Recent fluvial geology in western North Dakota

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RECENT FLUVIAL GEOLOGY
IN
WESTERN NORTH DAKOTA

by
Thomas M Hamilton

B. S. in Geology, Capital University 1965

A Thesis
Submitted to the Faculty
of the
University of North Dakota
in partial fulfillment of the requirements
for the Degree of
Master of Science

Grand Forks, North Dakota
June
1967
This thesis submitted by Thomas M Hamilton in partial fulfillment of the requirements for the Degree of Master of Science in the University of North Dakota is hereby approved by the Committee under whom the work has been done.

Chairman

Dean of the Graduate School
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ABSTRACT

The thesis here abstracted was written under the direction of Lee Clayton and approved by Walter L. Moore and John R. Reid as members of the examining committee, of which Mr. Clayton was chairman.

In western North Dakota many drainage channels that are incised into valley fill are characterized by steep, unvegetated sides. Many such channels contain actively eroding scarps. The origin of these scarps is most frequently related to increased water velocity causing accelerated erosion on a steepened reach of the valley flat.

The principal mechanism of scarp migration is soilfall which is initiated by undercutting of the scarp face. Groundwater seepage at the base of the scarp is effective in reducing the cohesion of the sediment, thereby increasing the erodibility. In a channel with several successive scarps, the height of these scarps becomes progressively greater downstream, indicating a relationship to discharge.

Three alluvial units and two paleosols within the recent valley fill have been identified. By carbon-14 dating it was established that the upper 15 to 20 feet of this sediment was probably deposited after A. D. 1775. An average rate of sedimentation on the order of 0.3 to 0.4 feet per year is believed to be characteristic of these
units.

The modern cycle of erosion and deposition in western North Dakota probably began in the middle or late 1800's. Many of the gullies were partially or entirely filled during the deposition of the upper alluvial unit and new gullies have been trench ed to their present depth since approximately 1936.

The periods of valley filling in western North Dakota were associated with periods of sub-normal precipitation. Modern precipitation records indicate that such a period occurred from 1917 to 1936 during which the upper unit was deposited
INTRODUCTION

Purpose

The rapid erosion and deposition occurring in western North Dakota is a response to changes in the water-sediment balance which is related to fluctuations in precipitation. Features and geometry believed to be involved in this balance and found associated with several channels in western North Dakota were studied in an attempt to better define the factors associated with each. The primary objective of the project was to determine the effects and history of groundwater seepage on erosion occurring within these channels. The recent erosional and depositional history of the basins was determined by variations in lithology and organic accumulations and by carbon-14 dating.

Acknowledgments

I wish to thank Dr. Lee Clayton, University of North Dakota geology department for his assistance, both in the field and in the writing of this report. Other members of the committee who provided helpful assistance are Drs. Walter L. Moore and John R. Reid of the University of North Dakota geology department. I am also grateful to Samuel S. Harrison for his help with the laboratory analysis of erosional processes and to Michael E. Young for his
assistance in the field. Laboratory assistance was also
given by Chester F. Royse and Edward T. Callender. I am
also indebted to the U. S. Park Service for permission and
assistance in conducting field work within the boundaries
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North Dakota Water Resources Research Institute with funds
provided by the U. S. Department of Interior, Office of

Location and Description of the Area

The drainage basins discussed in this report are in
western North Dakota in Billings, McKenzie, and Mountrail
Counties (fig. 1). Five of the channels studied are
tributaries to the Little Missouri River, the sixth is a
tributary to the Missouri River. The Little Missouri River
presently drains into the Garrison reservoir formed by the
damming of the Missouri River.

The streams studied are typical of the many ephemeral
streams in the Badlands of western North Dakota. Their
basins are cut into poorly consolidated Tertiary bedrock;
one is mantled by Pleistocene till. Their channels range
in size from tiny rills near the divides to channels a few
feet to a few tens of feet wide and deep which are incised
into valley fill.

The vegetative cover in the area ranges from dense,
on some northfacing slopes, to sparse on the most of the
Figure 1.--Location map of study area. The letters (A through F) indicate the approximate location of drainage basins studied.
hillslopes and alluvial surfaces. The most conspicuous vegetation present on the alluvial surfaces in McKenzie and Billings Counties is sagebrush, particularly of the species *Artemisia tridentata* and *Artemisia cana* (Stevens, 1950, p. 291). The trees, if present, are usually cottonwoods or aspen. The grass cover is light in most areas, and consists predominately of short prairie grasses.

**Climate**

Climatological data for western North Dakota shows a mean annual temperature near 42 degrees F. and an average yearly precipitation of about 15 inches. The difference between the yearly maximum and minimum temperatures can be as much as 170 degrees F. The closest available records to the area of study are from Beach and Dickinson, North Dakota.

Continuous weather records are sparse, but precipitation records from stations east and west of the Little Missouri River have been plotted (figs. 2 and 3) to show the fluctuations common to this area (U. S. Weather Bur. Records). Most of the yearly precipitation occurs between April and September, generally during thunderstorms. During some months of the record, no precipitation is recorded.

The annual precipitation is highly variable from year to year. The minimum annual precipitation was 6.21 inches at Beach in Golden Valley County (fig. 2) in 1934. Other stations in this part of the state have recorded amounts as low as 4 inches for a given year. The maximum annual precipitation from the available records was 31.2 inches at
Figure 2.--Annual precipitation at Beach, North Dakota (46° 55' N. Lat., 104° 01' W. Long.), 1906-1942.
Figure 3.--Annual precipitation at Dickinson, North Dakota (46° 53' N. Lat., 102° 48' W. Long.), 1892-1960.
Dickinson in Stark County (fig. 3) in 1941.

The 5-year moving averages show the general trends of the precipitation at Beach and Dickinson. The annual totals fluctuate from year to year, but the moving averages show that the periods of high and low precipitation are somewhat grouped, indicating a cyclic trend. The most evident period of low precipitation occurred roughly between the years 1917 and 1936, although some years of this period do show above normal precipitation. The later years of this period, were known as the "Dust Bowl Years". This drought resulted in a great reduction in the vegetative cover and severe blowing and drifting of the unprotected sediment (Bavendick, 1952). The pattern for the two records used is fairly consistent. No rainfall-intensity-frequency-duration curves are available for this part of the state, and those available for other areas cannot be considered as characteristic of the area of study (Stomme, H. G., State Climatologist, written communication, February 27, 1967). However, the extended period of low rainfall from 1917 to 1936 probably differed from the other more normal periods only in the frequency of occurrence of the precipitation and probably not greatly in intensity or duration of the individual storm. Runoff from the storage of winter snow is relatively unimportant in this part of the state. The mean potential evaporation at Dickinson is almost three times the mean precipitation (U. S. Weather Bur., 1965).
No previous investigations of gully erosion have been conducted in North Dakota, although this problem has been studied in other parts of the United States. The following discussion of gully erosion is a summary of some of the results and hypotheses of previous investigators.

The cause of gullying has been of interest, primarily because of the economic losses usually associated with this erosional process. Antevs (1952) compiled and summarized the various current hypotheses concerning the origin of gullies. Of these, two authors believed that the modern cycle of gullying in the western United States was associated with overgrazing; this cycle reportedly began on a large scale during the 1880's just after the large influx of ranchers beginning in about 1870 (Bailey, 1935, and Antevs, 1952). It was believed that overgrazing and trampling reduced or destroyed the vegetative cover causing greatly increased runoff and violently-erosive flash floods.

Other workers believed that the major factor in the origin of gullies was climatic change with overgrazing being only a contributing factor. Bryan (1925) believed that a climatic change toward lesser precipitation was responsible for the initiation of gullies. He postulated an increasing aridity which reduced the vegetation and promoted runoff, and which enlarged the magnitude and the erosive and transporting power of floods. This hypothesis was in opposition
to Bryan's (1922) earlier hypothesis that the change in climate was toward more humid conditions. Increased moisture was also thought to be the cause of gully ing by early workers such as Barrell (1908), and Gregory (1915 and 1917). They felt that an increase in moisture caused dense vegetation and longer, steadier streams with considerable erosive power in the valleys. Cattle trails were also believed to have served as the trigger to gully ing. Leopold (1951) felt that a change toward a greater intensity of the rainfall rather than in the total amount of water was the most important factor.

These hypotheses, though valid in many areas, fail to explain why no gullies formed in some areas of heavy over-grazing and why gullies formed in areas where no grazing has occurred (Schumm and Hadley, 1957).

More recent workers have placed greater emphasis on the basin geometry. Schumm and Hadley (1957) suggest that gullies formed only on steep gradient reaches within several valleys in New Mexico.

Brice (1966) noted the presence of steepened reaches of gullied valley flat in southern Nebraska. He also felt that the gullies might be initiated on these steepened reaches of the valley.

The mechanics of channel scarp migration have been most extensively studied in the laboratory, because direct field observation of scarp retreat is very difficult. Brush and Wolman (1960) attempted to maintain a nickpoint in nonco-
hesive sediment in a flume, but found that the pre-formed nickpoint quickly disappeared. They noted however, that the channel above the nickpoint became narrower and deeper with time while below the nick it became shallower and wider. They attributed the widening to increasing shear along the margins of the channel below the nickpoint. They also stated that seepage may be an important factor in channel scarp preservation and migration, though they gave no supporting evidence.

Groundwater seepage has been found to be the key element in the formation and migration of erosional scarplets on hillsides in the semiarid Great Plains (Hadley and Rolfe, 1956). The scarplets are developed in the surficial mantle overlying a fine-grained, impermeable weathered bedrock, where subsurface water reaches the surface and saturates the soil, making it readily susceptible to erosion by surface flow. A vertical face is maintained by water seepage at the base. They also found that there was a downslope migration of the sediment removed from the scarplet. Although no mention is made of seepage in conjunction with larger features, such as channel scarps, its importance might be inferred.

Leopold, Wolman, and Miller (1964) presented conclusions derived from field and laboratory studies concerning the maintenance of channel scarps. Their findings are summarized in the following list:

1. Nickpoints constructed in cohesive sediments are
maintained if the initial height of the nick is sufficient to create a plunge pool.

2. In order for the nickpoint to be maintained, the sediment must have a high resistance to shear (be cohesive).

3. The nick will be maintained only if the flow is sufficient to remove the sediment which accumulates at the base of the nick face.

