Erosional and Depositional Aspects of Hurricane Camille in Virginia, 1969

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CONTENTS

ILLUSTRATIONS

TABLES

SYMBOLS

		Dimensions
a	coefficient in the equation of longitudinal stream-profile	
A	cross-sectional flow area	L^2
A_d	drainage area	L^2
b	coefficient in various equations, representing slope of best-fit line	
c	coefficient representing portion of total rainfall that runs off the ground surface	
C	constant of integration in Hack (1957, p. 70) equation for stream profile	
d_{av}	average intermediate diameter of sediment particles	L
d_{10}	grain size for which 10 percent of the distribution is finer	L
d_{l5}	average intermediate diameter of the five largest rocks at a site	L
e	base of natural logarithms	
F	vertical fall from drainage divide	L
g	acceleration due to gravity	LT^{-2}
I	rainfall intensity	LT^{-1}
k	constant in Hack equation for stream profile	
L	horizontal distance from drainage divide	L
n	constant in Hack equation for stream profile	
Q	discharge of water	L^3T^{-1}
r_c	average radius of curvature of a channel bend	L
r_i	inside radius of curvature of a channel bend	L
r_o	outside radius of curvature of a channel bend	L
S	local slope (gradient)	
sk	skewness of a sediment size-frequency distribution	
S_o	sorting of a sediment size-frequency distribution	
V	mean velocity of flow	LT^{-1}
W	water-surface width of a stream	L
x	distance downstream from start of deposition	L
α	constant in equation of stream profile	
β	constant in equation of stream profile	
γ	constant in equation of stream profile	
Δh	elevation difference between high-water and low-water marks around a channel bend	L

ENGLISH-METRIC CONVERSION TABLE

Length:

Inches × 0.0254 = meters
Feet × 0.3048 = meters
Yards × 0.9144 = meters
Miles × 1.609 = kilometers

Area:

Square inches × 0.0006452 = square meters
Square feet × 0.09290 = square meters
Square yards × 0.8361 = square meters
Acres × 4047 = square meters
Square miles × 2,590,000 = square meters

Volume:

Cubic inches × 0.01639 = liters
Cubic feet × 28.32 = liters
Cubic yards × 764.6 = liters
Pints × 0.4732 = liters
Quarts × 0.9463 = liters
Gallon × 3.785 = liters
Acre-feet × 1233 = cubic meters

Weight:

Grains × 0.06480 = grams
Ounces (avoirdupois) × 28.35 = grams
Pounds (avoirdupois) × 453.6 = grams
Tons (short) × 907.2 = kilograms
Tons (long) × 1016 = kilograms

Specific combinations:

Feet per second × 1.097 = kilometer per hour
× 0.3048 = meters per second

Miles per hour × 1.609 = kilometers per hour
× 0.4470 = meters per second

Pounds per square inch × 70.3 = grams per square centimeter

Pounds per square foot × 0.4885 = grams per square centimeter

Tons (short) per square foot × 0.9765 = kilograms per square centimeter

Tons (short) per acre × 0.2241 = kilograms per square meter

Tons (short) per square mile × 0.0003502 = kilograms per square meter

Pounds per cubic foot × 0.01602 = grams per cubic centimeter

Cubic feet per second × 1.699 = cubic meters per minute
× 0.02832 = cubic meters per second

Cubic feet per second for 1 day × 1.983 = acre feet
× 2446 = cubic meters

Degrees Fahrenheit − 32 × 0.556 = degrees Celsius.

EROSIONAL AND DEPOSITIONAL ASPECTS OF HURRICANE CAMILLE IN VIRGINIA, 1969

By Garnett P. Williams and Harold P. Guy

ABSTRACT

Probably the worst natural disaster in central Virginia's recorded history was the flood resulting from an 8-hour deluge of about 28 inches (710 mm) of rain on the night of August 19–20, 1969. This study examines some of the intensive sediment erosion and deposition that resulted from the storm and flood.

