F(Z_2,\ Z_3,\ Z_4 | X_1,\ X_2,\ X_3,\ Z_1) =$

$$\frac{\text{SS regression } (D_I, X_1, X_2, X_3, DX_1, DX_2, DX_3) - \text{SS regression}(D_I, X_1, X_2, X_3)}{\text{MS residual } (D_I, X_1, X_2, X_3, DX_1, DX_2, DX_3)}$$

$$= \frac{13.97069 - 13.52893}{0.03131} = 14.1092$$

Comparing this F with $F_{7,57,0.999} = 5.497$

therefore the H_o is rejected at 0.0001 level and then conclude that there is very strong evidence that the two lines are not parallel.

2. The intercepts of the two equations (56 and 57) are not the same that is,

$H_0: Z_1 = Z_2,\ Z_3,\ Z_4 = B$. When Z_1 and $Z_2,\ Z_3,\ Z_4$ are zero, the model $Y = (B_0 + Z_1) + (B_1 + Z_2)X_1 + (B_2 + Z_3)X_2 + (B_3 + Z_4)X_3$ reduces the model $Y = B_0 + B_1X_1 + B_2X_2 + B_3X_3$ (i.e., the two intercepts are the same. The test of $H_o: Z_1 = Z_2, Z_3, Z_4 = 0$ is the usual test made for parallelism.

$$H_o: Z_1 = Z_2, Z_3, Z_4 = 0 \text{ [coincidence]}$$

$$F(Z_2, Z_3, Z_4, Z_1 | X_1, X_2, X_3) =$$

$$= \frac{[\text{SS regression}(D, X_1, X_2, X_3, DX_1, DX_2, DX_3) - \text{SS regression}(X_1, X_2, X_3)]/2}{\text{MS residual } (D, X_1, X_2, X_3, DX_1, DX_2, DX_3)}$$

$$= \frac{[13.97069 - 13.16580]/2}{0.03131} = 12.8536$$

Comparing this F with $F_{3,60,0.999} = 8.652$, the H_o is rejected at 0.0001 level and then conclude that there is a strong evidence the two regression lines intercepts are not the same.

Summary

The purpose of this dissertation was to examine the role of geology and human intervention in controlling the stream channel form. The literature survey of stream channel morphology and channel changes was outlined and identified in Chapter II. Emerging from these developments in channel form studies, subsequent to Leopold and Maddock's (1953) paper, the focus in this dissertation was on three approaches to explain how geology (lithology) controls the differential changes of channel form characteristics. These approaches were some of the products of different theories explaining the fluvial system adjustments such as hydraulic geometry, complex response, thresholds, allometry, and catastrophe theory. In the hydraulic geometry approach there is a considerable range of exponents in the downstream geometries, and so this casts doubt on the use of mean values samples of exponents to characterize the hydraulic geometries of particular areas (e.g., Leopold & Maddock, 1953). In most of the literature cited in Chapter II no attempt was made to explain the causes of variations in the hydraulic geometries as reflected in the exponents and the constants. The other approaches to explain channel form change, such as complex response, allometry, and catastrophe, do fail in some cases to explain the magnitude of the spatial variation of channel form in a given area. For example, a critical valley slope, altered runoff conditions, and fluctuation of water table cause cut and fill in the valley stream. Note that it should be remembered here that drainage area and network morphology are used as substitutes for runoff or discharge. Therefore, if discharge increases, the effect would be to shift the threshold slope down for a given drainage area or a given network volume. The failure of a single approach to explain the variation of the downstream channel form change in a given area, is because of the unique characteristics of the study area. Spatial variation of channel form was hypothesized in this study to be the result of the interaction of geology (lithology) among various morphologic variables in selected drainage basins in western Oklahoma. This was ascertained through general field observations and substantiated by a review of the literature.

The location of the study area is no less important in this study. The drainage area of the Washita River is the study area for the Southern Plain Watershed Laboratory in Chickasha, Oklahoma. The area is 1130 square miles within which the basins included in the study were selected. About 183 square miles (16.2%) of the Washita study reach has undifferentiated Quaternary age alluvium and terrace deposits. The remaining 947 square miles (83.8%) has Permian age sedimentary rock exposed at the surface or near the surface are covered by a thin veneer of poorly developed soil. In the study basins, the areas underlain by different lithology varies from basin to basin as follows; in West Fork and Salt Creek 90 percent underlain by Marlow, Dog Creek, and Blaine Formations, 6 percent underlain by Chickasha Formation, and 4 percent underlain by alluvium. On the other hand, in Kickapoo Creek, Devil's Canyon and Buggy Creek, about 72 percent is underlain by Rush Springs Formation and about 28 percent underlain by alluvium and terrace deposits. The alluvium composing the study stream channels perimeters is different from stream to stream. These differences are expressed in the depth and texture of the alluvium. The fine texture of the West Fork and Salt Creek alluvium is apparently due to the nature of the Permian materials (shale) that West Fork and Salt Creek drain. The fine texture of the source of the alluvium may partially explain the alluvium's great depth and therefor silt-clay percent may increase and in turn channel depth increases in the downstream direction. However, the thin veneer of poorly developed soil which covers a coarser material is due to the material that Kickapoo, Devil's Canyon, and Buggy Creek drain. This coarse material is sandstone. Streams on sandstone were found to have a great variation of channel form in the downstream direction due in part to landuse activities. The SCS floodwater structures are dominant in areas underlain by sandstone. The effect of these structures on the downstream water yield and thereby on the downstream channel form is high below the structures. However, determining the effect further downstream is difficult because of vegetation removal, cultivation, and farm pond constructions. The effect of these landuse activities on channel form were found upstream from the structures as well.

Examination of geological control and landuse impact on stream channel morphology was the subject of each hypothesis tested in this study. Firstly, streams on sandstone can be considered as candidates for the application of the catastrophe concept, which is a process that includes abrupt changes as thresholds are crossed. The algorithm using tractive force explains why gradient is not the only variable with a significant threshold. Similar thresholds exist for the resisting vegetation. This was evident in the significant relationship when channel cross-section area (C) was regressed against the force-resistance index (TV). Average correlation coefficients for streams on sandstone were 0.81, and the regression coefficients were found significant at 0.005 level. These statistical results reveal that various valley floor changes introduced by man (i.e., cultivation of valley floor) have their impact on differential channel form in the downstream direction. On the other hand, this hypothesis was found to be difficult to prove in streams on shale. Increasing discharge from shale basins increases the depth of channel in the downstream direction at a faster rate than the width of the channel.

Secondly, streams on shale are best considered from an allometric rather than from a steady state viewpoint. Drainage density in the sandstone basins was found to be higher than in the basins on shale as a result of the large area in sandstone basins converted from rangeland to cropland. However, a comparison of drainage densities is a purely spatial comparison and one which ignores the volume of the channels constituting the network. Regression relationships were then established between the channel capacity and total stream length so that a volumetric comparision was made between the groups (sandstone group and shale group). It was determined, by testing the hypothesis, that there is a relationship between channel capacity and network volume, and that network volume in shale basins is higher than in sandstone basins. Therefore, flood hazards are higher in streams on sandstone than in streams on shale. This was evident in the increases of channel size of streams on shale in the downstream direction with increasing discharge with the other factors held constant.

Thirdly, because of the great range of the channel size of the streams studied the tkpe of sediment characteristics collected from the stream bed and banks is considered to be an important control on channel shape. Schumm's (1960) relationship between channel dimensions and an index of the type of sediment load (M) was found significant if all the 65 observations, collected from streams on sandstone and shale, were plotted. The pattern of the points confirms Schumm's relationship between F and M, but when data for individual streams were used to compute a regression line for each stream, two groups of lines resulted. Local changes of channel dimensions would be and strongly affected by variations of resistance of bank sediments. For example, in streams on shale, M does not change greatly from cross-section to cross-section. If M remains constant downstream then the width and depth of the channel will increase at uniform rate with discharge, excluding the influence of other variables, and width-depth ratios will remain constant. On the other hand, streams on sandstone do have a variable M along the stream channels. The average correlation coefficient between F and M for streams on sandstone was found to be -0.51, while the correlation coefficient found by Schumm was -0.91. This result suggests that streams on sandstone are unstable, because the local cut and fill in these sandy streams could be a result of fluctuation of the water table. The position of the separate regression line for each stream, is apparently not determined by M.

Emerging from examination of the three approaches in streams on sandstone versus streams on shale a covariate model includes all three approaches, as predictions of channel form will not be reliable unless a dummy variable representing geology is included. The reliability test for the model generated an r^2 of 0.88 and F value of 320.74.

Subsequently, the extension of the morphometric approach to characterizing channel form variations to other streams in Oklahoma and elsewhere is worthy of further attention. Ideally such extension could cover streams under a wide range of environmental conditions, so that the analysis might be used on the world and local scales. If additional data of this type is collected from different regions, with differing climatic, geologic, and topographic conditions, the ability to identify potentially unstable segments of a stream channel will increase. Predicting the future behavior of currently unstable stream channels is one major goal of this study. The morphologic variables within geologic control, which form the causal model developed in this study, will provide a basis for prediction.

There appear to be a number of important themes which emerge from the present study. One is the utilization of the morphometric approach of relating channel form properties to drainage basin characteristics (i.e., stream network volume) and tractive force-resistance index as a valuable technique for studying man-induced changes in channel form. The brutal forcing of a channel into an unnatural straight alignment almost always produces serious consequences. Unless the new course is cut in resistant materials the channel will attempt to resume its meandering course. In addition, the greatly increased gradient of the straightened channel will normally cause incision and rejuvenation upstream, thereby producing a high sediment load that will probably enlarge and aggrade the aligned channel and cause flood plain destruction. Again, the channels that will benefit most from partial to full channelization probably can be identified by a careful study of channel morphology, which this study attempts. Furthermore, flow regulation by reservoirs in the area has led to a drop of channel capacity. By plotting channel capacity against stream network morphology, the results showed that the reduction in capacity is substantial immediately below dams.