4. If the depth of flow over the face is equal to or greater than the height of the face, the nickpoint will be obliterated.

5. Sapping at the base of the nickpoint is undoubtedly one of the more effective processes in headward extension.

6. Piping often occurs upstream from the nickpoint.

7. Slumping of the headwall at the base of the nickpoint occurs most commonly after a storm.

8. The erosive action of water over the vertical face of a gully head is generally not effective in causing headward erosion.

9. The depth of cutting of gullies is probably controlled by resistant layers.

Recent alluvium

Several terraces occur along the Little Missouri River, but little has been written about them. Laird (1950) described four terrace levels present along the Little Missouri River in the south unit of Roosevelt Park. The upper terrace was correlated with the advance of the Pleistocene ice sheet, which diverted the river in its lower reaches. But,
the ages of the lower terraces are still unknown. Howard (1960), in a study of the Cenozoic history of western North Dakota and eastern Montana, placed as age of Mankato to post-Mankato on some of the upper silts along the Missouri River. No mention was made of the age of the very recent alluvium, however.

The accumulation of sediments in present streams and valleys has aroused the interest of many workers. Leopold, Emmett, and Myrick (1966), in a study of ephemeral streams in New Mexico, measured the amount of sediment removed from hillslopes and the amount of sediment intransient in the streams; the alluvial units were dated by radiometric methods. Their findings, as well as those of Hadley and Schumm (1961), indicate that the sediment accumulation in small basins in the semiarid United States is relatively rapid.

The age of recent alluvium in basins in Wyoming has been established by Leopold and Miller (1954). They recognized the Lightning Formation, which consists of light-brown to tan sediment that is generally devoid of bedding. Deposition of this unit began about A.D. 1200 and ended about 1880. They note that no soil profile is observable on the surface of the Lightning Formation.

Brice (1966), in a study of the erosion and deposition in southern Nebraska, noted the presence of gullied recent alluvium, which is separated into a banded upper part and a less banded lower part. Carbon-14 dates indicate that much of this sediment has accumulated within the past 500 years.
Brice also notes the accumulation of from 5 to 6 feet of sediment on some valley flats between 1920 and 1953, as determined by the depth to which fence posts were buried.

Definition and Classification of Gullies

According to the most common usage, the essential features and processes associated with features called "gullies" are: the ephemeral transmission of flow, continuing erosion enlarging the channel, steepness of the channel walls, entrenchment into unconsolidated sediments, and its general size (larger than a rill). The term arroyo, as used by Bryan to describe vertical-walled, flat-floored channels of ephemeral streams of the semiarid Southwest, can be applied to channels having these qualifications, whether or not the long profile of the stream has pronounced vertical breaks, but the term arroyo cannot be applied to channels separated by lengths of ungullied alluvium (see following page).

The use of the term "nickpoint" or "knickpoint" will be avoided when referring to the vertical faces occurring in many of the gullies or arroyos, because this term has become extensively used when referring to any significant break in the gradient of a stream and not necessarily to the type described above. The term "channel scarp" or simply "scarp" will be applied to the vertical erosional faces occurring in a channel. This term will not refer to other erosional type scarps such as terrace scarps, step scarps, or hillside scarps. Use of the term "head cut" will also be avoided because the
application of this word to any channel scarp within the system has the connotation of being the headward terminus of that system. This will not be the case in a channel containing more than one channel scarp. The term "head cut" will be applied however, if the scarp is at the head of the gully. A scarp occurring in a tributary to a main gully will be referred to as a "side-channel scarp".

Gullies will be classified (Brice, 1966), on the basis of location (valley-bottom gullies, valley-head gullies, and valley-side gullies). This classification is rather arbitrary, because valley bottoms grade smoothly into valley heads and valley sides.

If more than one channel scarp exists along a single gully, it is called a discontinuous gully rather than an arroyo. The discontinuous gully or system of gullies displays characteristics common to new gullies (Leopold, Wolman, and Miller, 1964). This type of gully is characterized by a vertical channel scarp downstream from which the channel decreases in depth. Where the gully floor intersects or nearly intersects the original surface of the valley fill the height of the walls has decreased to zero and a fan occurs. This fan separates one channel scarp from another in the same valley (fig. 4).

The growth of a discontinuous gully system into a single, unbroken gully is explained by Leopold and Miller (1956). When the channel scarp of a downvalley gully reaches the lower edge of a fan of an upvalley gully, a stage of
Figure 4.—A composite sketch showing the elements of a discontinuous gully and their relationship.
coalescence is reached; after this stage the downvalley scarp will be cutting below the original surface of the valley fill. The discontinuous gullies integrate into a single gully as the downvalley scarp advances. This evolution is shown diagrammatically (fig. 5) as modified from Leopold and Miller (1956, p. 31).
Figure 5.-Longitudinal section showing the development of an arroyo from a discontinuous gully. Sketches modified slightly after Leopold and Miller (1956, p. 31).
METHODS OF INVESTIGATION

Drainage Basin Characteristics

Drainage basins were examined in the field and on aerial photographs. Factors pertinent to the study, such as width and depth, channel scarp spacing and height, and the general drainage characteristics of the basin were carefully recorded. Drainage area was determined with a compensating polar planimeter.

Comparison of aerial photographs taken of the study areas in 1938 and 1939 with photographs taken in 1958 made possible observations of channel change, the rate of channel scarp retreat, and gross estimates of the amount of alluvium removed from the basins. Results of these analyses are discussed in a later section.

Longitudinal Profiles

Longitudinal profiles of the dry channel bottoms were obtained by the use of a stadia-rod and a mounted level. A cord was measured at a distance which would permit clear reading of the stadia-rod and was used as a measure of the horizontal distance between stations.

An arbitrary starting point (datum) was selected for each profile run. The run was terminated at either the end of the stream or after a representative segment of the
stream had been traversed. The vertical change in feet at each station was recorded on a prepared data sheet.

After completion of the long-profile traverse of a given stream, the width and depth of the channel were measured at the established stations. Measurements of channel width were made with a steel tape and channel depth with a calibrated drop-line. The channel scarp heights were also measured with the steel tape.

Gaining and Losing Stream Segments

To ascertain the general groundwater flow pattern of the ephemeral streams in western North Dakota, investigations were made in basin A in Mountrail County, North Dakota (secs. 5 and 6, T. 152 N., R. 92 W.) (fig. 7). Following an early morning rain that yielded approximately 1 inch of precipitation the stream channel bottom was examined for evidence of either gaining water from the ground or losing water to the ground.

Following a procedure used by Lee Clayton, in a study of outwash channels in Alaska (oral communication), the gaining and losing segments were located by digging a pit (called a piezopit) in the channel bottom and noting whether the water runs into (losing) or out of (gaining) the pit. Those segments of the stream channel which were losing water required a much deeper pit to intersect the water table than did the gaining segments.

To establish the fluctuations of the groundwater level
through a channel scarp, a bank of piezometers was set at the base of a scarp. The segment of the channel upstream from the scarp was examined by means of auger holes dug with a 6-foot hand auger. The piezometers were set at measured depths of 4.0, 2.0, 1.5, 1.0 and 0.5 feet. To set the piezometers, a hole was augered to the required depth, the bottom of the hole was lined with gravel to prevent clogging, a plastic tube was inserted and the hole was backfilled with clay to minimize leakage.

The piezometers were ⅛-inch diameter plastic tubing, which were cut in measured lengths from a large roll. The tubes were open to the atmosphere and were therefore subject to some fluctuations due to changes in atmospheric pressure. This relationship is inverse; an increase in the atmospheric pressure results in a decrease in the head level (Todd, 1963).

The height to which the water level rises in the tube registers the potential that exists at the point in the flow system to which the tube has been inserted. Fluctuations in the head level is a measure of changes in piezometric potential and is directly related to direction (and amount or velocity) of groundwater movement. The piezometric head (fig. 6) is defined by the elevation to which water rises in a piezometer above some arbitrary datum; it is the hydrostatic head plus the hydrodynamic head plus the elevation head.

In an area of upward moving groundwater (discharge area) the piezometric-head levels in a series of piezometers will be greatest in the deepest piezometer and least in the
shallowest. In an area of downward moving groundwater (recharge area) the converse is true.

![Piezometric head diagram]

Figure 6.--A sketch showing the component parts of a piezometric head.

Sediment Sampling

Sediment samples were collected in all channels investigated. The samples from exposures characteristic of the alluvium of the area were placed in labeled sample bags. Care was taken not to take the sample from gravel stringers which are not characteristic of the sequence being sampled.

The samples to be analyzed in the laboratory were taken every 0.5-foot from the surface to the lowermost part of the exposure. A representative sub-sample was later split from the original sample and used for size and chemical analyses. The remainder of the sample was soaked and wet-sieved for
gastropods and other biological remains.

Sediment Sample Analysis

The samples were analysed in the laboratory to determine the chemical and physical properties of the alluvium.

The grain sizes (Wentworth Scale) present in the samples were determined by sieving and by standard pipette analysis. A representative fraction was split from each sample and dispersed in a solution containing 40 grams of calgon dissolved in one liter of distilled and de-ionized water. After dispersal using 100 ml of the calgon solution the sample was wet-sieved to separate any sand or gravel from the silt-clay fraction. The coarse fraction was dry-sieved using a set of Tyler-Standard Screens and the fine fraction was placed in a 1000 ml settling tube. The fine fraction was mixed in the settling tube and allowed to stand for 24 hours to check for flocculation. Withdrawals from the settling tube were made under a near-constant temperature of 68 degrees F. The withdrawal times and depths were calculated from Wadell's formula for the settling velocities of particles finer than 1/16 mm (Krumbein and Pettijohn, 1938). Each withdrawal was placed in a beaker. The beakers were placed in an oven and evaporated to dryness then removed and allowed to cool to room temperature. The weight of the contents of each beaker was obtained by the use of an analytical balance. The data were calculated using standard formulas.

The carbonate, in the form of CO₃, in each of the allu-
vial samples was determined by a volumetric analysis adapted from Herrin, Hicks, and Robertson (1958). Each sample was ground until it would pass through a 200-mesh Tyler sieve to insure a complete reaction between the sediment and the acid. From each powdered sample, exactly 1 gram was dissolved in 25 ml of 0.421 normal sulfuric acid (H₂SO₄). The acid-sediment combination was heated to approximately 90 degrees C. and maintained there for a period of 15 minutes to insure a complete reaction.