Most of the 150 people whom the flood killed in this mountainous area died from broken bones and other blunt-force injuries, rather than by drowning. The transport of sediment and other debris by the water therefore was very significant in loss of life and in property damage.

Erosion resulted mainly from debris avalanches down the mountain-sides and channel scour along streams and head-water tributaries. Total amounts of sediment yield from certain mountainous areas in Nelson County were about 3.2–4.6 million cubic feet per square mile, probably the equivalent of several thousand years of normal denudation.

Characteristics of the debris avalanches were that (1) they usually followed pre-existing depressions on hillsides and occurred on slopes greater than 35 percent, (2) the upslope tip of the avalanche scar tended to be located at the steepest part of the hillside, where the convex slope merged with the concave or planar zone immediately below, (3) hillsides facing north, northeast and east were more susceptible to avalanching than slopes facing other directions, and (4) debris-avalanches caused rapid and devastating surges of water and sediment in the mountain-stream channels. Such surges in some instances temporarily blocked the channel flow upstream.

Slightly more than half of the total sediment contributed to the stream system was from erosion of stream channels. Channel erosion was very irregularly distributed; some ravines 10–20 feet wide and 5–10 feet deep were scoured in places which formerly had only a very small channel, whereas other channels only a few hundred yards away experienced little or no channel erosion.

By the use of figures for the total amount of sediment removed from a drainage basin and the duration of the storm, estimates were made of the storm-average sediment-transport rate at the mouth of various basins. For drainage basins ranging up to about 1.5 square miles, the estimated storm-average sediment-transport rates varied from practically nothing to as much as 172,000 pounds per second (7.4 million tons per day).

The types of sediment deposits were (1) debris-avalanche deposits, rather rare, at the base of hillslopes, (2) mountain-stream channel deposits, usually in scattered sediment patches but locally occurring as large wedge-shaped deposits behind debris dams, (3) alluvial fans, (4) delta-like deposits at the junction of a stream and major highway, where water backed up during the flood due to plugging of a culvert, and (5) accretion deposits on flood plains. The highway deltas and some downstream flood-plain sediments consisted mostly of sand-sized grains, but the other types of deposits usually contained particles ranging from silt or clay to boulders 5–10 feet in diameter. Changes in grain size and in volume of deposition with distance downstream were measured, and sedimentary features of the various types of deposits are described.

INTRODUCTION

On the night of August 19–20, 1969, the central part of Virginia was deluged with some 28 inches of rain from the remnants of Hurricane Camille. The loss of life and property due to the flash floods, landslides and other sediment damage accompanying the storm has been called the worst natural disaster ever to strike Virginia. The storm was one of the most severe ever recorded in the United States.

The region most affected was Nelson County, a rural area midway between Charlottesville and Lynchburg (fig. 1). About 150 people, 125 of them in Nelson County, died in the storm and flood. Damage in Nelson County alone amounted to $116 million worth of public and private property, including about 150 homes and other buildings, 120 miles of roads, 150 bridges and culverts, hundreds of cars and trucks and 25,000 acres of cropland, including orchards. About $150 million in public funds, mostly federal, has been spent on recovery, including $8 million in debris clearance. Particularly heavy damage occurred in Massies Mill, Woods Mill, Roseland, Tyro, Lovingston, Norwood, Rockfish, and along Davis and Muddy Creeks (fig 1).

On the basis of the extensive damage to life and property, the Commonwealth of Virginia has made recommendations for future land use in Nelson County (Commonwealth of Virginia, 1970). Kuhaida (1971) also offers some pertinent observations.