A second theme is examining the three approaches provides conclusive evidence of the detailed form of change in channel process. It does also allow the initial delimitation of stream reaches which have undergone such changes, following which more detailed research in the future can be devoted to the mechanisms by which the channel changes have been brought about.

Subsequently, the establishment of a causal relationship between geology and the independent variables, to take into account their probable predictive value of channel form, was achieved. Geology as a cause of differential channel form development remains a subject of fertile research.

CHAPTER VII

CONCLUSIONS

The general hypotheses stated that geology and man's modifications of natural streams determine the spatial variations of downstream channel form change.

Three independent variables—force-resistance index, total stream length above cross-section area, and weighted mean silt-clay percent—estimated the channel cross-section area fairly consistently. They were used to determine the influence of geology and detect the effect of man's activities on channel form. All had significant regression coefficients when all the observations were used in the calculation. Analysis of covariance indicated that multiple correlation and R^2 were increased when geology was introduced as a dummy variable.

The t test was used in three subhypotheses to test the parallelism of the regression lines of two groups (the group of streams on sandstone versus the group of streams on shale). Each test indicated that there is a strong evidence of no coincidence between the two group's regression lines, thus rejecting the null hypothesis at 0.005 significant level.

The three subhypotheses were stated as causal relationships generated from other studies in other regions. As a result of varying lithology in the region, caution was exercised in using these relationships. A goal of this study was to emphasize environmental problems, such as varying lithology and mans' alterations of the natural system, in the examination of spatial variations in channel form in western Oklahoma. It was anticipated that only a slight facies change from sandy to shaley materials in Middle Permian Red beds of western Oklahoma would be sufficient to produce profound differences in stream channel characteristics.

Several generalizations can be stated with regard to the simple regression analysis, covariance technique, and discriminant analysis conducted in this study. The force-resistance index predicted channel cross-section area more appropriately than either tractive force or resistance offered by reparian vegetation cover. Due to varying lithology vegetation resistance index does not appear to correlate significantly with tractive force in this study area, contrary to Graf's (1979) study. Generally, streams on shale are deep, and the established riparian vegetation protects stream banks from slumping and failure. On the other hand, streams on sandstone are easily cut when vegetation in the valley floor is altered. Basically, the force-resistance index as a factor in relation to channel cross-section area is new and an important variable, particularly when lithology is included as an indicator.

While the total stream length above cross-section area was significant in the simple regression equation in predicting channel cross-section area, it was a more important factor in predicting network volume. This was particularly true when watersheds are treated with flood detention dams and when the network volume of watersheds on shale was compared to the network volume of watersheds on sandstone. It was found that the network volumes to the bankfull stage are not comparable between sandstone basins and shale basins, so that the flood-producing potential of the two groups' networks is not the same. The same was true that the network volume below flood-water retarding structures is less than the network volume above these structures. Contrary to large dams, deposition occurs just below the structures. This is perhaps due to reduction of flow and rapid growth of vegetation in the channel bed.

The weighted mean silt-clay percent was a relatively significant predictor for width-depth ratio of the channel cross-section when all observations are included. In this study area, the weighted mean silt-clay percent is not a significant estimator of channel width-depth ratio of streams on shale. The lower correlation for streams on shale caused a paradox in this study area (i.e., where Schumm's (1960) relationship between width-depth ratio and weighted mean silt-clay percent is -0.91, and in this study is -0.61). The increase of depth in the downstream direction for streams on shale is not determined by mean silt-clay percent. Discriminant analysis indicated differentiation between streams on sandstone and streams on shale as a function of width-depth ratio and mean silt-clay percent to denote the importance of geological control when estimating channel cross-sectional characteristics; cross-section area and width-depth ratio.

Along with the independent variables used in the covariance technique to predict channel cross-section area, geology was important in determining the spatial variation of downstream channel form change. Other studies have used these independent variables, without geology as an indicator factor in determining channel form, successfully. This research has almost ruled out their apparent importance without the geological factor for predicting the downstream channel form in this study area.

Summation

This research showed that it is necessary to disclose the effects of the environment upon the erosional and depositional mechanics of fluvial systems. It emphasized the environmental problem in fluvial research when examining the downstream channel form change over varying lithology, and signals the precaution necessary when results of research in one region are applied in another. The complexity of fluvial systems does not preclude the need for many studies over many environmental situations, in fact this is demanded.

Combining the subhypotheses of the study into a general model was unsuccessful. It was found, however, that the covariance technique including the properties of the three subhypotheses shows a consistency between the independent variables in estimating channel cross-section area.

Scientific laws may be probabilistic because observed phenomena in the real world are complex. However, when probabilistic statements are generated, strong relationships can be tested for their generality. The problem hypotheses in this research were generated from assumed causal relationships, thus, the probability to synthesize a deterministic conclusion was higher. This research, therefore, determined that steams on sandstone suggest themselves as candidates for the application of a process which includes abrupt changes as thresholds are crossed, while streams on shale are candidates for allometric change.

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APPENDIX 1

Data

APPENDIX 1

DATA

Geol = geology takes the value of 1 if it is sandstone formation, and 0 if it is shale formation
ST = stream; where K is Kickapoo Creek, D is Devils Canyon Creek, B is Buggy Creek, W is West Fork Creek, and S is Salt Creek.
W = bankfull channel width in feet
D = bankfull channel depth in feet
S = stream channel slope at the cross-section site feet/foot
C = stream channel cross-section area (width x depth) in square foot
F = width-depth ratio (width/depth)
τ_o = tractive force in pound/square foot ($\tau_o = \gamma dS$; where $\gamma = 62.4$, d is depth of flow, and S is channel slope)
H = hypsometric curve function (statistical moments; the slope of the curve at each cross-section point determined on the curve)
ΣL = total stream length above the cross-section site (miles)
Dd = drainage density (stream length/drainage area) in miles/square mile
S_c = silt-clay percent in the sediment composing the channel bed
S_b = silt-clay percent in the sediment composing the channel banks
M = weighted mean silt-clay percent in the channel perimeter
V = percent riparian vegetation cover; V = 0.7 (percent grass cover) + 0.3 (percent forest cover)
TV = force-resistance index; tractive force/percent vegetation cover.

OBS	GEOL	ST	W	D	S	C	F	T	H	L	Dd	SC	SB	M	V	TV
KICKAPOO CREEK																
1	1	K	9	4	0.0513	36	2.250	12.804	1.164	3.55	1.579	34.67	13.14	24.53	30.00	0.43
2	1	K	9	3	0.0591	27	3.000	11.063	1.391	5.29	1.922	37.42	13.42	27.82	45.91	0.24
3	1	K	10	3	0.0409	30	3.333	7.656	1.559	7.80	2.152	18.85	18.55	18.73	42.89	0.18
4	1	K	15	5	0.0711	75	3.000	22.183	1.878	9.21	2.992	6.28	34.58	17.60	36.22	0.45
5	1	K	20	7	0.0369	140	2.857	16.117	1.657	13.72	3.056	13.14	20.85	16.31	12.73	1.66
6	1	K	18	4	0.0482	72	4.500	12.030	1.708	14.22	3.112	13.42	20.57	15.62	23.83	0.50
7	1	K	12	2	0.0401	24	6.000	5.004	1.824	16.01	2.921	18.42	6.00	15.31	50.91	0.09
8	1	K	10	2	0.0359	20	5.000	4.480	1.889	17.55	3.177	27.42	23.14	26.19	39.99	0.11
9	1	K	15	7	0.0257	105	2.142	11.225	2.194	17.98	2.717	34.57	16.00	25.60	27.27	0.41
10	1	K	13	5	0.0209	65	2.600	6.520	2.249	19.43	2.482	34.27	23.14	29.60	31.81	0.21