The solution was back-titrated to a pH endpoint of 8.3 using a 0.468 normal solution of sodium hydroxide (NaOH). The percent carbonate present in the sample is reported as CO₃. Every fifth sample was run twice to check the reproducibility of the method.

The determination of the percentage of oxidizable matter in each alluvial sample was by the Schollenberger Method as modified from Jackson (1958). After being ground and sieved, 1 gram of each sample was oxidized using 0.40 normal chromic acid (K₂Cr₂O₇) in the presence of an excess of sulfuric acid, with the external application of heat. After completion of the oxidation process the remaining acid was back-titrated with a 0.225 normal solution of ferrous ammonium sulfate [Fe(NH₄)₂(SO₄)₂ · 6H₂O], using a diphenyl indicator. Every 5th sample was run twice to check the reproducibility of the method.

The results (discussed in later section) are given as percentage of oxidizable matter. This includes the organic
matter present as well as certain oxides which are probably present in the sediment and which react with the oxidizer. Because no corrections were made for these interferences, the results could not be reported as organic matter though the values found approximate the organic content. The amount of oxidizable matter in each sample is a measure of the reducing capacity of the environment, which was the desired result.

Correlation Procedures

The correlation of alluvial units between separate basins was accomplished by comparison of color, relative thickness of units, grain-size, the presence of paleosols and by the repetition of the sequence. The source of the alluvial sediment in all the basins studied was primarily the Paleocene Fort Union Group and older alluvium. Some of the alluvial sediment in the basin in Mountrail County, however, was derived from glacial drift.

Laboratory Models

A laboratory study was made in the fall of 1966 to test early hypotheses made during the field study and to better establish the factors related to the origin and migration of channel scarps in cohesive sediment. The study was conducted by Samuel S. Harrison of the geology department at the University of North Dakota and by the author.

The flume used was 12.0 feet in length, 0.5 foot in
width, and 0.88 foot in depth. It was constructed of wood covered with fiberglass, and had one glass side for observation. The bottom of the flume was approximately level throughout all runs.

The sediment used was overbank material from the Turtle River in Grand Forks County, North Dakota. The sediment averaged 20 percent sand, 64 percent silt, and 16 percent clay, which was comparable with grain-size averages obtained from the alluvium of western North Dakota. The sediment was mixed to a "soupy" consistency, poured into the flume and allowed to settle and compact for two or three days prior to individual runs.

The surface water feed was tap-water from a 3/4-inch diameter hose, which was placed on the upper end of the sediment. A plastic sheet was placed under the hose and extended 0.6 foot downstream on the sediment surface to prevent scour.

For the initial investigations on the mechanisms of maintenance and migration of a channel scarp, a vertical face was cut in the lower end of the sediment prior to the individual run. During later experiments on channel scarp initiation, the sediment surface was relatively smooth and sloping.
RESULTS OF INVESTIGATIONS

Drainage Basin Characteristics

Data from six drainage basins in western North Dakota were collected during the study. Other small basins in the area were observed but no measurements were made. The most detailed study of processes and basin geometry was made in a drainage basin in Mountrail County (secs. 5 and 6, T. 152 N., R. 92 W.). The basin is designated as (A) (fig. 7). The measurements were made in an attempt to determine which drainage basin characteristics correlate most closely with water and sediment yield and gully development.

Basin A has an area of 2.24 square miles. The main stream segment drains westward into the Garrison Reservoir, which has flooded its lowermost reaches. The valley is cut into the sand, silt, and clay of the Paleocene Fort Union Group, and the upper reaches are mantled by Pleistocene glacial drift. Recent sediments in the basin include alluvium, loess, colluvium and slope wash material. The once smooth and undissected valley floor is now trenches by the stream.

In the basin 10 channel scarps at least 1 foot high and 7 side-channel scarps are present. These scarps occur in the upper 3/4 of the main drainage channel and in its tributaries on the south-facing slopes. The tributaries from the
Figure 7.--Map of drainage basin A (secs. 5 and 6, T. 152 N., R. 92 W.). P-P' indicates segment of channel that was measured. For resulting profile see figure 12.
north-facing slopes are inactive and are covered with grass and trees. These channels from the north-facing slopes have a gently rounded bottom, mantled by glacial drift which has a well-developed soil.

The channel scarps range in height from 1.5 to 13.5 feet with the height generally increasing downstream. The scarp occurring farthest downvalley is a 2.3 foot scarp. It is the youngest scarp in the gully having been initiated in the fill deposited from the next upvalley scarp. The spacing of the scarps varies but they are more closely spaced farthest upstream. Throughout most of its length, channel width is greater than depth.

This gully can be classified as a discontinuous gully because the channel has a series of vertical scarps which are actively eroding upstream and which are separated by reaches of ungullied valley. Downstream from each scarp is a vertical walled channel which decreases in depth until it intersects or nearly intersects the valley flat, at which point a fan is formed. The fan is usually higher in the center than at the margins causing the watercourse to split in time of flow, unless the entire surface is inundated.

The size of the fans is related to the height and spacing of their associated scarps. Small fan surfaces are associated with small scarps and larger fans are associated with larger scarps, which are in turn related to the position in the drainage system.

Comparison of the data obtained during the field
investigation with data obtained from aerial photographs taken of the basin in 1938 reveal striking changes. The most obvious change has been in the channel width; the channel has more than doubled in width since 1938. The amount of scarp retreat that has occurred since 1938 increases with the drainage area; greatest retreat is associated with those scarps located farthest downstream.

The other basins studied have a geologic and geomorphic setting similar to basin A. Basin B (secs. 27, 28, and 34, T. 148 N., R. 99 W.) located in the north unit of Roosevelt Park has a drainage area of about 0.80 square miles (fig. 8). The contained stream has gullied its valley fill and forms a trench in the No. 2 terrace of the Little Missouri River. The channel depth is greater than the width for approximately 1/2 mile upstream from the river. Only one channel scarp was observed in the arroyo, and this was formed where the channel intersected bedrock. The character of the channel bottom changed from flat-bottomed to V-shaped where bedrock was encountered.

Basin C (sec. 33, T. 148 N., R. 99 W.), also located in the north unit of Roosevelt Park has a drainage area of about 0.39 square mile (fig. 9). The contained stream has gullied its valley fill and forms a fan on the surface of the No. 2 terrace. Many bedrock slump blocks buried by the alluvium have been re-exposed by erosion. Where the channel crosses the bedrock blocks, it becomes V-shaped. Many sinkholes are present in the lower reaches of the valley in the ungullied
Figure 8.--Map of drainage basin B (secs. 27, 28, and 34, T. 148 N., R. 99 W.). P-P' indicates segment of channel that was measured. For resulting profile see figure 13.

Figure 9.--Map of drainage basin C (sec. 33, T. 148 N., R. 99 W.). P-P' indicates segment of channel that was measured. For resulting profile see figure 14.
alluvium. These sinkholes are the result of piping collapse and form a pseudokarst topography.

Squaw Creek, also located in the north unit of Roosevelt Park and designated as (D), was examined but not mapped. The channel (an arroyo) is situated on a broad alluvial plain bounded by pediment surfaces and colluvium. The inner part of the valley is approximately 1/4 to 3/8 of a mile wide. The lower reaches of the stream have dense stands of willows and roses. A portion of the channel and one of its tributaries was examined, but no channel scarps were found. The basin has a drainage area greater than 15 square miles.

Jones Creek, designated (E), is located in the south unit of Roosevelt Park (secs. 3, 4, 5 and 6, T. 140 N., R. 101 W. and sec 1, T. 140 N., R. 102 W.). The basin (fig. 10) has a drainage area of about 4.35 square miles and its stream has been trenched into the No. 2 terrace of the Little Missouri River. No channel scarps are present, but a number of side-channel scarps were observed. The channel width is approximately equal to depth in its lower 1 mile. Comparison of the present channel with 1939 photographs indicates little change in channel pattern except for entrenchment of the present channel. The lower segment of the valley was un-trenched in 1939 as determined from the presence of man-made ditches upstream from a main road which crosses the stream. The ditches were required to funnel runoff, then spread out in a distributary pattern on a fan, beneath the bridge. At this same location, the stream today has gullied its channel
Figure 10.--Map of drainage basin E (Jones Creek) (secs. 3, 4, 5, and 6, T. 140 N., R. 101 W., and sec. 1, T. 140 N., R. 102 W.). P-P' indicates segment of channel that was measured. For resulting profile see figure 16.
to a depth of 15 to 17 feet. The absence of channel scarps indicates that rapid retreat occurred following initiation of the erosion cycle.

Basin F (fig. 11), located in the south unit of Roosevelt Park (sec. 24, T. 140 N., R. 102 W.) has a drainage area of 1.12 square miles. The stream gullied into the thick valley fill and has active channel scarps and side-channel scarps. The channel, on the average, is twice as wide as it is deep.

Longitudinal Profiles

The longitudinal profiles of the present channel bottoms and the valley flats are shown in figures 12 through 17. The channels lacking active scarps have a generally constant slope. These channels (figs. 15 and 17) have a drainage area greater than 4 square miles (basins D and E). The profile from basin D (fig. 15) shows the relationship between a main channel and the steeper gradient of one of its tributaries.

The valley-flat surfaces, as plotted on the longitudinal profiles, show an irregular slope. These changes in slope appear as alternating steep and gentle reaches; the steepened reaches are more pronounced in the smaller drainage basins. Some of the steepened reaches have slopes of 3 degrees (fig. 12) or greater, compared with an average slope of about 1 degree for most of the valley flat. These points of increased slope are located downstream from a major tributary
Figure 11.--Map of drainage basin F (sec. 24, T. 140 N., R. 102 W.). P-P'-P'' indicates the segments of the channel that were measured. For resulting profiles see figure 17.
Figure 12.--A longitudinal profile of a segment of the channel in basin A (fig. 7). The approximate position of the water-table following a soaking rain is shown.

Figure 13.--A longitudinal profile of a segment of the channel in basin B (fig. 8).
Figure 14. -- A longitudinal profile of a segment of the channel in basin C (fig. 9).