This report is a study of the extraordinary erosion

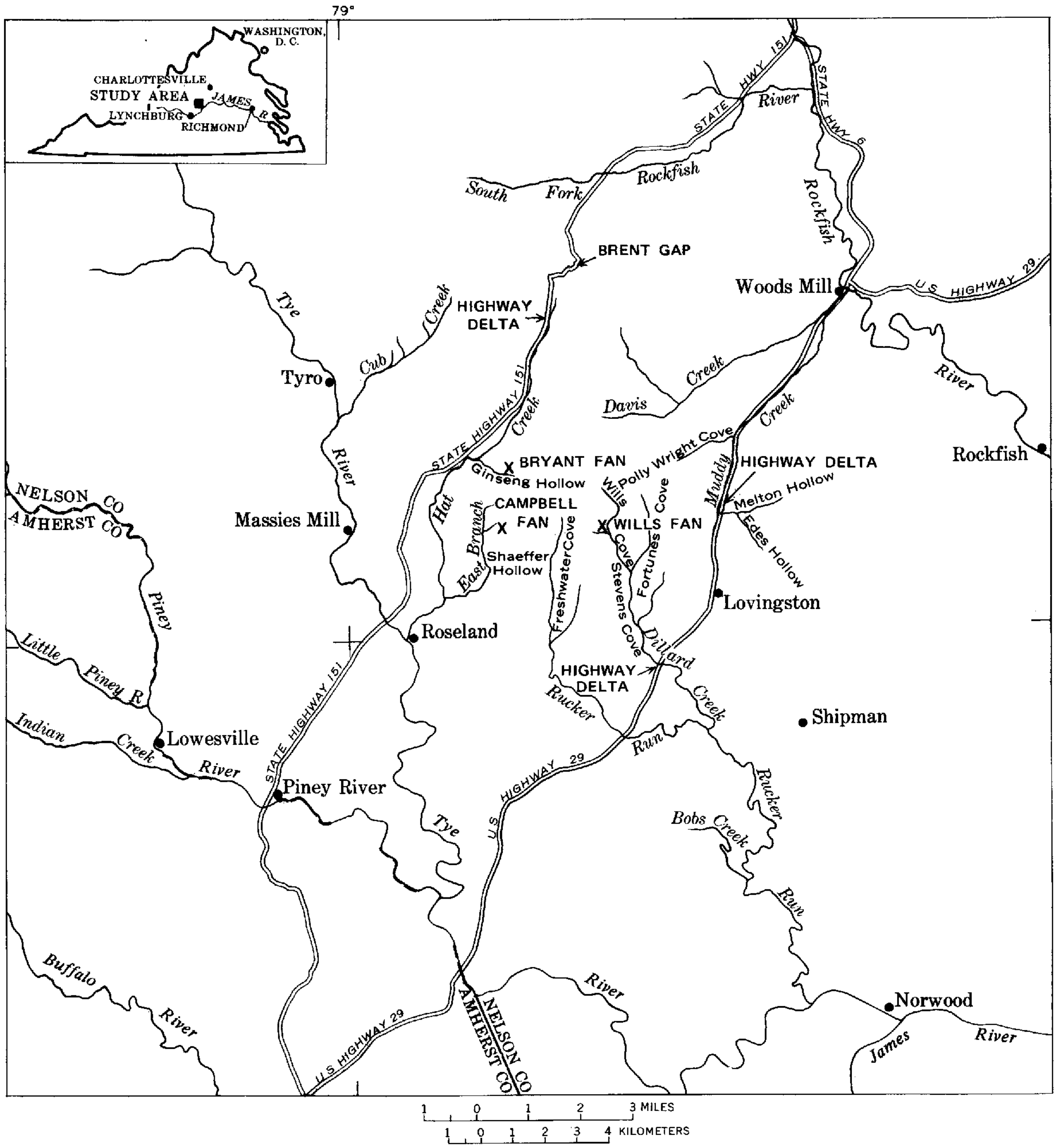


FIGURE 1.—Study area.

and deposition which the storm inflicted on parts of Nelson County. Most of our field observations and measurements were made within 3 months after the flood. Erosion and deposition, somewhat less severe, also occurred in regions bordering Nelson County, but time limitations permitted only a brief reconnaissance of these nearby areas.