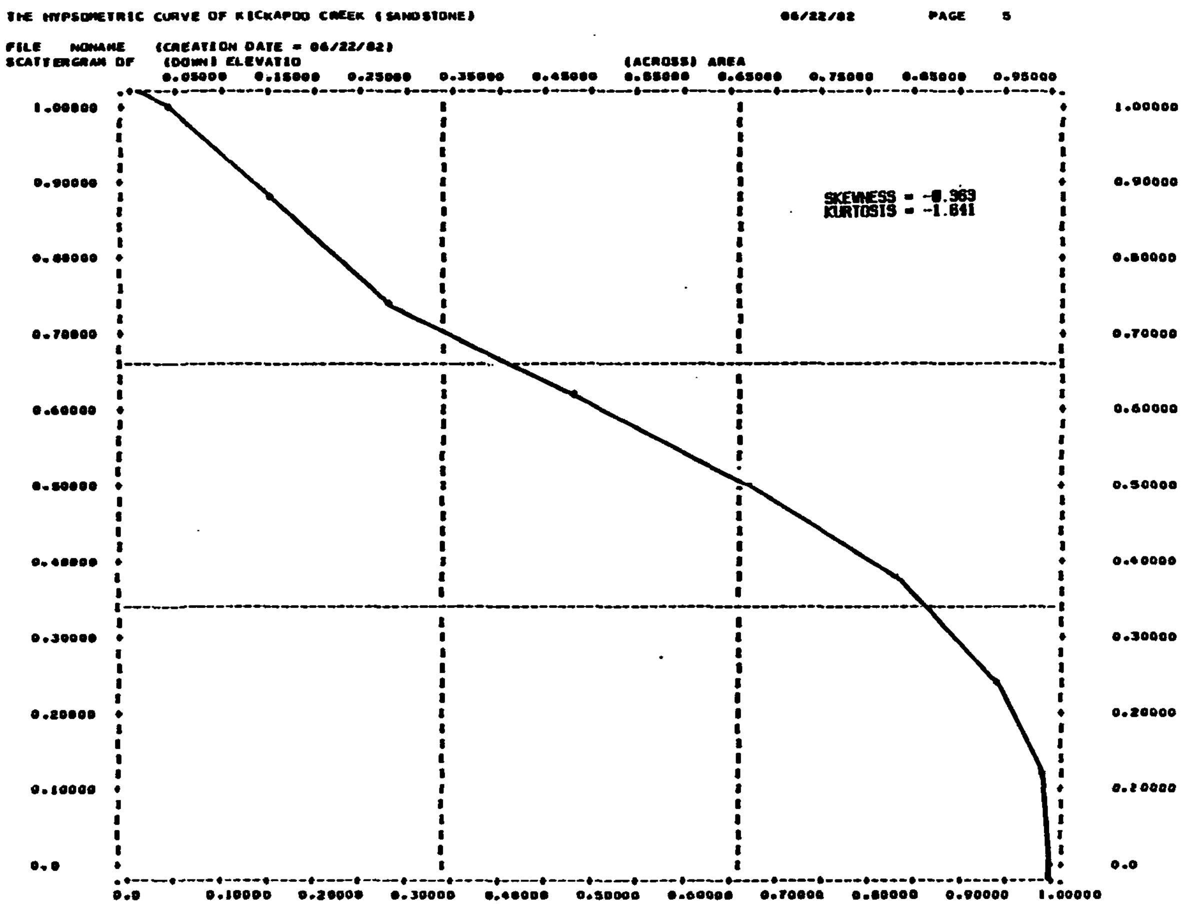
OBS	GEOL	ST	W	D	S	C	F	T	H	L	Dd	SC	SB	M	V	TV
11	1	K	8	4	0.0165	32	2.000	4.118	2.249	22.01	2.455	34.57	34.57	34.82	44.54	0.09
12	1	K	10	3	0.0305	30	3.000	5.709	2.249	27.49	2.662	13.14	20.28	15.81	50.91	0.11
13	1	K	18	4	0.0557	72	4.500	13.902	2.249	30.98	2.529	13.42	15.03	13.91	44.54	0.99
DEVIL'S CANYON CREEK																
14	1	D	13	4	0.0451	52	3.250	11.256	1.772	5.50	2.353	60.57	40.42	56.32	27.27	0.41
15	1	D	9	4	0.0342	36	2.250	8.536	1.913	5.60	1.098	65.50	45.00	55.85	50.91	0.16
16	1	D	8	3	0.0741	24	2.666	13.871	2.048	6.01	1.221	48.85	13.43	26.52	45.22	0.31
17	1	D	12	2	0.0989	24	6.000	12.342	2.119	7.25	1.750	18.50	22.00	19.37	44.54	0.28
18	1	D	14	5	0.0911	70	2.800	30.295	2.257	9.29	1.210	34.57	34.57	34.57	50.91	0.60
19	1	D	7	3	0.1201	21	2.333	22.483	2.253	10.20	1.998	50.22	44.22	47.45	42.92	0.52
20	1	D	8	3	0.0875	24	2.666	16.380	2.382	13.23	2.001	59.33	60.02	59.62	40.38	0.40
21	1	D	15	5	0.0599	75	3.000	18.688	2.389	16.97	2.000	20.28	13.14	17.42	30.00	0.62
22	1	D	25	8	0.0618	200	3.125	30.850	2.430	18.80	1.906	20.57	34.57	26.03	19.10	1.62
23	1	D	15	5	0.0702	75	3.000	21.902	2.438	23.87	1.995	18.00	25.22	20.80	31.82	0.69
24	1	D	12	4	0.0519	48	3.000	12.954	2.444	25.90	2.009	23.69	25.00	24.21	38.82	0.33
BUGGY CREEK																
25	1	B	15	3	0.0261	45	5.000	4.885	0.270	5.90	1.291	29.14	22.00	27.50	40.92	0.12
26	1	B	19	2	0.0255	38	9.500	3.182	0.274	6.21	1.812	25.09	10.25	22.50	42.44	0.07
27	1	B	18	2	0.0219	36	9.000	2.733	0.289	10.00	1.991	21.71	11.14	19.78	39.19	0.07
28	1	B	21	3	0.0204	63	7.000	3.818	0.372	10.92	2.298	26.28	15.14	23.80	19.10	0.20
29	1	B	29	5	0.0165	145	5.800	5.148	1.080	31.21	1.865	23.42	18.57	22.17	25.45	0.20
30	1	B	22	4	0.0152	88	5.500	3.793	1.587	40.92	2.072	23.14	16.57	21.38	25.45	0.15
31	1	B	19	5	0.0119	95	3.800	3.713	1.866	70.23	2.073	33.42	7.42	24.45	12.73	0.29
32	1	B	18	4	0.0115	72	4.500	2.870	2.208	98.80	2.062	26.66	9.00	21.22	19.10	0.15
33	1	B	12	5	0.0111	115	4.600	3.463	2.491	129.22	2.518	28.50	11.55	23.36	12.72	0.27
WEST FORK OF SALT CREEK																
34	0	W	13	4	0.0378	52	3.250	9.434	1.277	9.92	4.112	70.57	66.28	68.92	37.26	0.25
35	0	W	19	4	0.0312	76	4.750	7.787	1.364	10.46	4.269	70.57	63.42	68.42	37.91	0.21
36	0	W	20	5	0.0256	100	4.000	7.987	1.405	14.00	4.001	63.43	66.28	64.38	41.81	0.19
37	0	W	21	8	0.0244	168	2.625	12.180	1.432	21.53	4.196	70.57	84.85	76.74	39.30	0.31
38	0	W	17	5	0.0186	85	3.400	5.803	1.599	41.02	4.987	70.57	66.28	68.98	25.45	0.23
39	0	W	19	6	0.0137	114	3.166	5.129	1.752	56.53	4.165	77.71	70.57	74.44	21.00	0.24

OBS	GEOL	ST	W	D	S	C	F	T	H	L	Dd	SC	SB	M	V	TV
40	0	W	28	9	0.0137	252	3.111	7.693	1.776	59.02	2.812	56.00	41.71	50.39	31.82	0.24
41	0	W	35	10	0.0130	325	3.500	8.112	1.807	62.68	4.450	63.42	34.57	52.93	25.85	0.31
42	0	W	25	9	0.0130	225	2.777	7.301	1.807	66.02	2.099	41.71	34.57	38.72	13.64	0.54
43	0	W	31	11	0.0130	341	2.818	8.923	1.807	67.72	4.206	48.85	41.71	45.88	20.18	0.44
44	0	W	39	12	0.0122	468	3.250	9.135	1.864	85.55	4.239	56.28	49.14	66.16	24.54	0.37
45	0	W	49	14	0.0175	686	3.500	15.288	1.905	82.82	4.391	66.82	58.22	63.69	15.88	0.96
46	0	W	44	19	0.0181	836	2.315	21.459	1.905	98.23	3.981	70.56	66.28	68.57	41.82	0.51
47	0	W	51	20	0.0228	1020	2.550	28.454	1.905	104.09	4.441	74.43	63.43	69.59	25.91	1.09
48	0	W	45	23	0.0235	1035	1.956	33.727	1.946	108.92	4.011	74.57	70.57	72.54	27.27	1.23
49	0	W	47	25	0.0209	1175	1.880	32.604	2.012	114.59	4.324	70.28	63.43	66.75	31.82	1.02
SALT CREEK																
50	0	S	12	6	0.0962	72	2.000	25.908	0.682	8.11	3.921	63.71	63.42	63.57	44.54	0.58
51	0	S	15	7	0.0613	105	2.142	26.775	0.818	9.03	4.043	72.00	66.28	69.23	30.91	0.87
52	0	S	14	9	0.0557	126	1.555	31.281	0.895	18.10	4.001	72.00	70.57	71.19	27.27	1.15
53	0	S	20	10	0.0193	200	2.000	12.043	1.923	25.87	3.164	74.85	70.57	72.71	25.10	0.48
54	0	S	22	12	0.0159	264	1.833	11.906	1.222	36.90	2.921	82.00	70.85	76.18	33.63	0.35
55	0	S	22	11	0.0155	242	2.000	10.639	1.371	41.10	2.809	84.85	75.14	79.99	39.09	0.27
56	0	S	23	12	0.0151	276	1.916	11.306	1.446	49.59	2.700	56.28	53.14	54.67	24.54	0.46
57	0	S	20	11	0.0145	220	1.818	9.952	1.477	55.00	3.550	63.43	60.57	61.93	40.91	0.24
58	0	S	21	10	0.0142	210	2.100	8.861	1.611	63.80	2.992	70.85	52.28	61.79	21.82	0.41
59	0	S	23	8	0.0129	184	2.875	6.439	1.595	72.87	3.448	67.71	56.00	62.90	41.49	0.24
60	0	S	21	9	0.0119	189	2.333	6.683	1.643	112.01	3.941	70.57	53.43	62.65	39.08	0.17
61	0	S	28	12	0.0112	336	2.333	8.386	1.656	132.30	3.808	77.71	74.85	76.39	19.10	0.44
62	0	S	31	18	0.0111	558	1.722	12.467	1.825	139.00	4.000	67.81	87.43	78.35	19.10	0.65
63	0	S	37	22	0.0109	814	1.681	14.963	1.884	149.27	4.172	77.71	70.28	73.68	15.82	0.94
64	0	S	44	23	0.0100	1012	1.913	14.352	1.914	158.40	4.090	70.85	63.15	66.91	10.91	1.32
65	0	S	51	25	0.0091	1275	2.040	14.196	1.928	178.87	4.004	70.71	56.00	62.81	16.36	0.87