Figure 15. -- A longitudinal profile of a segment of the channel in basin D and one of its tributaries.
Figure 16.--A longitudinal profile of a segment of the channel in basin E (fig. 10).

Figure 17.--A longitudinal profile of the confluence of two channels in basin F (fig. 11).
to the main channel.

Gaining and Losing Stream Segments

The gaining and losing segments of basin A were located as described on pages 19 and 20. The stream discharge at the time of the investigation was estimated to be approximately 0.50 cfs, having abated following the rain.

The position of the groundwater table following the rain in basin A is indicated on the longitudinal profile of that channel (fig. 12). Areas associated with slightly steepened reaches of the channel bottom were generally gaining water on the downstream side. Associated with these gaining reaches were steep banks in the inset fill of the channel and coarse sediment on the channel bottom. The channel bottom was very soft in these segments. Upstream from a channel scarp the channel bottom was found to be losing water to the ground. In these reaches there was an absence of banks in the inset fill, the channel bottom was firm and had a layer of fine sediment covering it which reached a few inches in thickness.

Upstream from the 2.3-foot inset scarp in basin A, three holes were augered and used as observation wells (fig. 19). The water-table became progressively lower in the direction of the scarp. Stream water and groundwater temperatures were taken at this point. The air temperature in the shade registered 82 degrees F., the surface water temperature was 86 degrees, and the temperature of the water
Figure 18.—A diagram showing the fluctuations in head level as the result of changes in the groundwater level between soaking rains in basin A.

Figure 19.—A plot showing the position of the water-table upstream from a channel scarp in basin A.
seeping at the base of the scarp was 68 degrees. To compare this with the temperature of groundwater from a more regional flow system, the temperature of water seeping from a lignite aquifer on the valley side was measured and registered 47 degrees F.

The piezometers installed at the base of the 13.5-foot scarp in basin A showed marked fluctuations between rainstorms. An attempt was made to measure the head levels prior to and following a soaking rain. The results (fig. 18) plus additional observations indicate that the water level rose following a rain and peaked within one to three days, depending on the duration and intensity of the storm, and then began to fall. The head readings obtained during July show the highest level in the 4-foot piezometer, indicating upward groundwater movement. Toward the middle of August, the reverse is apparently true because the head level in the 4-foot piezometer fell below the plane of reference (became negative). If no errors resulted from clogging, this area went through a transition from the upward movement of groundwater early in the summer to the downward movement of groundwater later in the summer.

Laboratory Models

The initial run was made to determine if a vertical face could be maintained in cohesive sediment in a flume. The sediment was nearly dry, hard, and contained many transverse and longitudinal shrinkage cracks; some extending to
the bottom of the flume. A vertical face was cut into the sediment prior to the beginning of the run. The run started with low discharge. The water soaked into the desiccation cracks and after one minute emerged at the face of the scarp. No water traversed the entire sediment surface of the flume. All erosion was by piping; cracks were enlarged until blocks of the sediment were separated.

Run number 2 was made for the same reason as stated above. But, the sediment was not allowed to dry completely to minimize shrinkage cracking. However, three transverse cracks about 0.02 foot wide and 0.25 foot deep, located about 3 feet above the scarp were present. The discharge was maintained at 0.65 gpm (about 0.05 foot deep) and there was flow over the scarp.

The sediment above the scarp was little eroded by channel flow, however, rapid erosion occurred at the face of the scarp; this appeared to be entirely by undercutting or removal of sediment from the scarp base and subsequent block collapse. A layer of water approximately 0.01-foot deep was noted adhering to and running down the face of the scarp even when the scarp was overhanging. Once a crack developed in the sediment just above the scarp it immediately filled with water and collapse of the downstream overhanging block followed. Soilfall ceased after the channel flow was terminated.

In run number 3 and in subsequent runs, an attempt was made to eliminate the water layer which was clinging to the
scarp face and appeared to be the major erosive process. The channel width was restricted and the flow depth increased to 0.25 foot (almost 1/2 the height of the scarp); the layer of water adhering to the scarp face could not be eliminated.

The effect of plunge pool activity was controlled by placing absorbent material at the base of the scarp to reduce the impact of the falling water and funnel it away from the base. Undercutting continued as previously observed. The reduction of the plunge pool activity caused the accumulation of sediment in blocks at the base of the scarp, thereby reducing the effective height of the scarp.

In subsequent runs of longer duration, the water table was observed through the glass-side of the flume. Although the velocity of the water in the channel and over the scarp face was relatively low, the groundwater intersecting the base of the scarp sufficiently reduced the cohesion of the sediment at the base of the scarp and made it more susceptible to erosion. The height of the intersection of the water-table with the scarp face appeared to control the height of effective undercutting.

Runs numbers 4 and 5 were made to determine if a channel scarp could be initiated on sloping sediment. In run number 4 the channel bottom had a slope of approximately 30 degrees. As soon as flow was established, the base of the slope was eroded away; the base was soft as a result of groundwater seepage. A vertical scarp was developed and began to retreat. In run number 5 the sediment had a primary slope of 6 1/2 degrees.
With a flow approximately 0.03-foot deep, a scarp developed 1.0 foot from the downstream end after 3 minutes.

The rate of retreat of channel scarps decreased with time, being relatively rapid at first and becoming progressively slower. After the initial rapid retreat of the scarp, the channel below the scarp began to widen. The height of the scarp increased with time, but appeared to reach a quasi-equilibrium height for a given discharge.

Sediment Sample Analysis

The grain-size analyses of the alluvium indicate that the predominate grain size is in the silt range. The size composition is variable but 20 percent sand, 60 percent silt, and 20 percent clay is typical of alluvium in western North Dakota.

The cumulative weight-percentage curves of 23 selected samples from the gully wall of Jones Creek (basin D) (sec. 6, T. 140 N., R. 101 W.) are shown in figures 20 and 21. The results show the clay content increasing at depths of about 5 feet and 16 feet. Much of the sediment, excluding these zones of high clay content, show good sorting with the mean grain size in the very fine sand (0.125 mm) to coarse silt (0.0625 mm) range. The analyzed samples from 7 to 16 feet below the surface show very consistent curves with little variation in the mean grain size. The clay content from this alluvial sequence is shown in figure 22.

The results of the carbonate analyses from Jones Creek
Figure 20.--Graphs showing the cumulative weight percentages of each sediment size class from the alluvium in Jones Creek. The numbers above each graph indicate the depth below the surface from which each sample was taken.
Figure 21. -- Graphs showing the cumulative weight percentages of each sediment size class from the alluvium in Jones Creek. The numbers above each graph indicate the depth below the surface from which each sample was taken.
Figure 22.—Changes in the chemical and physical properties of the sediment with depth. The letters (A through E) indicate recognizable units in Jones Creek.
(fig. 22) show marked fluctuations with depth. The upper 5 feet of sediment show minor fluctuations. At a depth of about 5 feet the percentage of carbonate in the sediment shows a marked decrease, with the lowest percentage (4.1 percent) occurring at depth of 6.0 feet. The carbonate content increases from this low value to a depth of 7.5 feet where the content becomes relatively constant to a depth of about 15.0 feet. Another marked decrease in the carbonate content can be noted with the low value of 5.0 percent occurring at a depth of 16.5 feet. The low carbonate content corresponds in depth to the zone of high clay.

The results of the analyses of oxidizable matter are also shown in figure 22. The upper 5 feet of sediment has intermediate but fairly constant values, excluding the high values of the very upper surface. Beginning at a depth of 5.0 feet, the oxidizable matter increases to 4.53 percent at a depth of 5.5 feet. Below this zone the content fluctuates, but another discernible peak is present at a depth of 16.5 feet. Below this zone the values of oxidizable matter are very low, in the range of 0.80 to 0.20 percent. The high peaks for the percentage oxidizable matter occur at the same depths as those previously mentioned for the low carbonate and high clay content.

**Stratigraphy**

Five stratigraphic units within the valley fill of Jones Creek were observed during the field investigations.
These alluvial units were differentiated on the basis of position, color, grain size, and the presence or absence of bone.

Unit A, the oldest unit exposed, is clayey sand underlain by gravel composed predominantly of clinkers. The unit is variable, but does not exceed 5 feet in thickness. The unit rests directly on Paleocene bedrock in all exposures examined. The physical and chemical properties of the upper part of the unit are shown in figure 22.

Unit B is a light gray to yellowish-brown silty clay and typically ranges in thickness from 1 to 2.5 feet but may be absent. Many trees buried by the overlying unit were observed to be rooted in this zone. Laboratory analyses indicate a greater clay content (up to 50 percent) and a lower carbonate content than adjacent older and younger units. As can be observed in figure 22, the percentage of oxidizable matter, although lower than the overlying unit, does have a peak in this zone.

Unit C is from 8 to 15 feet thick and is light gray to very pale brown clayey silt. Grain-size composition averages 20 percent sand, 50 percent silt, and 30 percent clay. The unit contains a greater amount of carbonate and oxidizable material than overlying and similar Unit E. An abundance of bone was found, the majority of which belongs to the modern species of bison. One big-horn sheep skull and some gastropods (unidentified) also have been found. This unit is present in most tributary valleys in western North Dakota.
Figure 23.--Composite sketch showing the occurrence and relative positions of the alluvial units.
and accounts for the bulk of the exposed valley fill.

Unit D is brown to dark brown silty clay and typically ranges in thickness from 0.5 to 2 feet but is absent in places. Laboratory analyses indicate a higher clay content (up to 75 percent) than adjacent older and younger units and an increase in oxidizable material. The carbonate content is 65 percent lower than in the underlying unit. The unit has a granular and crumbly structure. The upper 0.5 foot contains an abundance of plant remains; some living cottonwood trees rooted in this unit extend up through Unit E.

Unit E, a sandy silt and the upper deposit, generally ranges in thickness from 1 to 10 feet, but is absent in places. The sediment averages 25 percent sand, 60 percent silt, and 15 percent clay and is pale yellow in color. This unit is lighter in color and contains less clay than does the underlying similar Unit C. The unit is devoid of bison bones, but does contain small terrestrial and aquatic gastropods (unidentified).
INTERPRETATION OF RESULTS

Channel Scarp Initiation

The problems of an interpretation of the mechanisms involved in the formation and evolution of a channel scarp are those of reconstruction from existing data of the conditions of the alluvial flat directly preceding, during, and after the formation of the scarp. The conditions associated with the origin of channel scarps in western North Dakota will be presented in their probable order of importance in the following discussion.