Because much of the sediment deposition from the flood interfered with the daily operations of man, a massive sediment-cleanup campaign began soon after the flood. Bulldozers effectively destroyed most of the alluvial fans and many flood-plain deposits in the weeks and months following the flood. Necessary though this work was, it did limit the scope and thoroughness of a study of the erosion and deposition. All of the data presented in this report were obtained before man's interference with the deposits. However, the writers would like to emphasize that anyone wishing to study flood deposits should act as soon as possible after the flood.

We acknowledge with thanks the many helpful manuscript suggestions of Dwight R. Crandell, Luna

B. Leopold, and John T. Hack. Jon W. Nauman provided valuable assistance with some of the field work. Mr. and Mrs. S. K. Wills, Robert Bryant and Guy Spencer, residents of Nelson County, allowed us to roam at will over their land. Many excellent photographs were provided by the Virginia Division of Mineral Resources, the Virginia Department of Highways, Ed Roseberry and others.

GENERAL DESCRIPTION OF STUDY AREA

The devastated area lies within the James River drainage basin in the Blue Ridge geomorphic province. Most of the region is mountainous (fig. 2), with peaks towering as much as 1,200 feet above the valleys. The climate is temperate-humid. Trees, bushes, small plants and dead leaves cover all of the hillslopes and ridges. Tree varieties include walnut, poplar, oak, ash, locust, hickory, and sycamore.

Bedrock is mainly Precambrian gneisses, granites and schists of the Lovingston and Marshall Formations (Bloomer and Werner, 1955). The geology is very complex, with a variety of rock types. Jointing is present in some places and apparently absent in others. On hillslopes the bedrock generally lies from a few inches to about 20 feet below the surface. The soil, formed in place from the bedrock, is typically a reddish or brown loam and generally contains some rock fragments.

THE STORM

Hurricane Camille struck the State of Mississippi on the Gulf of Mexico on August 17, 1969. As it moved inland to the north and then curved eastward, the hurricane weakened considerably. Eastern Kentucky received only 1–2 inches of rainfall.

On the night of August 19 a rare combination of circumstances brought about a rapid intensification

FIGURE 2.—Upstream view of Wills Cove and Dillard Creek, a typical drainage basin studied in this report. The photograph, taken a few days after the flood, shows valley flood-plain deposits and hillside scars due to debris avalanches. Main stream from Fortunes Cove enters from right foreground. (Photograph courtesy of Virginia Division of Mineral Resources.)