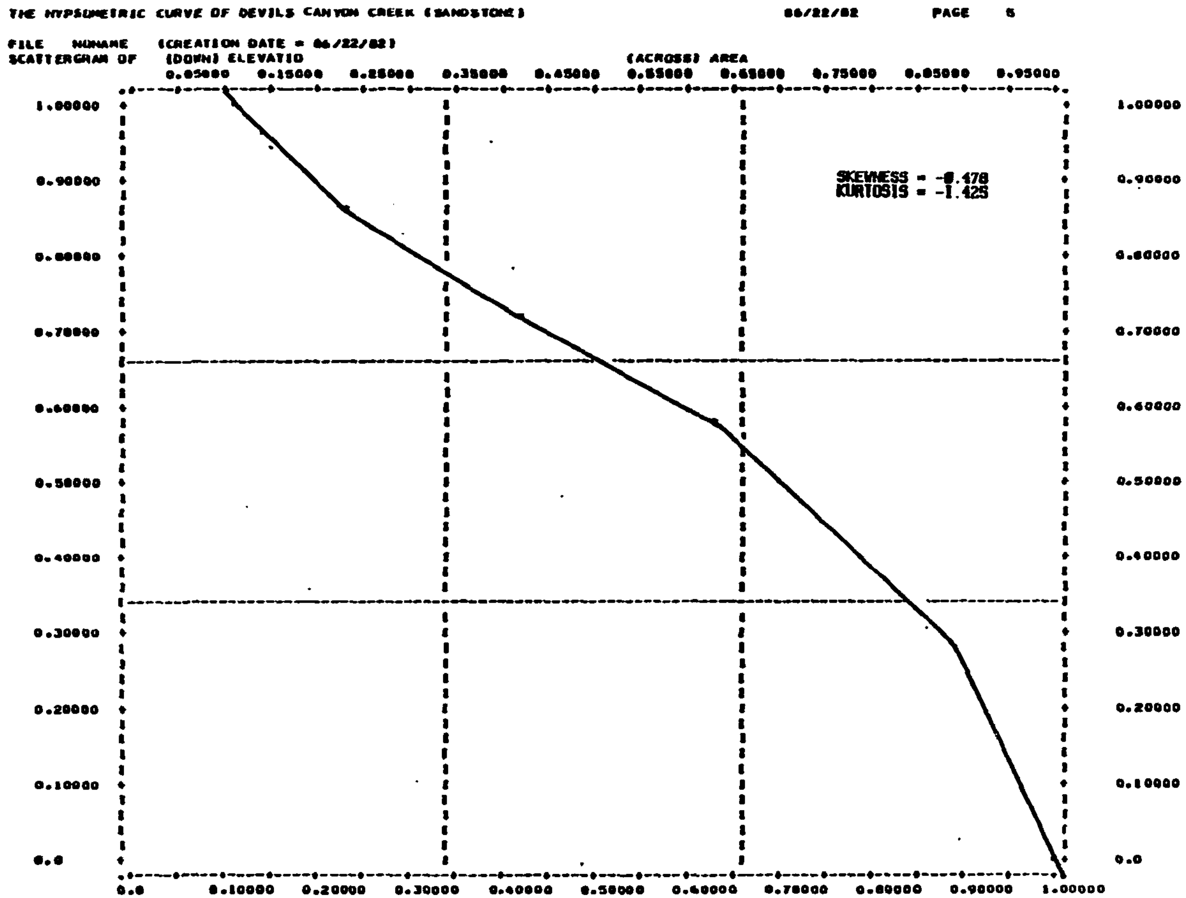
APPENDIX 2

Hypsometric Curves of the Study Drainage Basins

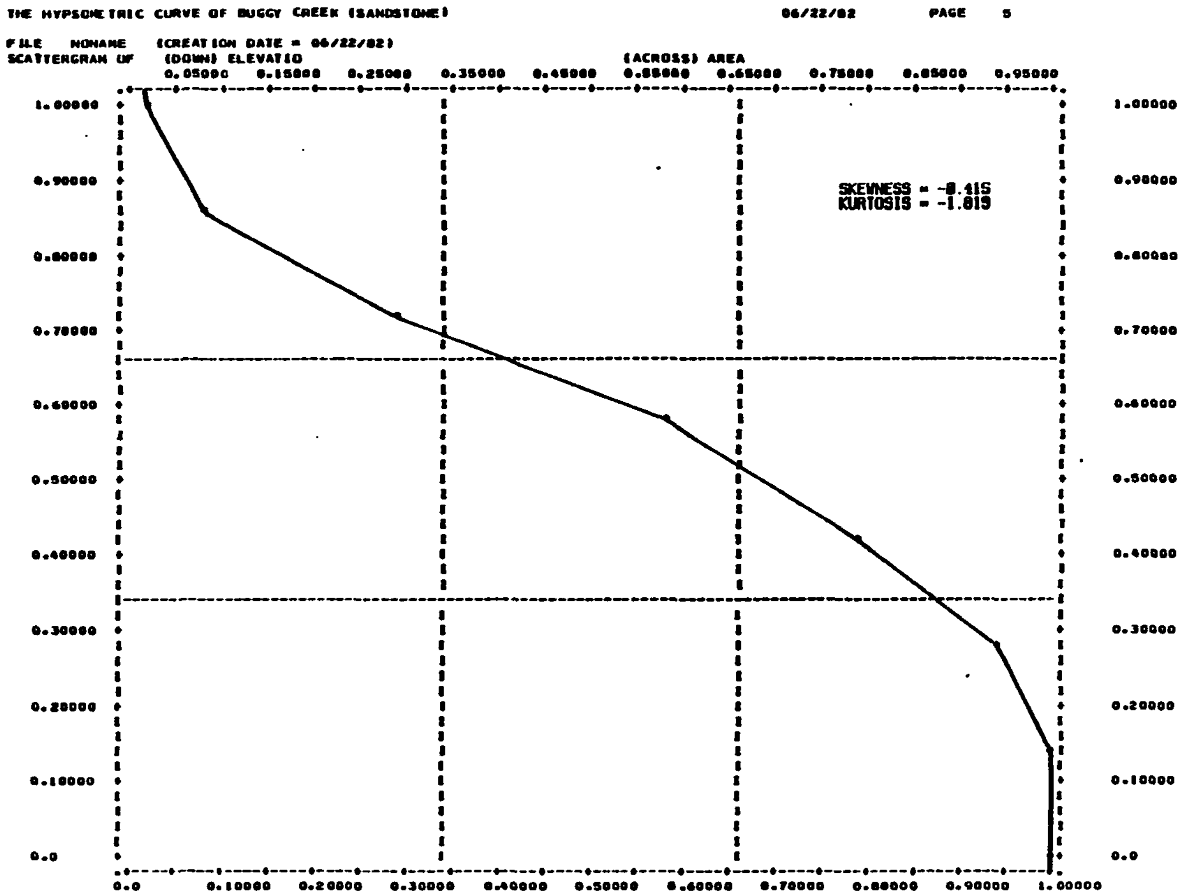
Appendix 2


THE HYPSOMETRIC CURVE OF KICKAPOO CREEK (SANDSTONE)
06/22/82
PAGE 5
FILE NONAME (CREATION DATE = 06/22/82)
SCATTERGRAM OF (DOWN) ELEVATIO
(ACROSS) AREA
0.05000 0.15000 0.25000 0.35000 0.45000 0.55000 0.65000 0.75000 0.85000 0.95000
1.00000
0.90000
0.80000
0.70000
0.60000
0.50000
0.40000
0.30000
0.20000
0.10000
0.0
SKEWNESS = -0.369
KURTOSIS = -1.641
0.0 0.10000 0.20000 0.30000 0.40000 0.50000 0.60000 0.70000 0.80000 0.90000 1.00000