Steepened reaches

The longitudinal profiles obtained during the field investigations show the presence of steepened reaches on the valley flat. These reaches occur downstream from large tributaries and are most pronounced in the smaller drainage basins. The position of this interruption in gradient can be most logically explained by a consideration of the hydrodynamics of the system during alluviation.

The streams of western North Dakota have undergone alluviation primarily during periods of sub-normal precipitation (see section on "Geomorphic and climatic implications", p. 82). As a result of the low rainfall, the slope vegetation is reduced and slope erosion occurs in greater proportions than under more normal conditions. This abundance of
available sediment causes alluviation in the main stream and lower reaches of the tributaries. Although there is much greater runoff from the exposed slopes, the total amount of sediment removed from the slopes is greater than the capability of the water available in the larger channels to remove it from the basin.

The tributaries enter the main stream carrying proportionally more bedload than the main stream as a result of their steeper gradient and consequently higher energy. In areas where the main streams flow east-west, the south-facing slopes and associated streams contribute more sediment than those facing north, due to microclimatic factors. The main stream cannot carry away all of the sediment supplied to it by the tributaries because of its lower gradient. An additional loss of some water to the ground further decreases the ability of the stream to move the sediment. During a period of sub-normal rainfall the groundwater level of the basin would be much lower than during a period of higher precipitation and would allow for greater percolation through the stream bed. This loss by percolation could amount to as much as 1 foot of water depth per day (Babcock and Cushing, 1942).

The result of these combined factors lead to adjustments in the transport and deposition by the main stream. A fan forms on the channel bottom adjacent to the tributary and the fan-front migrates downstream. Brice (1966) noted the presence of steepened reaches of the alluvial flat in
valleys in southwestern Nebraska, but stated that he believed that the majority of these reaches were formed upstream from major tributaries to the main stream because of damming of the stream by deposition at its confluence with the tributary. Most of Brice's information was obtained from topographic maps with a contour interval of 10 feet; these might not be entirely accurate for use in positioning the steepened reaches. In the area studied the steepened reaches of streams in western North Dakota occur downstream from the major tributaries.

Most fans are composed of coarser materials which probably formerly comprised the tributary-bedload; this coarseness results in increased permeability of the fan surface. This increased permeability leads to greater percolation of water through the fan surface and a decrease in the surface discharge, causing further deposition (negative feedback).

A climatic shift toward greater precipitation results in the re-establishment of the hill slope cover which retards slope erosion. The presence of vegetation on the slopes also retards the overland flow and allows greater infiltration of water on the slopes resulting in less runoff than for a storm of equal intensity and duration occurring during a drier period. As a result of this greater infiltration, the basin water-table rises.

Evidence of local recharge of the groundwater in basin A is its relatively high temperature. In this basin, a
groundwater temperature of 68 degrees F. was nearer to air temperature (82 degrees F.) on July 1, 1966 than to the temperature (47 degrees F.) of water from a lignite bed on the valley side. The temperature of water from this regional lignite aquifer more closely approximates the mean annual temperature of 42 degrees F. for western North Dakota, indicating that the water has been in the ground for some time. The basin groundwater temperature on the other hand seems to indicate local recharge with probably very little of the groundwater coming from regional flow systems.

Definitive data for this area is lacking but it is unlikely that the intensity or duration of individual rainstorms during drier periods was significantly different from the intensity and duration of rainstorms during wetter periods. The probable difference in precipitation occurring during these periods was in the frequency of occurrence. During the wetter periods the number of significant storms was probably greater than for a similar length of time during a drier period. This greater frequency of occurrence also contributed to the maintenance of a higher water-table.

In the following discussion, data from basin A (fig. 12) will be used to develop a mathematical model of conditions probably occurring in western North Dakota during the initiation of a channel scarp. The model will include calculated velocities above and on the steepened reach, probable changes in bedforms, and the changes in rates of transport on these reaches. The slope of the valley flat in basin A
is approximately 0.028 ft/ft downstream from the steepened reach, 0.056 ft/ft on the steepened reach, and 0.024 ft/ft above the reach. By the use of the Manning equation for channel flow (Chow, 1964), it is possible to calculate flow velocities which might be expected on the steepened reach. The equation is:

\[ V = \frac{1.5}{n} (R)^{2/3} (S)^{1/2} \]

where \( V \) is velocity, \( n \) is a channel roughness coefficient, \( R \) is the hydraulic radius (cross-sectional area of flow/wetted perimeter), and \( S \) is the slope, in ft/ft.

The slope was taken from the longitudinal profile. Assuming a width of 100 feet and various flow depths, the hydraulic radius can be calculated (the hydraulic radius nearly approximates the depth for flow in wide shallow channels). A value for bed roughness of 0.030 was selected (Olson, 1965). This is an empirical value derived for flow in natural, clean and straight channels.

Ten values of velocity are shown in Table 1. These values were calculated for the segment of the valley flat above and on the steepened reach. No attempt was made to compensate for the change in depth of flow as a result of the increased velocity on the steepened reach. The values obtained give orders of magnitude of flow in these two reaches of the stream.

By use of the calculated velocities and cross-sectional
Table 1.--Velocities of stream flow in a channel with varying slopes.

<table>
<thead>
<tr>
<th>depth of flow</th>
<th>width of flow</th>
<th>hydraulic radius</th>
<th>slope = 0.024 velocity</th>
<th>slope = 0.056 velocity</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1 foot</td>
<td>100 feet</td>
<td>0.1</td>
<td>0.3875 fps</td>
<td>0.600 fps</td>
</tr>
<tr>
<td>0.25 foot</td>
<td>100 feet</td>
<td>0.25</td>
<td>3.10 fps</td>
<td>4.80 fps</td>
</tr>
<tr>
<td>0.50 foot</td>
<td>100 feet</td>
<td>0.50</td>
<td>4.96 fps</td>
<td>7.68 fps</td>
</tr>
<tr>
<td>0.75 foot</td>
<td>100 feet</td>
<td>0.75</td>
<td>6.43 fps</td>
<td>9.96 fps</td>
</tr>
<tr>
<td>1.00 foot</td>
<td>100 feet</td>
<td>1.00</td>
<td>7.75 fps</td>
<td>12.0 fps</td>
</tr>
</tbody>
</table>

areas, the discharge can be determined for the selected stream segment in basin A. At a velocity of 3.0 fps (known in New Mexico to be a reasonable value), the stream has an instantaneous discharge of 75 cfs. The effective drainage area above this steepened segment of the stream is known (approximately 1.5 square miles) and therefore the amount of rainfall required to give this flow can be determined, assuming that there is little deficiency in the antecedent moisture of the basin. A flow at a rate of 3 fps would be 0.25 foot deep and would require a rainfall of approximately 2 inches over a 3-hour duration. A flow of 5 fps would be 0.50 foot deep and would require a rainfall of about 5 inches under the same conditions as above. No data are available
concerning the frequency of occurrence of storms of this magnitude in western North Dakota and therefore, no return intervals for storms of these magnitudes can be estimated. The values of 2.0 and 5.0 inches of precipitation over a three hour period are known to occur in some storms, however.

Leopold and Miller (1956) measured velocities exceeding 5 fps in arroyos 100 feet wide and with flows of 1 foot depth or less. The slope of the channel bottom was not given but from the large size of the basin it could be assumed not to exceed 1 degree. Therefore, the previously calculated velocities for basin A appear reasonable.

Schumm and Hadley (1957) noted that alluviation within valleys in New Mexico steepened the gradient and postulated that a critical slope might be reached which would initiate the cycle of erosion. They measured slopes of the alluvial surfaces as great as 3 degrees, which is approximately equivalent to the slope of 0.056 ft/ft in basin A.

The bed shapes that might be expected above and on the steepened segment of the stream in basin A are suggested by the Froude number associated with the flow. Using the values of velocity for a depth of flow of 0.25 foot, the Froude number can be calculated by the equation:

\[ F = \frac{V}{\sqrt{gR}} \]

where \( F \) is the Froude number, \( V \) is the velocity, \( g \) is the acceleration of gravity, and \( R \) is the hydraulic radius. The Froude number for a slope of 0.024 ft/ft (shallow reach) with
a depth of flow of 0.25 foot is approximately 1.0. The Froude number for the steepened reach (0.056 ft/ft) with the same depth of flow is approximately 1.7. A Froude number less than 1.0 indicates subcritical (tranquil) flow; the slope is considered mild. A Froude number greater than 1.0 indicates supercritical (rapid) flow; the slope is considered steep (Brush and Wolman, 1960). Bed forms associated with Froude numbers of 1.0 are transitional forms between ripples and a plane bed or antidunes. Antidunes occur at a Froude number of 1.7 (Allen, 1965). This indicates that during flow on the steepened reach in basin A the bed forms may have been those associated with higher energies of flow as opposed to more tranquil flow above and below this reach.

Assuming that the flow on the steepened reach of basin A was of higher velocity and had bed forms associated with higher energy environments than on the adjacent upstream and downstream sides, it logically follows that more sediment transport was occurring on this reach. A consideration of the rates of sediment transport can now be attempted.

Erosion along the channel bottom is a function of the transport rate at any point. Many contemporary workers feel that shear forces acting along the bed are of primary importance in the consideration of erosion (Brush and Wolman, 1960). The rate of transport per unit width is a function of the shear occurring at the bed and is defined by Brush and Wolman (p. 61) as:

\[ q_s = f(T_o) \]
where $q_s$ is the rate of transport per unit width of the channel and $T_0$ is the shear at the bed. The shear occurring at the bed can be expressed:

$$T_0 = \gamma ds$$

where $\gamma$ is the specific weight of water, $d$ is the depth of the water, and $S$ is the slope of the energy profile. Because the rate of transport ($q_s$) varies directly with the depth-slope product, any change in this product will cause $q_s$ to vary. Because the energy profile, from which the energy slope is obtained, is dependent upon the depth and velocity of the water (Olson, 1965), the highest energy slope occurs on the steepened reach. Brush and Wolman state that a maximum value of shear occurs on the upper point of steepening of the channel because the depth-slope product is maximum. They further state that downstream on a steepened reach the slope of the energy profile increases, but the depth decreases, resulting in a lower shear. Upstream from the steepened reach the depth of water increases, but the slope of the energy profile decreases, again resulting in lower shear.