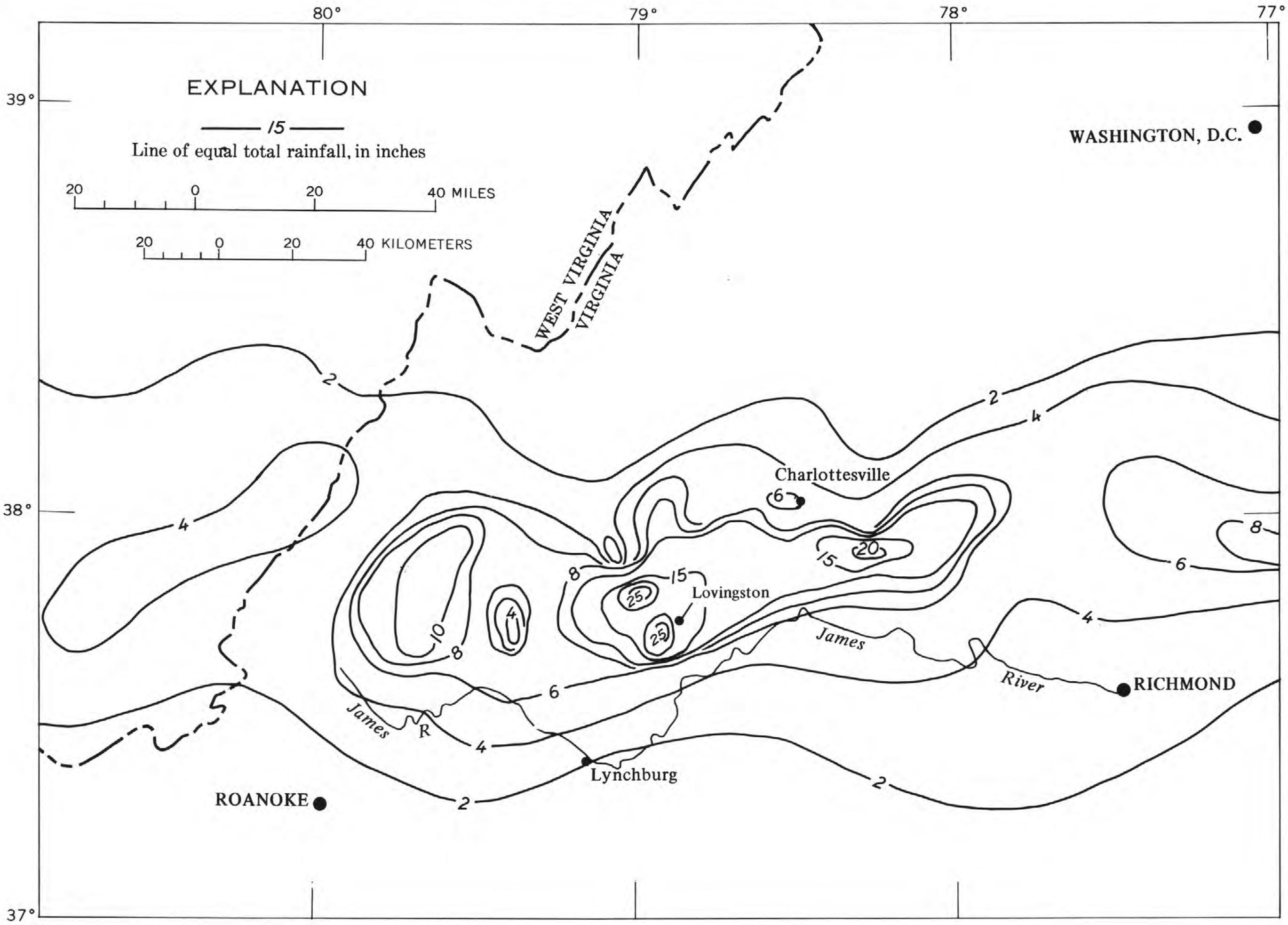


FIGURE 3.—Rainfall from noon August 19 to midnight August 20, 1969. Adapted from DeAngelis and Nelson (1969).

of the rainfall in the region just east of the Blue Ridge Mountains, especially around Nelson County, Virginia (Schwarz, 1970). The air in the region southeast of Nelson County was nearly saturated with moisture that had been accumulating for several days before the storm. This moist air extended up to an altitude of about 3,000 feet. The hurricane itself contained a large amount of moisture at higher elevations. As Camille arrived, the remnant circulation of the hurricane created a flow of wind in a northwesterly direction at the lower levels. These winds drew moist air from the southeast, and this moisture moved under and joined that of the hurricane to form a very deep layer of moist air. Thunderstorms developed which, because of the unusual thickness of the layer of moist air, were particularly intense and persistent. Figure 3 shows the rainfall distribution in central Virginia for the period covering the storm.

In some places the orography may have contributed to the intense rainfall by funneling the northwest-flowing air into the mountain ridges. However, the effect of orography on the rainfall is difficult to determine. Over 21 inches of rain fell in a region about 15 miles southeast of Charlottesville where the terrain, while not flat, is devoid of any steep or high mountains. Thus some areas received torrential rain with little or no influence from nearby mountains.