Appendix 2--continued



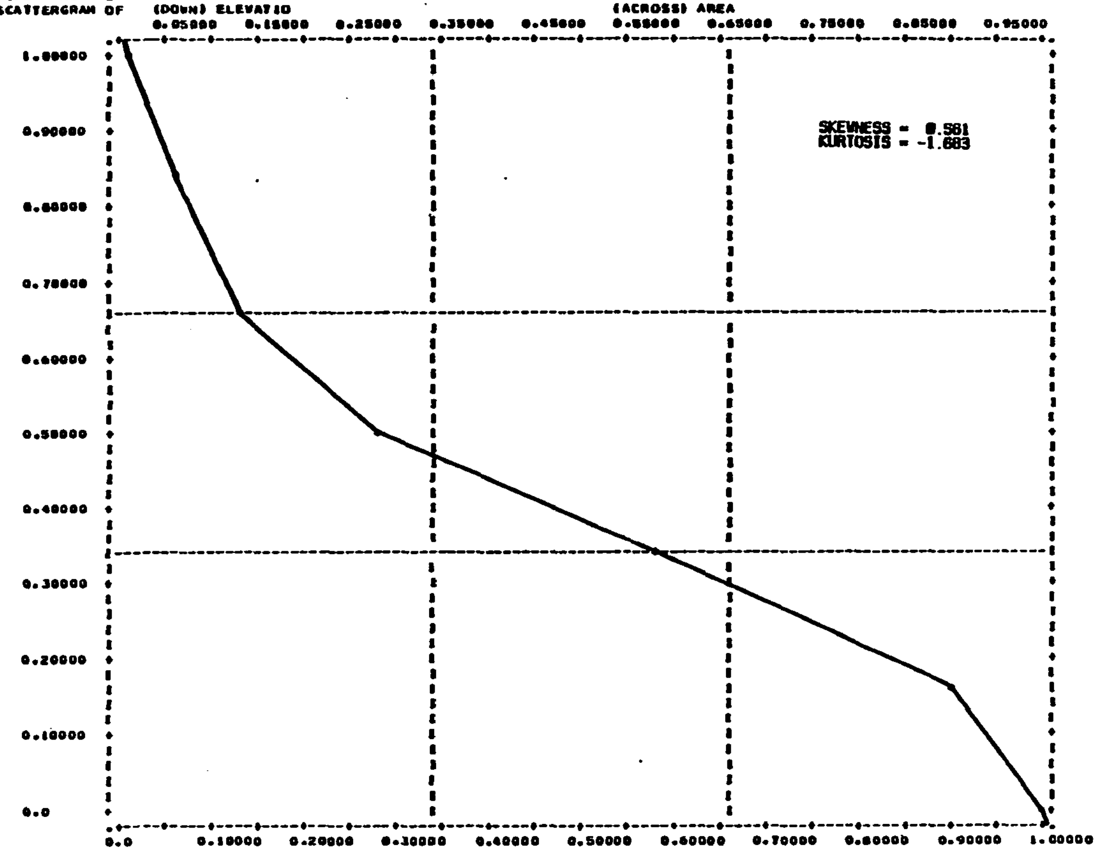
Appendix 2--continued



Appendix 2--continued

THE HYPSOMETRIC CURVE OF WEST FORK OF SALT CREEK (SHALE) 06/22/82 PAGE 5

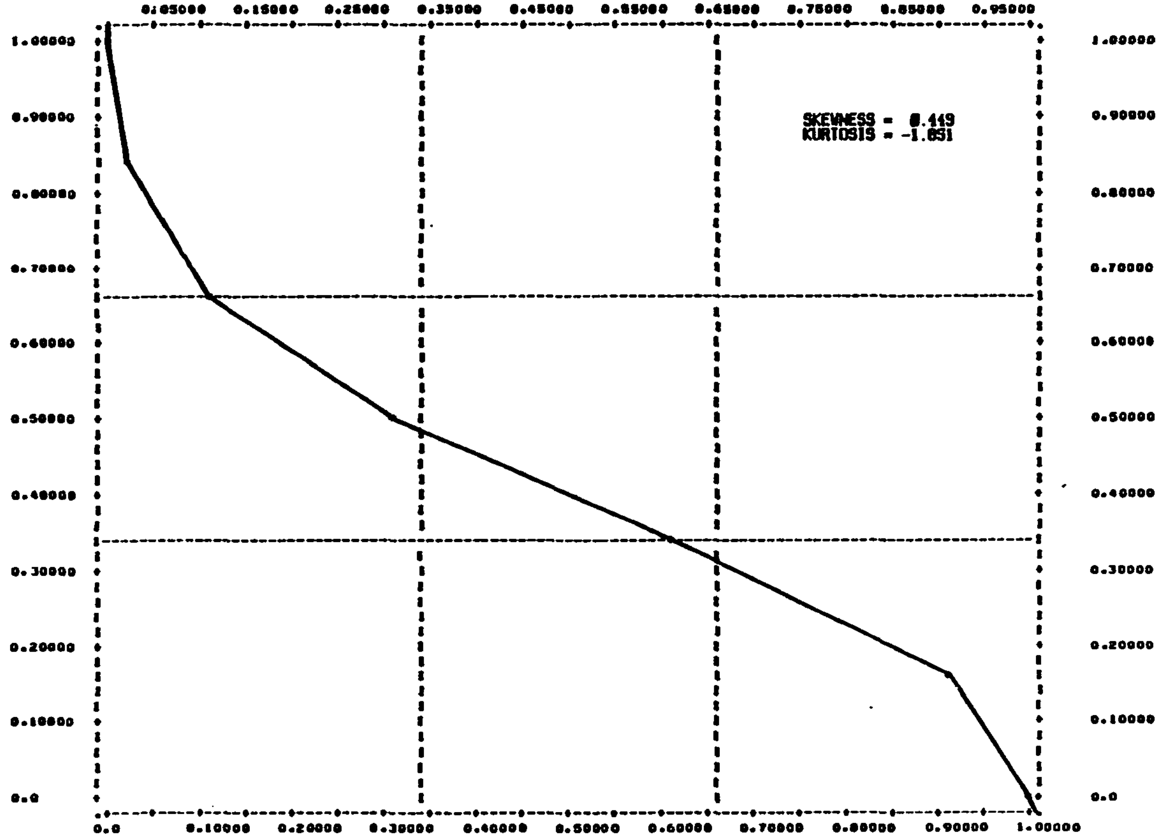
FILE NONAME (CREATION DATE = 06/22/82)
SCATTERGRAM OF (DOWN) ELEVATIO (ACROSS) AREA



Appendix 2--continued

THE HYPSOMETRIC CURVE OF SALT CREEK (SHALE) 06/22/82 PAGE 5

FILE NONAME (CREATION DATE = 06/22/82)
SCATTERGRAM OF (DOWN) ELEVATIO (ACROSS) AREA



APPENDIX 3

Procedure for Hydrometer Analysis

Procedure for Hydrometer Analysis

Calibrate the hydrometer as follows: Add 10 ml of 5% sodium hexametaphosphate to a 1000 ml settling cylinder. Add water bringing the solution to the 1000 line. Insert the hydrometer, and take the reading when the instrument stabilizes. Remove the hydrometer, and take the temperature (in °F) of the solution. For each degree above 69°F, add 0.2 to the hydrometer reading to determine the standardized correction factor. If the temperature is less than 69°F, subtract 0.2 from the reading for each degree below 69°F to arrive at the correction factor.

An over-drived sample should be sieved in #10 mesh sieve. Place 70 grams from the sediment passed the #10 sieve in a baffled dispersion cup, adding 100 ml. of 5% sodium hexametaphosphate and 400 ml of water. Allow this mixture to stand at least 10 minutes. (Sodium hexametaphosphate complexes or 'ties up' all negatively charged exchange sites on the clay colloids and insures that all sites have the same charge. The clays consequently repel each other rather than floculating.)

To further disaggregate the soil particles, mix the suspension for 5-10 minutes with the motor mixer. Then carefully transfer the mixture to a sedimentation cylinder using a wash bottle as necessary. Fill the cylinder to the 1000 ml line with water.

Place your hand <u>firmly</u> over the cylinder opening, and rock the cylinder back and forth repeatedly to mix the suspension completely. Return the cylinder to the upright position. After 20 seconds, lower the hydrometer into the suspension. At exactly 40 seconds, record the hydrometer reading.

Take the temperature, and add 0.22 to the hydrometer reading for each degree above 69°F, or subtract 0.2 for each degree under 69°F. Then subtract the original hydrometer correction factor (from first step; calibrating the hydrometer) from this adjusted reading.

Thoroughly mix the suspension again, and take another reading at exactly 2 hours. Adjust this reading for temperature as above, and subtract the hydrometer correction factor.

The adjusted 40-second reading gives the concentration of silt and clay in grams per liter. In other words, the percentage of silt and clay is the adjusted 40-second reading, divided by 70 grams, times 100, i.e.,

$$\text{percent silt and clay} = \frac{\text{adjusted 40-second reading}}{\text{70 grams}} \times 100$$

Note that 50 minus this combined percentage gives percent sand.

The percentage of clay is determined in the same manner using the 2-hour reading. (Only clay remains in suspension after 2 hours.)

$$\text{percent clay} = \frac{\text{adjusted 2-hour reading}}{\text{70 grams}} \times 100$$

Percent silt is determined by subtracting from 100, the combined percentage of sand and clay, i.e.

percent silt = 100 - (percent sand + percent clay)

However, a table was designed by the civil engineering department at the University of Oklahoma and was used to tabulate the hydrometer readings. The percentage of silt and clay in the sediment was calculated at the 40 second reading assuming that a diameter of 0.10 is the boundary between sand and silt and clay. The choice of a 0.10 mm particle diameter as a boundary was based upon the lithology in the area ranging from sand to silty sand in the Permian beds.

Schumm (1960) calculated the weighted mean percentage of silt-clay for each cross-section as:

$$(M = \frac{\begin{array}{c}\text{percent silt-clay}\\ \text{in channel bed}\end{array} \times \begin{array}{c}\text{channel}\\ \text{width}\end{array} + \begin{array}{c}\text{percent silt-clay}\\ \text{in channel bank}\end{array} \times 2 \begin{array}{c}\text{channel}\\ \text{depth}\end{array}}{\text{width} + 2\ \text{depth}}$$

as a parameter to express the character of the sediment composing the channel perimeter.

A SAMPLE OF

HYDROMETER ANALYSIS TABULATED RESULTS

Test by M. D. Radi

Soil Sample No. Kickapoo Creek (Sandstone) Sec. #11 banks sample

Wt. of original specimen (W) = 70 gr.

Hygroscopic moisture content (M) =

Wt. of dry soil specimen $W = \frac{100W_1}{100+1}$

Date 5-11-82 Time 10:00 PM

Dispersing Agent Calgon

Specific gravity of soil (G) = 2.65

Specific gravity correction (a) = $1.65G/_{2.65(G-1)} = 1$

Time Minutes	Hydro. Rend.	Temp. of	Disp. Agt Correct.	Corr. Hyd. Read.	Max. Part. Dia. mm.	Classification according to U.S.D.A.	Particle Diam. Correction Sp. Grav. T II	Temp. T III	Hyd. Rdg. T IV	Correct Particle diam. mm.	% Soil Adjustment in Suspension %Passing
T	R_1	t	C_1	R	d_1		C_2	C_3	C_4	d	P
40 sec	30	73	-5.8	24.20	0.10	Fine Sand	1.00				34.57
2	22	73	-5.8	16.20	0.041	Very Fine Sand	1.00				23.14
5	18	73	-5.8	12.20	0.026	Silt	1.00				17.42
15	15	73	-5.8	9.20	0.015	Silt	1.00				13.14
30	13	73	-5.8	7.20	0.011	Silt	1.00				10.28
60	12	73	-5.8	6.20	0.0074	Silt	1.00				8.85
250 or 120	11	73	-5.8	5.20	0.0037	Silt	1.00				7.43
1440 or 250	11	73	-5.8	5.20	0.0015	Clay	1.00				7.43

$R = R_1 + C_1$: $d = d_1 \times C_2 \times C_3 \times C_4$: $P = (R.\ a.\ 100)/W$

CORRECTION TABLES

Table I
Hyd Rdg Corr for "Calgon"

T	CR
65	-7.6
66	-7.4
67	-7.2
68	-6.9
69	-6.7
70	-6.5
71	-6.3
72	-6.1
73	-5.8
74	-5.6
75	-5.4
76	-5.2
77	-4.9
78	-4.7
79	-4.5
80	-4.3
81	-4.1
82	-3.8
83	-3.6
84	-3.4

C_1

Table II
Specific Gravity Corr

S.G.	Kg
2.50	1.050
2.55	1.030
2.60	1.015
2.65	1.000
2.70	.985
2.75	.970
2.80	.960
2.85	.950
2.90	.935
2.95	.925