The increasing velocity of the water on the entire length of the steepened reach results in a continued increase in the energy profile, thereby increasing the value of the slope of the energy profile (fig. 24), the entire steepened reach undergoes greater bed transport. The maximum rate of transport need not confined to the point of inflection of the channel bottom. At the toe of the steepened reach the
slope decreases and there is a transition back to a mild slope. At this point the depth of water increases but the energy slope greatly decreases, resulting in a lower shear and consequently a lower rate of transport. The result would be deposition of some of the sediment scoured from the steep slope.

Assuming that the amount of sediment transport is proportional to the shear, more material will be scoured from the steepened reach than from any other point of the channel bottom. This removal of sediment from the steepened reach acts to further steepen the slope resulting in positive feedback to the system; the steeper the slope, the greater the erosion on that slope. The limit of slope steepness is 90 degrees.

Another factor affecting erosion is the level of the groundwater table in the basin. When the water-table is lower than the channel bottom the stream loses water to the ground. The force or forces imparted by the water moving...
through the sediment act to compact the sediment, decreasing the erodibility. During a period of higher rainfall the basin water-table would maintain a higher position and if two or more rainstorms occurred within an interval of a few days, the water-table would rise following the first storm resulting in less insoak during the second. If the water-table underwent sufficient rise so as to intersect or nearly intersect the toe of the steepened reach, this portion of the slope would be more susceptible to erosion during the following storm or storms. Hence, the initiation of a channel scarp may not entirely depend upon the amount of water moving through the system during a given storm, but in part on the conditions existing in the basin prior to the period of runoff.

The laboratory study showed that channel scarps were initiated in cohesive sediment on slopes as low as 6 1/2 degrees. The scarps were initiated near the toe of the slope and progressed upstream.

A summary of the processes occurring on the steepened reach of a valley flat during scarp initiation is as follows: the increased slope results in increased water velocity causing a change from near tranquil flow above the reach to rapid flow on the reach; the increasing velocity results in a higher energy slope and greater erosion causing scouring of the steepened reach. The removal of sediment from the reach increases the angle of slope further increasing the erosive power of the water. However, the antecedent
conditions of the basin (groundwater level and soil moisture) may be of greater significance that the discharge of a single storm in the initiation of a scarp.

**Piping collapse**

Piping is the term applied to a process of subsurface erosion by water moving through sediment (Parker and Jenne, 1967). Piping occurs when surface runoff enters a drainable crack or other opening in the sediment where the hydraulic gradient is sufficient to cause erosion along the walls of the opening and carry away the fine-grained material. This process occurs most frequently in arid and semiarid regions.

The alluvium of western North Dakota contains abundant evidence of piping. Many of the pipes observed extend almost vertically downward for 10 feet or more before they become integrated with subterranean tunnels. After sufficient piping, the overlying sediment is susceptible to collapse and the overlying sediment may fall into the pipe leaving a channel to the surface. If this occurs in a drainage-way, the channel is broken by a scarp at the point where collapse ceased.

Channel scarps initiated by piping collapse have been observed in Mountrail County, North Dakota. Piping may have occurred in the steepened reach of other alluvial valleys, especially in smaller basins where the angle of steepening is greater, supplying the necessary hydraulic gradient to cause subsurface erosion.
Migration of larger channels

A situation contributary to scarp initiation might occur when the main river shifts away from the mouth of a tributary causing the tributary to form a fan on the floodplain of the main river. The loss of continual channel flow to carry away the sediment causes progressive steepening of the fan.

In the situation described above a scarp might form by one of two processes. Continued steepening of the fan accompanied by sapping at the toe by groundwater might bring about increased erosion during periods of runoff by the processes previously described. After initiation, the sediment is splayed out at the toe of the fan as the scarp migrates upstream. A second method of initiation might be due to the re-occupation of the old channel by the main stream. Should this occur, the toe of the fan would be cutoff and a scarp initiated.

Other methods of initiation

The initiation of a side-channel scarp in a tributary to a gullied reach can occur either by processes previously described or more simply due to the rapid change in base level as a gully in a main stream progresses upstream past the confluence of the tributary and the main stream.

The activities of man and animals as possible initiators of gullies has been discussed by many workers in the field. Animal trails on hillslopes may form channels and concentrate the runoff resulting in increased erosion. Animal burrows
may be a factor in the initiation of piping. Heavy overgrazing in a given basin may also cause accelerated hillslope erosion leading to alluviation in the valleys and the formation of oversteepened reaches.

Mechanics of Channel-Scarp Migration

The following discussion of channel-scrap migration is based on observations made in the field and in the laboratory.

Laboratory analysis

Brush and Wolman (1960) found that a vertical face could not be maintained in noncohesive sediment and from this it was inferred that the first requirement for channel scarp maintenance is that the sediment have a high resistance to shear (be cohesive). Therefore, the sediment used in the laboratory experiments conducted for this report was cohesive, averaging about 20 percent sand, 60 percent silt, and 20 percent clay. This is comparable to the alluvium studied in western North Dakota.

During initial investigations with the laboratory model it was believed that the majority of the erosion was controlled by the activity of the plunge pool. However, early experiments indicated that the base of the face was being eroded by a layer of water which was clinging to the vertical face. In an attempt to alleviate this condition the depth of flow was increased. It was noted that no matter what the depth of water, even when the flow was divided into a thin scarp flow and a much larger flow separated
from the scarp face, this layer adhered to the face and was very active in eroding the base of the scarp. Negligible erosion occurred upstream from the scarp on the upper surface of the sediment where the greater volume of water was flowing.

Observations of the water-table made through the glass side of the flume showed an apparent correlation between the height of the water-table and the location of active removal of sediment from the scarp base (undercutting). The face was actively undercut below the point at which the water-table intersected the face. This can be explained by saturation of the zone below the water-table, reducing the cohesion of the sediment and making it more susceptible to erosion.

Pooling of the water at the base of the scarp as a result of damming by debris caused the water-table to remain high. This pooling was the result of extended periods of low flow, which were sufficient to cause undercutting and collapse of the scarp face but were not sufficient to remove the debris.

To determine the effects of plunge-pool activity on the migration of the scarp, the plunging water was kept from impacting the sediment below the scarp. The layer of water adhering to the face continued to cause undercutting of the face, even in the absence of a plunge pool. The plunge-pool activity is therefore believed to be of major importance only in breaking up the blocks of sediment debris and making
it available for transport. In this respect, the plunge pool would be effective in controlling the height of the scarp face.

The hydraulic impact of the water separated from the scarp face forms a plunge pool a short distance downstream from the base of the scarp (fig. 25). The position of this plunge pool is a function of the amount of flow over the scarp. The impacting water in the plunge pool does not cause migration of the scarp face, however, as the scarp migrates upstream, the plunge pool also migrates upstream while maintaining its separation from the scarp.

![Figure 25.-A sketch showing the relative position of a plunge pool downstream from a channel scarp.](image)

Most of the sediment was removed from the scarp face in large blocks by soilfall and slumping, which was initiated by undercutting of the face, leaving a block of sediment unsupported from beneath. After sufficient undercutting of the face a crack appeared in the sediment at the upper
surface as a result of elastic failure of the overhanging sediment. The crack immediately filled with water. The hydrostatic pressure exerted by this column of water aided in overcoming the remaining shear strength of the sediment and thus hastened the process of soilfall. The pressure exerted on the sediment can be calculated (table 2). The presence of deep cracks in the sediment around a channel scarp was observed in the field.

Table 2.--The amount of hydrostatic pressure exerted by a column of water filling a crack in sediment.

<table>
<thead>
<tr>
<th>depth of crack (ft)</th>
<th>pressure exerted (lbs/ft²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5</td>
<td>32.25</td>
</tr>
<tr>
<td>5.0</td>
<td>322.50</td>
</tr>
<tr>
<td>10.0</td>
<td>640.50</td>
</tr>
<tr>
<td>20.0</td>
<td>1290.00</td>
</tr>
<tr>
<td>25.0</td>
<td>1612.50</td>
</tr>
</tbody>
</table>

If the block did not collapse within a short period of time after cracking, the large hydraulic gradient caused the initiation of piping at the base of the block. The piping released the water, removing the effects of hydrostatic pressure, but the erosive effects of the water moving along the vertical fracture was very great and quickly caused
block failure.

The frictional drag from surface flow continuing over the face of the scarp undoubtedly imparted some force which aided in overcoming the shear strength of the sediment. Upward groundwater seepage pressure was probably effective in the sloughing of some of the sediment at the scarp base, although no means could be found to measure this effect. The factors affecting undercutting and block failure are shown diagrammatically in figure 26.

- **A.** Factors acting on sediment during process of undercutting.

- **B.** Factors acting on sediment during process of soilfall.

Figure 26.--Cross-sectional diagrams showing the factors involved in the removal of sediment from the scarp base and soilfall.

Field observations

Difficulties arise in the direct observation of active channel scarps because of the infrequent summer rains in
western North Dakota. Arroyos were observed during very low flow, however, the rapid runoff from these basins normally results in a quick peak flow and rapid recession following a storm.

Observations of groundwater flow through a channel scarp indicate that the water-table does rise following a rain and intersects the base of the scarp. This water-table rise peaks one to three days after the rain. Therefore, the effects of groundwater seepage at the base of the scarp may not be effective during the surface flow peak from a given storm. As was noted in an earlier section, the greatest effects may be related to two or more closely spaced storms, the erosive effect of the second is enhanced by the high water table provided by the first.

Groundwater seepage greatly affects slumping and soil-fall at the scarp following a storm. This is shown by the abundant blocks of sediment which collect at the base of the scarp and which show no evidence of having been eroded after falling, indicating that there was little or no channel flow after collapse.