The deluge was of catastrophic cloudburst proportions. Rainfall of 12–14 inches was common in Nelson County and nearby areas, with reliable reports of 27–28 inches in the central part of Nelson County (Camp and Miller, 1970; Schwarz, 1970). Other sources have reported unconfirmed amounts of as much as 46 inches (Simpson and Simpson, 1970, p. 31).

The storm lasted about 7 or 8 hours, from around 2030 hours August 19 until 0330 or 0400 August 20. Except for a brief lull at about 2330 or 2400 hours, the rainfall, thunder, and lightning were intense during the entire storm. A particularly strong cloudburst struck about 0250, cutting off electricity and telephones for many of the inhabitants. At this time lightning—mainly sheet or horizontal lightning—was so severe that the sky was virtually a continuous bluish-white.

The staggering amount of rain easily ranks among the largest quantities ever measured in the United States. DeAngelis and Nelson (1969, p. 458), for example, compare it to the following:

12 inches in 42 minutes at Holt, Mo., in 1947,
19 inches in 2 hours, 10 minutes at Rockport, W. Va., 1889,
22 inches in 2 hours, 45 minutes at D'Hanis, Tex., in 1935,
31 inches in 4 hours, 30 minutes at Smethport, Pa., in 1942,
34 inches in 12 hours, at Smethport, Pa., in 1942.

Schwarz (1970, p. 853) states that Camille's rainfall in Virginia was within about 80–85 percent of the estimated maximum possible rainfall, for areas up to 1,000 sq mi over a 12-hour period. Yarnell (1935, p. 50–51) presents figures showing that the maximum 8-hour rainfall for central Virginia might be 3.5–4.0 inches once every 10 years and 5.5–6.0 inches once every 100 years. After 25 years of additional data, Hershfield (1961) presented charts which for the eastern U. S. are nearly the same as Yarnell's. Hershfield's curves show that for central Virginia a rainfall of 7 inches in 12 hours (no 8-hour data given) would be expected only once in 100 years. Thus the Camille rainfall in Virginia was about 3–4 times the amount expected once every 100 years, for an 8-hour storm. Thompson (1969) reported that for Virginia the amount of rainfall associated with this storm occurs, on the average, only once in more than 1,000 years!

THE FLOOD

Comparisons of this flood with previous floods in the James River basin are difficult because of the different lengths of record at the gaging stations. A comment by Camp and Miller (1970, p. 21) indicates the magnitude of the 1969 flood: "The flow of the James River at Richmond peaked at 222,000 cubic feet per second and is considered to be the second highest discharge on the James River since Jamestown was settled in 1607." Towns along the James River and its main tributaries were inundated to depths of as much as 14 feet.

At Buena Vista, a town about 15 miles west of Nelson County on the Maury River, the measured flood discharge of 105,000 cfs (cubic feet per second) is not likely to be equaled or exceeded, on the average, more than once every 130 years (Camp and Miller, 1970). When one considers that the flow on the James River at Richmond was the second highest since at least 1607, discharges as large as the 1969 flood occur at Richmond an average of about once every 180 years. These estimates of 130- and 180-year floods provide a very rough idea of the recurrence frequency of the Camille flood at stations on two major rivers, the Maury and the James, near Nelson County.

There are no streamflow records for most streams in Nelson County, where much of the heaviest rainfall occurred. Data for the few sites where gaging stations have been established go back only a few decades. For the nine sites in the area at which a peak discharge measurement was obtained (six measurements were by indirect methods), the August 1969 flood exceeded the previous record by factors ranging from 1.5 to 8. The discharges in the deluged upstream basins undoubtedly are less common in those basins than the 130- and 180-year occurrences estimated for the Maury and James Rivers.