C_2

Table III
Temperature Corr

Temp	Kt
65	1.020
66	1.013
67	1.007
68	1.000
69	.992
70	.984
71	.977
72	.971
73	.965
74	.960
75	.955
76	.950
77	.945
78	.940
79	.935
80	.925
81	.920
82	.915
83	.910
84	.905

C_3

Table IV
Hydr Level Corr

Hydr.	Kl
0	1.000
5	.973
10	.947
15	.920
20	.895
25	.867
30	.840
35	.812
40	.787
45	.757
50	.726
55	.694
60	.661

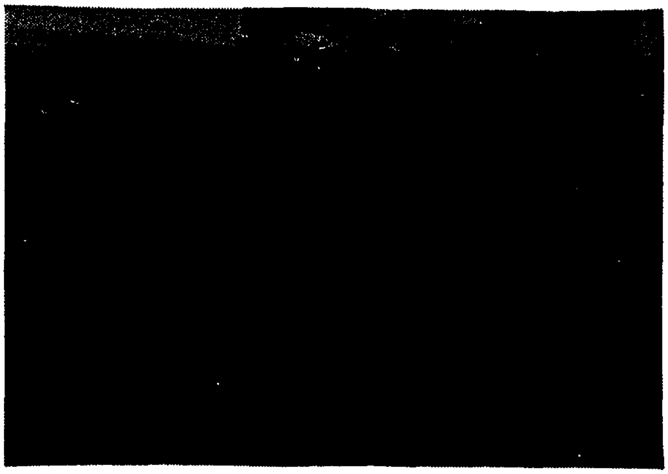
C_4

APPENDIX 4

Treatment of the Picture Distortion

APPENDIX 4

Treatment of the Picture Distortion



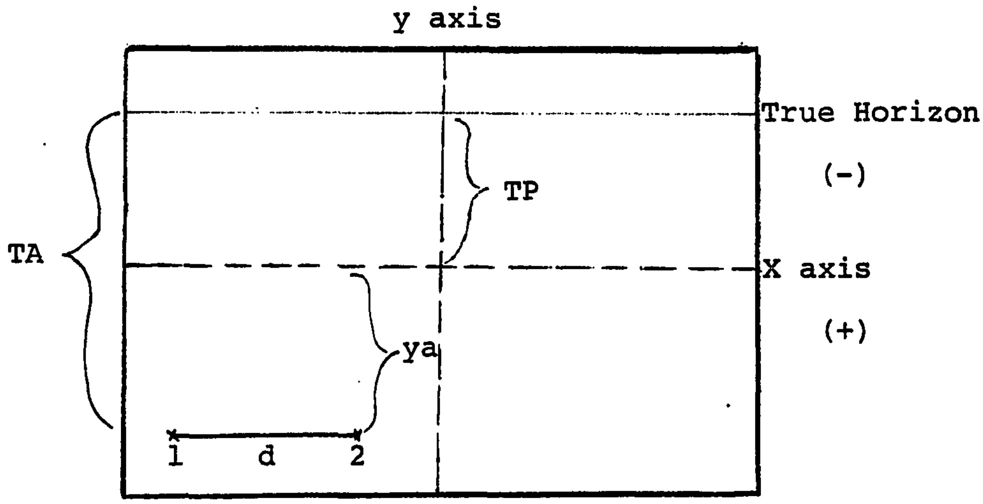
Cross-section #12 on Kickapoo Creek
Picture taken in August 1980.

$$V = \rho_1(T) + \rho_2(G)$$

Where V is the vegetal retardance index, ρ_1 and ρ_2 are correction factors, T is the percent cover of trees, and G is the percent cover of grass.

A quadrate is drawn covering the adjacent areas to the cross-section site on the flood plain and upstream and downstream directions. The horizontal lines (4 lines) are belt transects that the percent vegetation cover that touches each line will averaged by the four horizontal lines within the quadrate (Goudi, 1981, p. 269).

The distortion in the picture was treated mathematically by calculating the ground length of the lines transects as follows (Harold Malde, U.S.G.S. Journal of Research, vol. 1, 1973, p. 193).



The Picture

d = the distance between points 1 and 2 on the photo.

D = the distance between points 1 and 2 on the ground.

TA = the distance between the true horizon and the line d on the photo.

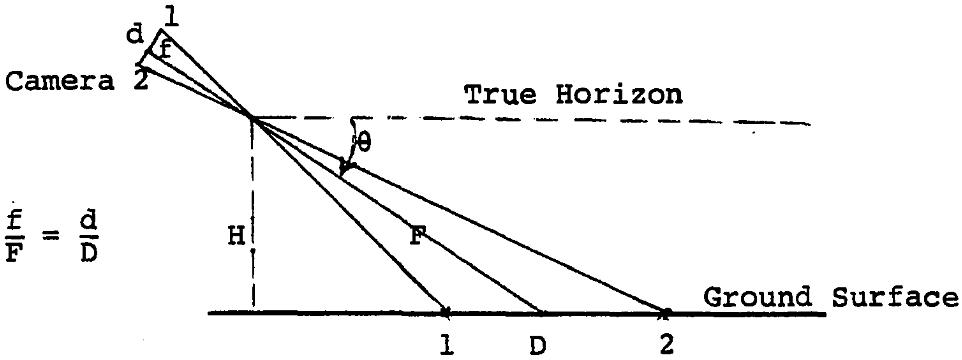
TP = the distance between the true horizon and x axis.

ya = the distance between the x axis and the line d.

In order to calculate D the following equation is used:

$$D = \frac{d\ H}{TA\ \cos\theta}$$

$$TA = f \tan\theta \pm ya$$



$\frac{f}{F} = \frac{d}{D}$

Where f is the focal distance or the distance between the film and the lines, F is the distance between the camera and the ground surface, H is the height of the camera which in this case is the eye-level height of the person taking the picture, and θ is the angle between the true horizon and a line running between the center of the line on the picture and the center of the line on the ground.

Example:

d = 5 inches (which is the length of the first line transects of the foreground on the picture)

D = ?

TP = 1 1/8 inches

TA = 2 inches

ya = 3/4 inches

H = 5'8" = 68 inches

θ = 11° 38'

f = 35 mm = 1.378 inches

$$D = \frac{d\ H}{TA\ \cos\theta} = \frac{5(68)}{2(0.999)} = \frac{340}{1.998} = 170.1702 \text{ inches}$$

$= 14.181$ feet

APPENDIX 5

Estimate and Proof of the Correction Parameters in the Vegetal Retardance Index Equation

APPENDIX 5

Estimate and Proof of the Correction Parameters in the Vegetal Retardance Index Equation

As indicated in Chapter I that cross-section areas of the stream channel holds a certain relationship with the ratio between tractive force and vegetal retardance index, this relationship is written as:

$$C_1 = a + b \frac{\tau_o}{V_1}$$

$$C_2 = a + b \frac{\tau_o}{V_2}$$

$$C_3 = a + b \frac{\tau_o}{V_3}$$

$$\vdots$$

$$C_n = a + b \frac{\tau_o}{V_n}$$

where C is the cross-section area, a and b are the regression constants, τ_o is the tractive force, and V is the retardance index of vegetation. The least square method is applied to determine ρ_1 and ρ_2 using an iterative values. Since the need is to estimate ρ_1 and ρ_2 in V ($V = \rho_1(T) + \rho_2(G)$, See Chapter IV), tractive force (τ_o) must be held constant. The following observations were chosen from the data with equal tractive force value:

Stream and Cross-Section #	C	τ_o	T	G
D-2	36	8.536	20	63
W-8	325	8.112	68	7
W-10	341	8.923	64	2
S-9	210	8.861	2	35
S-12	336	8.361	60	3
W-3	100	8.000	35	45
W-11	468	8.000	10	25
K-3	30	8.000	40	45
W-2	76	8.000	24	40

Where C is the cross-section area, τ_o is the tractive force, T is the percent coverage with trees, and G is the percent coverage with grass. Eleven iterative values were given for ρ_1 and ρ_2 as follows:

	ρ_1	ρ_2
V1	0.1	0.9
V2	0.2	0.8
V3	0.25	0.75
V4	0.3	0.7
V5	0.35	0.65
V6	0.4	0.6
V7	0.5	0.5
V8	0.6	0.4
V9	0.7	0.3
V10	0.8	0.2
V11	0.9	0.1

Then the possible values for V with different ρ_1 and ρ_2 are:

V1	V2	V3	V4	V5	V6
58.70	54.40	52.25	50.10	47.95	45.80
13.10	19.20	22.25	25.30	28.35	31.40
8.20	14.40	17.50	20.60	23.70	26.80
31.70	28.40	26.75	25.10	23.45	21.80
8.70	14.40	17.25	20.10	22.95	25.80
44.00	43.00	42.50	42.00	41.50	41.00
23.50	22.00	21.25	20.50	19.75	19.00
44.50	44.00	43.75	43.50	43.25	43.00
38.40	36.80	36.00	35.20	34.40	33.60

V7	V8	V9	V10	V11
41.50	37.20	32.90	28.60	24.30
37.50	43.60	49.70	55.80	61.90
33.00	39.20	45.40	51.60	57.80
18.50	15.20	11.20	8.60	5.30
31.50	37.20	42.90	48.60	54.30
40.00	39.00	38.00	37.00	36.00
17.50	16.00	14.50	13.00	11.00
42.50	42.00	41.50	41.00	40.50
32.00	30.40	28.80	27.20	25.60

And, the channel cross-section area (C) was regressed against V1 through V11 eleven times. The results are:

ρ_1 and ρ_2			
0.1 & 0.9	r_{cv1} - 0.80 R^2 0.64	$\text{Log}_{10} C = 3.72 - 1.12 \text{Log}_{10} V1$	F = 12.71
0.2 & 0.8	r_{cv2} - 0.89 R^2 0.79	$\text{Log}_{10} C = 4.83 - 1.83 \text{Log}_{10} V2$	F = 27.17
0.25 & 0.75	r_{cv3} - 0.93 R^2 0.86	$\text{Log}_{10} C = 5.51 - 2.29 \text{Log}_{10} V3$	F = 44.01
**0.3 & 0.7	r_{cv4} - 0.95 R^2 0.89	$\text{Log}_{10} C = 6.15 - 2.71 \text{Log}_{10} V4$	F = 60.46
0.35 & 0.65	r_{cv5} - 0.92 R^2 0.85	$\text{Log}_{10} C = 6.54 - 2.96 \text{Log}_{10} V5$	F = 42.67
0.4 & 0.6	r_{cv6} - 0.85 R^2 0.73	$\text{Log}_{10} C = 6.43 - 2.87 \text{Log}_{10} V6$	F = 19.11
0.5 & 0.5	r_{cv7} - 0.60 R^2 0.36	$\text{Log}_{10} C = 4.97 - 1.87 \text{Log}_{10} V7$	F = 4.00
0.6 & 0.4	r_{cv8} - 0.36 R^2 0.12	$\text{Log}_{10} C = 3.53 - 0.92 \text{Log}_{10} V8$	F = 1.04
0.7 & 0.3	r_{cv9} - 0.19 R^2 0.04	$\text{Log}_{10} C = 2.73 - 0.38 \text{Log}_{10} V9$	F = 0.28
0.8 & 0.2	r_{cv10} - 0.09 R^2 0.009	$\text{Log}_{10} C = 2.39 - 0.15 \text{Log}_{10} V10$	F = 0.07
0.9 & 0.1	r_{cv11} - 0.02 R^2 0.0007	$\text{Log}_{10} C = 2.21 - 0.15 \text{Log}_{10} V11$	F = 0.01

The same method was tested by another group of observation with a constant tractive force value.

Stream and Cross-Section #	C	τ_o	T	G
K-7	24	5.004	8	70
K-12	30	5.709	21	65
B-5	145	5.148	68	12
W-5	85	5.803	70	8
W-6	114	5.129	58	7

Then, possible values for V with different ρ_1 and ρ_2

V1	V2	V3	V4	V5
63.80	57.60	54.50	51.40	45.20
60.60	56.20	54.00	51.80	50.40
17.60	23.20	26.00	28.80	34.40
14.20	20.40	23.50	26.60	32.80
12.10	17.20	19.75	22.30	27.40

V6	V7	V8	V9	V10
39.00	32.80	26.60	20.40	14.20
43.00	38.60	34.20	29.80	25.40
40.00	45.60	51.20	56.80	62.40
39.00	45.20	51.40	57.60	63.80
32.50	37.60	42.70	47.80	52.90

And, the channel cross-section area (C) was then regressed against V1 through V11 eleven times. The results are:

ρ_1 and ρ_2		
0.1 & 0.9	r_{cv1} - 0.32	$Log_{10} C = 2.239 - 0.269 Log_{10} VI$
0.2 & 0.8	r_{cv2} - 0.93	$Log_{10} C = 3.74 - 1.31 Log_{10} V2$
0.25 & 0.75	r_{cv3} - 0.93	$Log_{10} C = 4.16 - 1.57 Log_{10} V3$
**0.3 & 0.7	r_{cv4} - 0.94	$Log_{10} C = 4.78 - 1.94 Log_{10} V4$
0.4 & 0.6	r_{cv6} - 0.83	$Log_{10} C = 6.285 - 2.84 Log_{10} V5$
0.5 & 0.5	r_{cv7} - 0.47	$Log_{10} C = 7.71 - 3.73 Log_{10} V6$
0.6 & 0.4	r_{cv8} - 0.73	$Log_{10} C = -5.06 + 4.29 Log_{10} V7$
0.7 & 0.3	r_{cv9} - 0.90	$Log_{10} C = -2.36 + 2.60 Log_{10} V8$
0.8 & 0.2	r_{cv10} - 0.93	$Log_{10} C = -0.85 + 1.66 Log_{10} V9$
0.9 & 0.1	r_{cv11} - 0.94	$Log_{10} C = -0.007 + 1.15 Log_{10} V10$

Moreover, the same method was tested by a third group of observations with another constant tractive force value:

Stream and Cross-Section #	C	τ_o	T	G
K-2	27	11.063	55	38
D-1	72	11.256	80	5
K-9	105	11.225	69	10
S-5	264	11.906	75	15
S-7	276	11.306	62	8

The possible values for V with different ρ_1 and ρ_2 are:

V1	V2	V3	V4	V5	V6
39.70	41.40	42.25	42.675	43.10	44.86
12.50	20.00	23.75	25.625	27.10	35.00
15.90	16.80	24.75	26.225	27.70	33.60
21.00	27.00	30.00	31.500	33.00	39.00
13.40	18.80	21.50	22.850	24.20	29.00

V7	V8	V9	V10	V11
46.50	48.20	49.90	51.60	53.30
42.50	50.00	57.50	65.50	72.30
39.50	45.40	51.30	57.20	63.10
45.00	51.00	57.00	63.00	69.00
35.00	40.40	45.80	51.20	56.60

Then the results of regressing channel cross-section area against V1 through V11 are:

ρ_1 and ρ_2		
0.1 & 0.9	r_{cv1} - 0.59	Log_{10} C = 3.49 - 1.23 Log_{10}V1
0.2 & 0.8	r_{cv2} - 0.67	Log_{10} C = 4.23 - 1.63 Log_{10}V2
0.25 & 0.75	r_{cv3} - 0.67	Log_{10} C = 5.50 - 2.41 Log_{10}V3
0.275 & 0.725	r_{cv4} - 0.67	Log_{10} C = 5.92 - 2.65 Log_{10}V4
**0.3 & 0.7	r_{cv5} - 0.68	Log_{10} C = 6.37 - 2.92 Log_{10}V5
0.4 & 0.6	r_{cv6} - 0.66	Log_{10} C = 8.41 - 4.097 Log_{10}V6
0.5 & 0.5	r_{cv7} - 0.57	Log_{10} C = 9.99 - 4.92 Log_{10}V7
0.6 & 0.4	r_{cv8} - 0.35	Log_{10} C = 8.11 - 3.64 Log_{10}V8
0.7 & 0.3	r_{cv9} - 0.05	Log_{10} C = 2.99 - 0.55 Log_{10}V9
0.8 & 0.2	r_{cv10} - 0.16	Log_{10} C = -0.41 + 1.39 Log_{10}V10
0.9 & 0.1	r_{cv11} - 0.31	Log_{10} C = -2.13 + 2.32 Log_{10}V11

APPENDIX 6

Photos for Cross-Sections Sites

SANDSTONE FORMATION

KICKAPOO CREEK

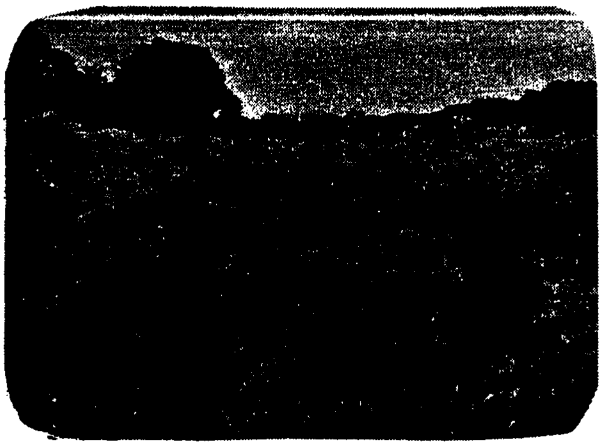
1

2

3

4

5



6

KICKAPOO CREEK

7

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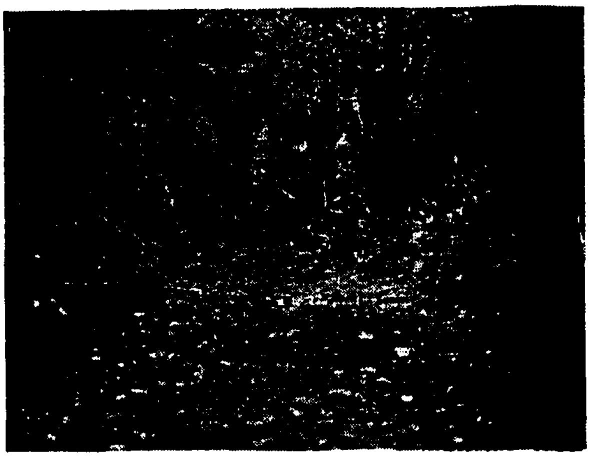
11

13

SANDSTONE FORMATION

DEVIL'S CANYON CREEK



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2

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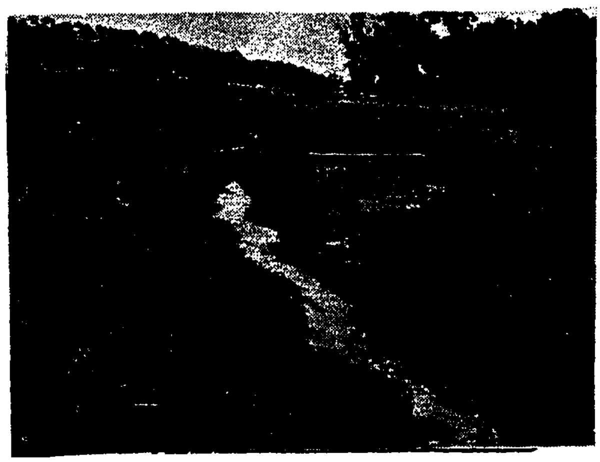

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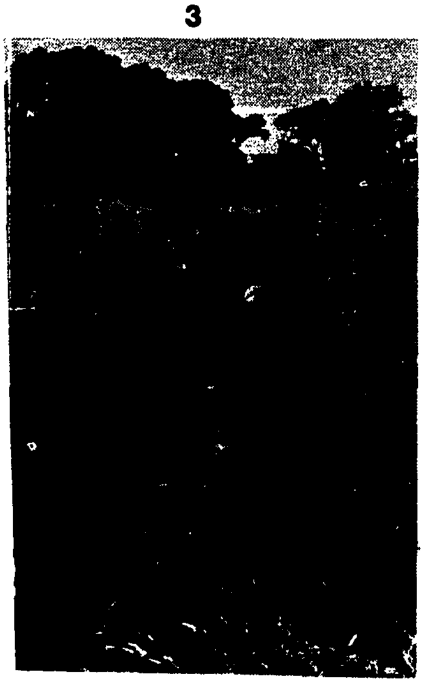