Summary of laboratory and field interpretations

To summarize the processes acting on a channel scarp during migration, a process-response model (Krumbein and Graybill, 1965) can be constructed (fig. 27). This model illustrates the main process elements on the left and the resulting responses to these elements on the right. The energy factors effective in removal of sediment from the scarp face include: frictional drag of the water moving
Sediment removal from scarp base (undercutting)

PROCESS ELEMENTS

ENERGY FACTORS
- frictional drag
- stream power
- seepage force
- gravity

MATERIAL FACTORS
- sediment grain size (cohesion)
- saturation

GEOMETRY
- vertical face

RESPONSE ELEMENTS

GEOMETRY
- overhanging face

MATERIAL
- sediment removed from face

Mass movement (soilfall)

PROCESS ELEMENTS

ENERGY FACTORS
- frictional drag
- hydrostatic pressure
- gravity

MATERIAL FACTORS
- sediment grain size (cohesion)

GEOMETRY
- overhanging face

RESPONSE ELEMENTS

GEOMETRY
- vertical face

MATERIAL
- sediment debris (blocks)

vertical face overhanging face

Figure 27.--Process-response models of undercutting and soilfall and the changes in geometry that occur in channel scarp retreat.
past the sediment, stream power (including plunge pool) acting primarily at the base of the scarp to remove sediment debris, groundwater seepage forces exerting forces acting to loosen the sediment, and the effects of gravity acting in a downward direction on the sediment. The material factors included in the process elements are: cohesion as a result of the fine grain size which maintains the scarp face, and the saturation of the sediment by groundwater and pooled water at the scarp base. The initial geometry involves a vertical or nearly vertical face. These factors can be considered to be acting on individual grains or collections of grains within the scarp face. The response to these processes is primarily in the geometry of the scarp which changes from vertical to overhanging as undercutting proceeds. Reinforcing positive feedback can occur by the collection of sediment slightly downstream from the scarp base which acts to pool the water and reduce sediment cohesion. Undercutting of the face also allows the force of gravity to be more effective.

This process can also be represented by a mathematical model showing the resulting undercutting as a function of the observed forces. The complex interrelations between the factors involved however does not allow them to be defined in the true mathematical sense of independent and dependent variables. As in the process-response model, the primary response will be selected as the dependent variable and the forces involved as the independent variables. The resulting expression is:
where \( Y \) represents undercutting and is a function of: 
\( S_a \) (saturation), \( F_s \) (seepage force), \( g \) (gravity), \( F_d \) (frictional drag), and \( C_0 \) (cohesion). As indicated by the expression, \( Y \) is inversely proportional to some function of \( C_0 \).

A process-response model of the soil fall occurring at the scarp face can be constructed in the same manner (fig. 27). In this case the factors can be considered to be acting on blocks of the sediment. The energy factors effective in producing elastic failure of the sediment include: frictional drag of the water moving past the sediment, hydrostatic pressure exerted by a column of water filling a crack in the sediment, and gravity acting downward on the overhanging block of sediment. The material factor is the sediment cohesion acting to maintain a vertical face and the initial geometry is represented by an overhanging face resulting from undercutting. The response to these factors is soilfall forming a vertical or nearly vertical face and the associated sediment blocks which accumulate at the scarp base.

The process of soilfall can be expressed by a mathematical model with the dependent variable being the removal of sediment blocks from the scarp face. The resulting expression is:

\[
Z = f\left(\frac{g, H_p, F_d}{C_0}\right)
\]
where $Z$ represents soilfall, $g$ is gravity, $H_p$ is hydrostatic pressure, $F_d$ is frictional drag, and $C_0$ is cohesion.

Because soilfall is dependent upon undercutting of the face, the combined processes of undercutting and soilfall can be combined into a two-phase process-response model (fig. 28) with the order of occurrence being from left to right.

The combined effects can be represented by a mathematical expression which is simply a combination of the two previous equations; $X$ represents scarp migration.

$$X = f \left[ \frac{(S_a, F_s, g, F_d)}{C_0} + (g, H_p, F_d) \right]$$

Gully Widening and Sediment Removal

The effects of groundwater seepage can be observed by block collapse from the gully wall after a rainstorm. This is probably the result of bank storage which accumulates during periods of stream flow. During the recession of stream flow water in the bank sediment undergoes rapid drawdown. The forces exerted may be sufficient to cause the sloughing of sediment from the walls because saturation decreases the cohesion of the sediment and increases the total sediment weight by the addition of pore water.

The increased weight of the water-sediment combination, the decrease in sediment cohesion, and the increased seepage pressure result in slumping of the gully wall. Large masses of collapsed sediment can be observed lying across the gully.
Figure 28.--A 2-phase process-response model showing the combined effects of undercutting and soilfall on channel scarp migration.
floor following a rain. This movement apparently occurs after the runoff because the blocks lack noticeable subsequent erosion.

The sediment which falls and slumps following a rain constitutes the material available for transport during the next period of runoff. This process appears to be one of the most important, if not the most important process of channel widening.

The sporadic migration of channel scarps as reported by many investigators may be related to the frequency of occurrence of rainstorms during a given year. If a succession of rainstorms occurs the water-table in the basin rises following the first storm and intersects the base of the channel scarp, making the sediment much more susceptible to erosion during following storms. Therefore, the years of very rapid migration of a channel scarp may be related to a greater number of more closely spaced rainstorms.

An estimate of the amount of alluvium removed from basin A was made by a comparison of photographs taken in 1938 with photographs taken in 1958 and with data collected in the field. Determination of the amount of alluvium removed was made using values of the amount of scarp retreat and channel widening measured from air photographs and channel depths measured in the field. The estimates do not include channel deepening because no methods were available to determine the change in depth. The amount of alluvium removed by the retreat of the highest downvalley scarp in
basin A between 1938 and 1966 was calculated to be approximately 286,000 cubic feet. Gross calculations of the entire amount removed from the basin since 1938 was 9,450,000 cubic feet or approximately 567,000 tons. Averaging this figure for the interval 1938 to 1966 yields 20,250 tons of alluvium removed from the basin per year. These figures do not include material derived from the slopes and therefore cannot be implied to be the total sediment yield from the basin.

A comparison of 1939 photographs of Jones Creek (E) (fig. 10) indicates that a great amount of sediment had been removed, but no estimations of the total amount were made.

Stratigraphy

The alluvium of western North Dakota contains evidence of distinct periods of erosion and deposition. Three alluvial units (designated A, C, E, bottom to top) were distinguished in the gully of Jones Creek. Zones of secondary alteration are apparently present on the upper surface of Units A and C (see fig. 23, p. 51). These have been identified as paleosols on the basis of physical and chemical analyses (fig. 22, p. 48) and the lateral persistence of the zones in the field. The units (described on pages 49 and 50) were observed in several gullies over a distance of approximately 100 miles along the Little Missouri River and therefore probably occur throughout most of western North Dakota.
Age of units

A buried tree rooted in Unit B (lower paleosol) was radiocarbon dated at less than 185 years B. P. (I-2325), indicating that overlying units C through E, 15 to 20 feet thick, have been deposited since about A. D. 1765.

The age of Unit A is unknown because no evidence has been found to indicate the time of deposition. The soil forming processes acting during the formation of Unit B ceased prior to about 1765. The eroded surface of Unit B indicates that following its formation the area underwent a period of degradation. Several trees rooted in Unit B and buried by Unit C were observed. Unit B is preserved best where these trees had established a root system capable of holding the soil together and thereby retarding erosion. In other exposures of Unit B, the local relief was probably not great enough to have promoted erosion. The largest trees were about 2 feet in diameter and were probably well established before erosion began. The smallest trees observed were less than a foot in diameter and were estimated to be about 20 years old before burial. If the youngest trees present began to grow just preceding degradation they might be used to approximate the length of time associated with the period of degradation. Trees not established before degradation began would not have been seeded or if seeded would not have survived through the period of erosion. Because the youngest buried trees observed were estimated to be 20 years old, the period of degradation preceding the
deposition of C was probably no longer than 20 years.

Following the erosion of Unit B, alluviation began, which resulted in the formation of Unit C. The beginning of this period is after 1775. Unit E, which directly overlies C in places, was deposited sometime after 1880. This date can be established because Unit E lacks bison bones, indicating that it formed after the 1880's when bison became extinct in this area (Robinson, 1966). The total amount of time involved in the deposition of Unit C can be further shortened by considering the time required for the development of the soil on its upper surface. By visible and chemical comparison of this soil with the faintly developed soil on the present surface, it is felt that assuming the same climate existed for both at least three times the amount of time required for the present soil was required for the development of Unit D. Because the present surface has been exposed for approximately 30 years (see following paragraph), the period of time required for the formation of Unit D would be on the order of 100 years. When this figure is subtracted from the beginning of deposition of Unit E, which is about 1920, a date of about 1820 is obtained for the end of deposition of Unit C.

As was mentioned previously, the absence of bison bones in Unit E indicates that the earliest date for its deposition is sometime after the 1880's. However, the presence of living cottonwood trees rooted in Unit D, estimated to be approximately 40 to 50 years old, indicate that the beginning
of deposition of that unit was about 1920. The presence of 20 to 30 year-old trees on the present surface indicates that deposition ended about 1935.

**Depositional rates**

Units C through E, which are 15 to 20 feet thick, probably formed between A.D. 1775 and 1935, a total of 160 years. However, the deposition of this sediment was interrupted for a period of approximately 100 years, during which time a soil formed and underwent a slight period of erosion. This leaves a period of approximately 60 years for the actual deposition of these sediments. The depositional rates indicated are on the order of 0.3 to 0.4 feet per year. These figures are of the same order of magnitude as the 0.34 feet per year calculated for the deposition of recent alluvium in a somewhat similar environment in southwestern Nebraska (Brice, 1966).

It is not likely that this deposition was constant for each year. Discolorations and fine banding in Unit C probably indicate times during which the deposition was much less than average. Conversely, at times deposition probably exceeded 0.4 feet per year. It is felt however, that the range from 0.3 to 0.4 feet per year is representative of the average sedimentation rate occurring during this period.

**Correlation**

In studies of the postglacial chronology of alluvial valleys in Wyoming, Leopold and Miller (1954) identified
several formations ranging from Mankato to recent in age. These formations were correlated with similar units throughout the western United States. The most recent period, which began about A. D. 1200 and ended approximately in 1880, resulted in the deposition of the Lightning Formation. The recent valley fill of western North Dakota, possibly excluding Unit E, can be tentatively correlated with the upper part of the Lightning Formation.