This study unfortunately has no water-discharge measurements in the upstream reaches of the drainage basins, where much of the sediment movement occurred. Flow depths, according to eyewitness accounts, sometimes changed radically within minutes. One farmer described the flow in Cub Creek as increasing in depth by intermittent jumps of as much as 3 feet at a time. A resident of Woods Mill reported that the water in his back yard rose about 5 feet in 20 minutes. Another Woods Mill resident estimated that the water near his house rose about 8 feet in less than 30 minutes. Even with allowances for errors in estimates, the discharge of even the larger streams undoubtedly changed rapidly, and in the mountains the discharge of streams draining only a few hundred acres probably changed even faster as a result of the impact from the debris avalanches (Guy, 1971).

Even if the discharge at a given site had been monitored closely throughout the flood, the investigator would have the problem of what discharge to relate to the resulting sediment deposits.

ESTIMATED STORM-AVERAGE DISCHARGES

Measured rainfall amounts can provide a rather crude estimate of the average water discharge for the duration of the storm for the smaller upstream drainage basins. This method involves the so-called rational formula

$$Q=cIA_d$$

where Q is water discharge or surface runoff (in acre-inches per hour, for this formula), c is a coefficient representing the proportion of total rainfall that runs off the ground surface, I is average rainfall intensity in inches per hour and A_d is drainage area in acres. Flow rate in acre-inches per hour is about equal to flow rate in cubic feet per second. The coefficient c is usually estimated on the basis of previous experience with similar areas and reflects the role of infiltration losses, surface detention, valley and channel storage, and general physical characteristics of the drainage basin (Foster, 1948, p. 343). The effects of these factors probably were minimized for this heavy storm, especially for the small steep drainage basins in the study area. Reasonable values for c therefore are taken to be about 0.90 for basins of 100 acres or less, decreasing by 0.02 for each additional 100 acres of basin size up to 2,000 acres. In the upstream regions the flood probably lasted from an hour or so after the onset of heavy rainfall to an hour or so after the rain stopped, a period of about 7 hours. According to the rainfall data mentioned above, plus the figures which Camp and Miller (1970) listed, the average total rainfall over the general study area can be taken as about 20 inches. If 20 inches accumulated during the 7-hour period, the average intensity I was about 2.9 inches per hour.

The mountain drainage basins (fig. 1) studied in some detail in this report, selected because they showed some of the most widespread sediment erosion and deposition, are Ginseng Hollow (drainage area A_d=0.667 sq mi), Polly Wright Cove (A_d=0.953 sq mi), Wills Cove (A_d=1.577 sq mi), and Freshwater Cove (A_d=2.333 sq mi). (Some neighboring regions, such as the Davis Creek area and Fortunes Cove, also experienced considerable erosion and deposition and might have been included in the investigation had time permitted.) If one uses the rational formula with the assumption just mentioned, the approximate 7-hour average discharges were 1,500 cfs for Ginseng Hollow, 1,400 cfs for Polly Wright Cove, 2,100 cfs for Wills Cove and 2,700 cfs for Freshwater Cove.

ESTIMATED PEAK DISCHARGES

Investigations made after the flood usually, of necessity, deal with the peak discharge. Two ways to estimate the peak discharge in the upstream reaches of interest are (a) to extrapolate a best-fit line relating drainage area to measured discharges for downstream sites or (b) to measure the channel slope and estimate the cross-sectional flow area for each study reach, for use in the Gauckler-Manning formula. Both of these attempts were fruitless in the present case. A plot of drainage area versus peak discharge per square mile, with data for the few downstream sites measured, showed too much scatter to permit a reliable extrapolation to small drainage areas. The slope-area method faltered over a reliable estimate of the resistance coefficient, as there was no way to determine this value for a stream carrying and depositing particles ranging from clay to large boulders, along with trees and other debris.