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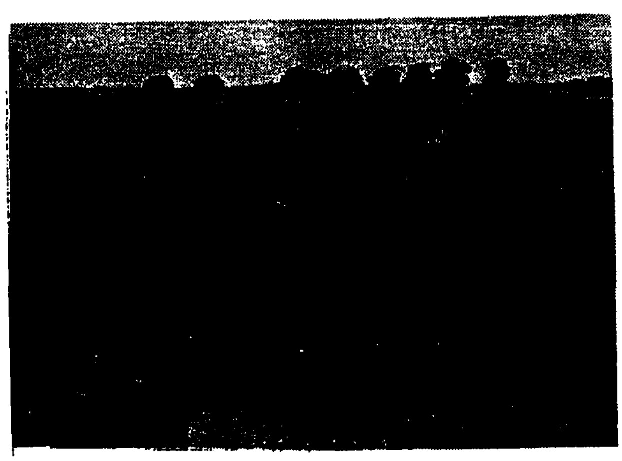


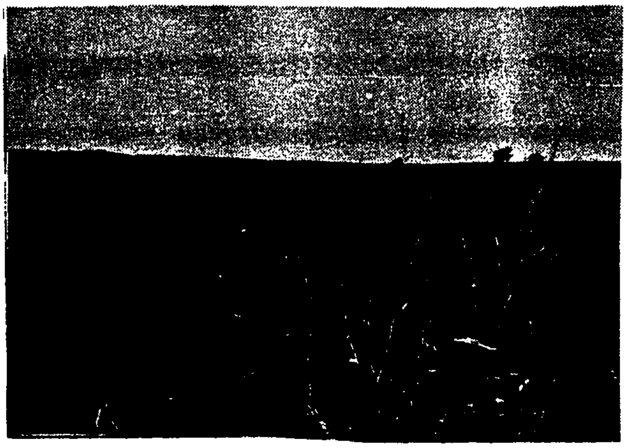
7

DEVIL'S CANYON CREEK

8

9

10

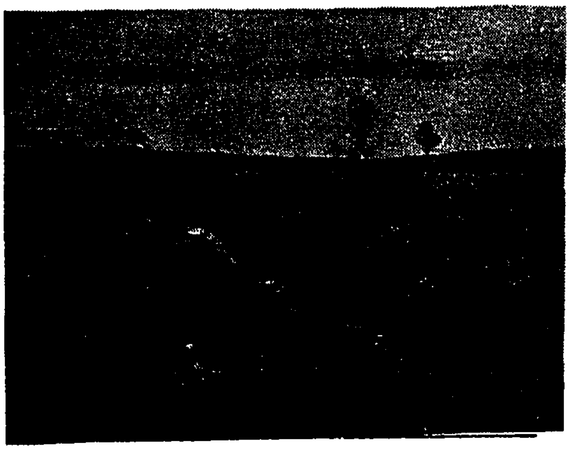


11

SANDSTONE FORMATION

BUGGY CREEK

1



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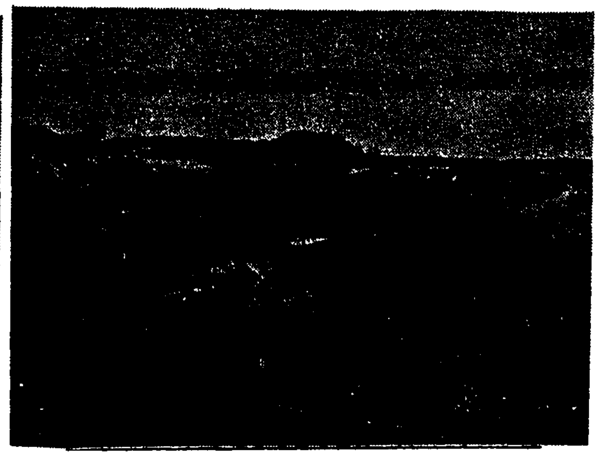
BUGGY CREEK

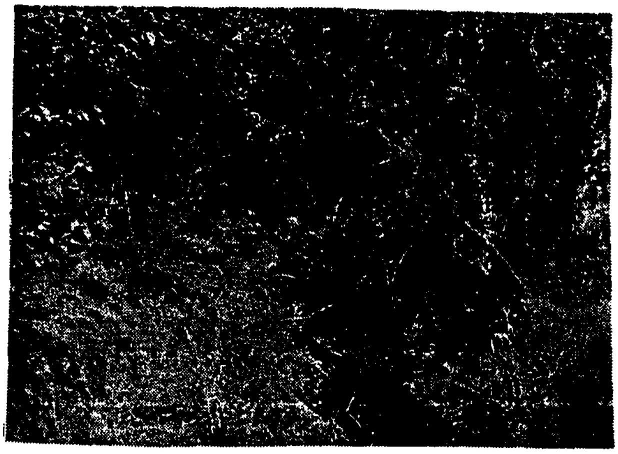
7



8

SHALE FORMATION
WEST FORK CREEK

1



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WEST FORK CREEK

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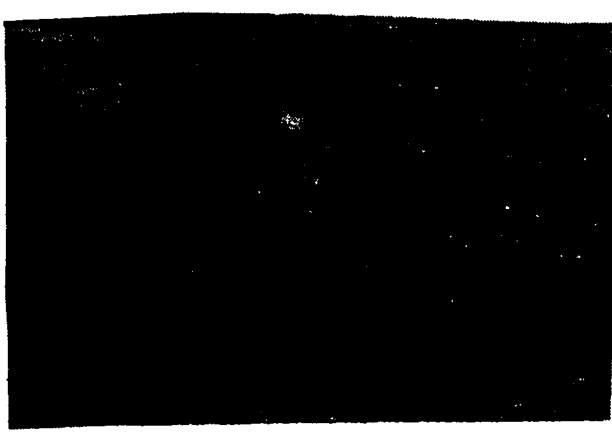
10



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WEST FORK CREEK

13

14

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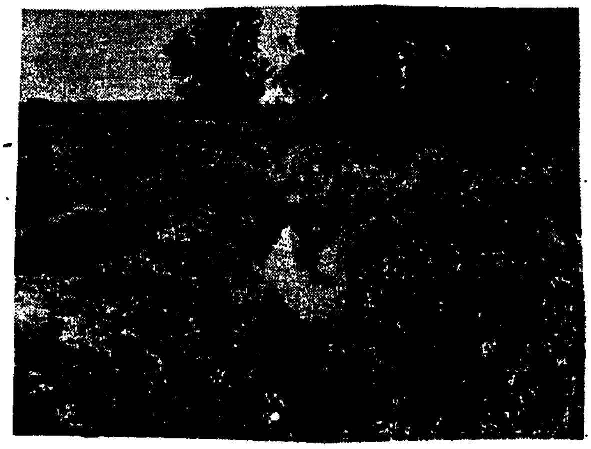
16

SHALE FORMATION
SALT CREEK

1

2



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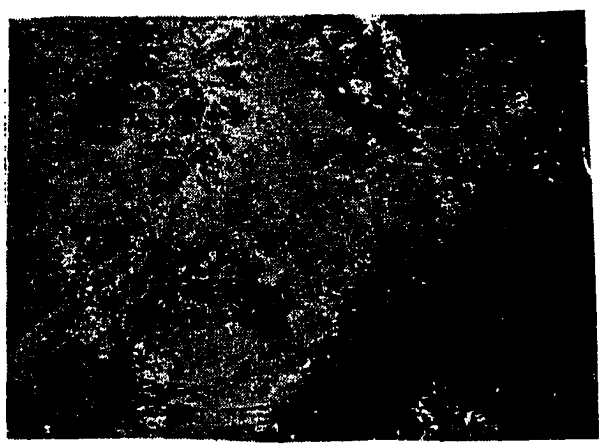

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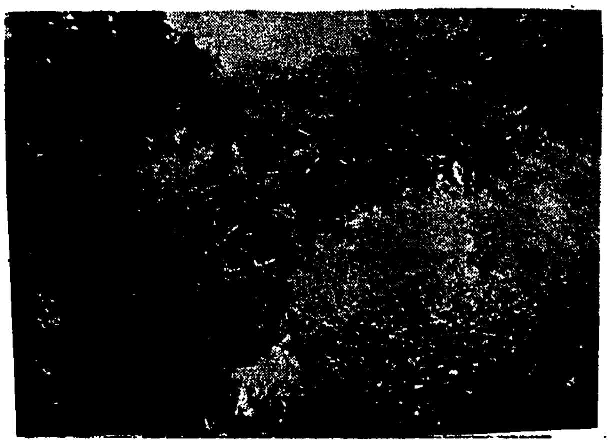
SALT CREEK

7



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SALT CREEK

13

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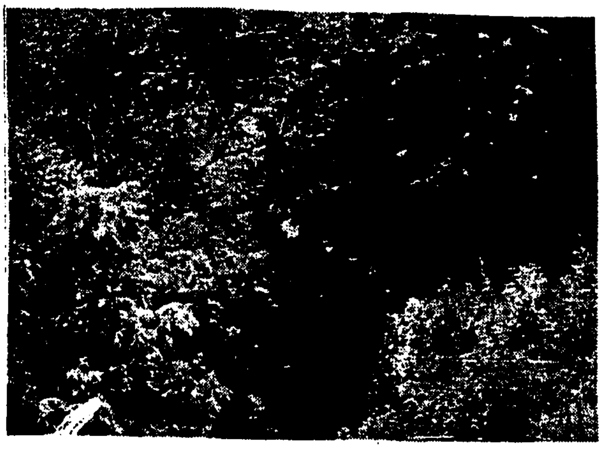

16