In a study of erosion and deposition in Nebraska, Brice (1966) describes a sequence of recent alluvium which is divided into a banded upper part and a less banded lower part. The banding results from concentrations of clay and organic matter. A carbon-14 date from charcoal in the upper alluvium yielded a date of $420 \pm 160$ years B.P., suggesting a correlation with the Lightning Formation and probably also with the recent alluvium of western North Dakota.

Any attempt at exact correlation between recent alluvial deposits should be done carefully, because the factors which cause aggradation in the individual valleys may have been important at different times. In a semiarid environment, only a slight change in the factors controlling either erosion or deposition may result in a local change in the rate of that process. A correlation of this type can at best group only similar occurrences which may have been independent of one another in time.

**Geomorphic and climatic implications**

The sequence of alluvial fills is related to either
climatic fluctuation, animal activity, or a combination of both. The deposition of sediment prior to 1880 was most likely the result of a climatic shift because the influx of settlers and large scale grazing in this area did not occur until after 1880.

Prior to the deposition of Units A through E in western North Dakota, the area underwent a cycle of stream incision, during which time most pre-existing valley fill was stripped from the valleys. A remnant of this incision is in the form of a basal lag gravel (lower Unit A) which overlies the Fort Union Group in all areas where exposed by present erosion. This incision was probably caused by a greater than average amount of precipitation which may have caused the streams of the area to become unstable and to begin downcutting. This entrenchment may in part be associated with 39 years (1663 to 1702) of excess precipitation in North Dakota (Will, 1946). This correlation is strictly speculative as no age has yet been determined for this unit.

Following this incision, a period of quiescence was reached, which was probably associated with a return to more normal precipitation. During this time the streams resumed a state of equilibrium with the environment. The quiescence resulted in the development of a soil (Unit B).

A climatic shift toward lesser precipitation, probably about 1775, caused a reduction in the slope cover with an increase in the sediment yield to the basins and a decrease in the stream flow to carry it away. The result was the
deposition of 10 to 12 feet of sediment, here designated as Unit C.

A return to normal precipitation about 1820 caused an increase in the slope cover thus decreasing the sediment yield to the basins. This decrease in slope erosion brought an end to the period of alluviation and was followed by the formation of a soil (Unit D).

Gullying probably began in the lower reaches of some of the larger tributaries as early as the late 1880's. Unpublished field notes made in 1905 by A.G. Leonard, State Geologist, indicate that some gullying had begun at that time. However, the number of gullies described does not seem as great as is evident today.

Renewed valley filling in this century was again associated with a decrease in precipitation. Precipitation records for North Dakota have been evaluated by Bavendick (1952), showing that beginning with the year 1917 and culminating in 1936, when Unit E was deposited, western North Dakota underwent an almost continuous series of droughts. Records show that the yearly precipitation fell below the average of slightly more than 15 inches to as low as 5 or 6 inches in 1934. This period was accompanied by strong winds which caused much blowing and drifting of sediment. Bavendick (p. 17) cites measurements indicating that from 2 to 7 inches of soil was removed from many fields in 1934.

Size analyses of the sediment from Unit E indicate good
sorting with the mean grain size in the 0.0625 to 0.031 mm range, which is characteristic of loess. In the exposures examined, any wind blown sediment present had probably been reworked by water before final deposition.

The rapid sedimentation which occurred in western North Dakota is thought to be almost entirely controlled by climate. Because modern records show precipitation deficiencies of enough magnitude (figures 2 and 3) to eliminate most of the established vegetation, it is not inconceivable that similar fluctuations have occurred in the near past.

Erosional and depositional history

The tributary valleys in western North Dakota have probably undergone many periods of erosion and deposition during and since the Pleistocene. The exposed alluvial sequence indicates that within the past few hundred years the valleys have undergone four periods of erosion and three periods of deposition (fig. 29, p. 87 and 88).

The valleys were probably cut to near their present depth during the Pleistocene. Periods of valley filling have also been associated with the Pleistocene. Howard (1960) concluded that the upper terrace silts along the Missouri River are Mankato in age and stated that deposition probably continued beyond the Mankato climax. The tributary valleys were most probably filled during this time also.

Another period of valley filling was probably asso- ciated with the post-glacial Hypsithermal interval. The No. 2 terrace along the Little Missouri River, as described
by Laird (1950), may possibly correlate with this interval. If this assumption is true, the tributary valleys were filled again.

During the period prior to A.D. 1700, the tributaries again underwent erosion which stripped most of the pre-existing valley fill from the valleys. The remnant of this incision is represented by Unit A of the recent valley fill in western North Dakota. This unit is a sand and gravel lag unconformably overlying the Paleocene sediments. LaMarche (1966) cites evidence indicating that rapid downcutting was occurring in southwestern Utah around 1700. He attributes this erosion to local rather than widespread factors, stating that in no other part of the United States has a similar cycle of erosion been noted for this recent period. The erosion in western North Dakota is likewise attributed by the writer to local climatic factors that were apparently occurring coincidently with these in southwestern Utah.

Following this period of entrenchment was a time of relative quiescence and the formation of a soil on the upper surface of the lag sediments. The eroded nature of this paleosol suggests that a period of at least slight entrenchment preceded the deposition of the next alluvial unit. The alluviation resulting in Unit C then filled the channels and raised the surface of the valleys some 10 to 15 feet. Following this alluviation was another period of soil formation.
Figure 29.—Sketches depicting recent erosion and deposition in western North Dakota (see description on following page).
Figure 29.--Composite sketches depicting the recent erosion and deposition in western North Dakota. A, Tributary valleys were trenched prior to the 1700's. B, Deposition of Unit A and formation of a soil (Unit B). C, A period of erosion about the middle 1700's. D, Deposition of 10 to 15 feet of sediment (Unit C) and the formation of a soil (Unit D) during the late 1700's and early 1800's. E, Another period of erosion beginning about the late 1800's. F, Deposition of the upper unit (Unit E) between about 1920 and 1936. G, Present erosion exposing the alluvial sequence.
Entrenchment of the tributary streams began again in the middle or late 1800's and is indicated by the eroded nature of the upper surface of Unit C. This entrenchment was interrupted by the deposition of another alluvial unit during the 1900's which either partially or entirely filled the gullied reaches. Present erosion has again cut the channels to bedrock and exposed the sequence of alluvium.

Comparison of channel patterns from photographs taken in 1938 and 1939 with those taken in 1958 show that few changes have occurred during this period. This lack of change was also observed in the field in the presence of lag channel gravels and cross beds now exposed in the gully walls which generally correspond to present channel positions. The pebbles were almost entirely clinkers which had been eroded from burned-beds in the valley walls. The pebbles were predominately disk-shaped and were imbricated with upstream dip. The positions of these gravels exposed on the gully walls indicated the approximate limits of the previous channel walls. A portion of Jones Creek and the locations of these gravels are shown in figure 30. The coincidence of the path of the present stream with that of buried stream channels has been noted in various other basins. The explanation of this may simply be that the buried channel had a surface expression which concentrated runoff from the valley flat, thereby initiating the new gully system in the same position as the old. The cause may be more complex than this simple relationship.
Figure 30.—A map of a segment of Jones Creek (basin E) approximately 1/2 of a mile upstream from the mouth, showing the relationship between the past and present channel.
Because the channel gravels had a higher permeability than the surrounding silts they would form a conduit for the groundwater in the basin. If this were true, the areas overlying and bounding the buried gravels might be more susceptible to erosion during periods when the basin groundwater level was high. This is only conjecture because the relationships between groundwater and surface water which cause erosion are poorly understood.
SUMMARY

The vertical-walled gullies in the valley fill in western North Dakota are the result of rapid headward erosion by a series of channel scarps. The profiles of the valley flats and gully floors suggest that the primary origin of these scarps is related to increased erosion on a locally steepened reach of the valley flat that is formed downstream from a major tributary. In a given basin, the origin of a channel scarp is related to the angle of steepening of the slope, the amount of discharge, and also probably related to the moisture conditions of the basin preceding each runoff.

Other possible modes of origin of channel scarps include: collapse by piping, steepening due to migration of a larger channel, rapid change in base level, and possibly by the activities of animals.

The vertical or nearly vertical scarp face is maintained because of high cohesion of the sediment. Rapid erosion at the base of a scarp is related to groundwater seepage which reduces the cohesion and makes the sediment more susceptible to erosion by water flowing past it on the face. The height of the scarp, if not bottoming in bedrock, is related to the amount of water available to remove the sediment from the scarp base which is in turn related to the position of the
scarp in the drainage basin.

Channel scarp retreat is a cyclic process involving removal of sediment from the face of the scarp resulting in soilfall or slumping of the overhanging sediment which again renews the vertical face. The sporadic migration of a channel scarp may be the result of the spacing of rainstorms. If two rains occur within a few days of one another, the basin water-table rises following the first rain and intersects the scarp face making it more susceptible to erosion during the following storm.

The majority of the alluvium removed from the gully is derived from the walls of the gully by slumping and results in gully widening. This sediment becomes available for transport by a later period of runoff. Gully wall slumping is a response to saturation of the banks by groundwater in bank storage during runoff and is aided by positive seepage pressure associated with rapid drawdown of the bank storage during recession of the flow.

The late-recent alluvium of western North Dakota can be divided into five individual units, two of which are paleosols. By carbon-14 dating and the estimation of tree age it was established that the upper 15 to 20 feet of this sediment was probably deposited between 1775 and 1936. This 160-year period can be further shortened by approximately 100 years which presumably were required for the formation and subsequent partial erosion of a soil developed between these units. An average rate of sedimentation on the order
of 0.3 to 0.4 feet per year is believed to be characteristic of these units.

The established age of this alluvium makes possible a tentative correlation with similar alluvial units in Wyoming and Nebraska, which have been studied and dated by investigators in those areas.

The periods of valley filling in western North Dakota are associated with periods of sub-normal precipitation which decreased the slope cover, thereby increasing the sediment yield to the valley floor. Modern precipitation records indicate that such a period occurred from about 1920 to 1936 during which time the upper unit was deposited.

The alluvial units and nature of the intervening contacts indicate that the tributary valleys in western North Dakota have undergone four periods of erosion and three periods of deposition within the past few hundred years.
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