A third possible way to estimate the peak discharge in the small upstream drainage basins is to resort again to the rational formula $Q=cIA_d$, in which I represents a peak rate of rainfall intensity and c and A_d have the same values used above. If one may judge from the data which Jennings (1950) compiled, a reasonable estimate of the peak rate of rainfall intensity in the basins studied here is about 25 inches per hour, even though such an intense burst probably did not last for more than 5–10 minutes. If this peak intensity and the assumptions regarding c are valid, the rational formula yields approximate peak discharges of 9,100 cfs for Ginseng Hollow, 12,200 cfs for Polly Wright Cove, 18,200 cfs for Wills Cove and 23,100 cfs for Freshwater cove, or about 6–9 times as large as the values (estimated above) for the 7-hour average discharge. These peak values approximate the Myers rating of 100, that is, the peak discharge per square mile is about equal to $10{,}000/A_d^{0.50}$. Some indirect measurements which Camp and Miller (1970, p. 59) reported for this storm also approach this rating; so the rational-formula assumptions have some support.

All of the above discharge estimates are for water alone, exclusive of the large amounts of sediment and vegetative debris contributed by the hillsides and stream channels.

GEOMORPHOLOGICAL FEATURES OF DRAINAGE BASINS

The amount and extent of the erosion and deposition can be better understood if some background information is provided on the landscape and stream-channel characteristics of the study area. The following concerns selected geomorphological aspects and their interrelations.

SLOPES OF DRAINAGE AREAS

Steep hillslopes provided excellent conditions for mass wasting. Figures 2 and 4 indicate the general steepness of the regions studied.

The areal distribution of slopes for the three drainage basins of figure 4 is shown in figure 5. These slopes, listed in table 1, were obtained from the contour maps (scale 1:24,000) by map measurements of the local slope (50 ft upslope and 50 ft downslope) at each of many grid points, shown as black dots on figure 4 and spaced 400 feet apart. Subbasin drainage areas were outlined as in figure 4, and the areas were measured by planimeter. The general slope representing each such drainage area is the average of all local slopes (grid points) within that area.

TABLE 1.—*Drainage area and average surface slope in tributaries and subbasins of Wills Cove, Ginseng Hollow, and Polly Wright Cove*

Drainage basin	Drainage area (sq mi)	Average surface slope (percent)
Wills Cove		
West tributary	0.238	46.2
Middle tributary	.145	46.5
Falls tributary		
Subbasin 1	.040	---
2	.037	---
3	.029	---
4	.068	---
5	.046	---
6	.030	---
7	.026	---
8	.052	39.5
9	.047	45.0
10	.033	41.4
11	.027	48.0
12	.024	40.0
13	.114	39.1
14	.060	41.5
15	.032	---
16	.063	---
17	.064	---
18	.014	---
19	.022	---
20	.006	---
21	.008	---
22	.038	---
Subtotal (1–22)	.880	---
Noncontributing	.314	---
Total, Falls tributary	1.194	
Average slope, basins 8–14[1]		41.3
Total Wills Cove	1.577	---
Ginseng Hollow		
Lower	0.172	48.1
North tributary	.086	42.7
South tributary	.115	43.9
Main> 3400[2]	.222	38.0
East bowl	.082	32.1
Total	.677	
Average slope, entire basin		41.5

See footnotes at end of table.

TABLE 1.—*Drainage area and average surface slope in tributaries and subbasins of Wills Cove, Ginseng Hollow, and Polly Wright Cove*—Continued

Drainage basin	Drainage (sq mi)	Average surface slope (percent)
Polly Wright Cove		
Tributary 1	0.043	49.3
2–5	.086	38.6
6	.037	39.3
7	.026	40.0
8	.069	45.8
9	.060	45.5
10	.034	53.0
11	.037	54.2
12	.043	47.5
13	.024	52.0
Total (1–13)	.459	
Average slope, tributaries 1–13		45.6
Lower	.130	37.7
Noncontributing	.364	21.8
Total, entire basin	.953	
Average slope, entire basin		35.4

[1] Weighted according to contributing drainage area.

[2] Refers to all drainage area upstream from a station 3400 ft up the main channel, the measurement of this distance having begun at the mountain front (apex of alluvial fan).

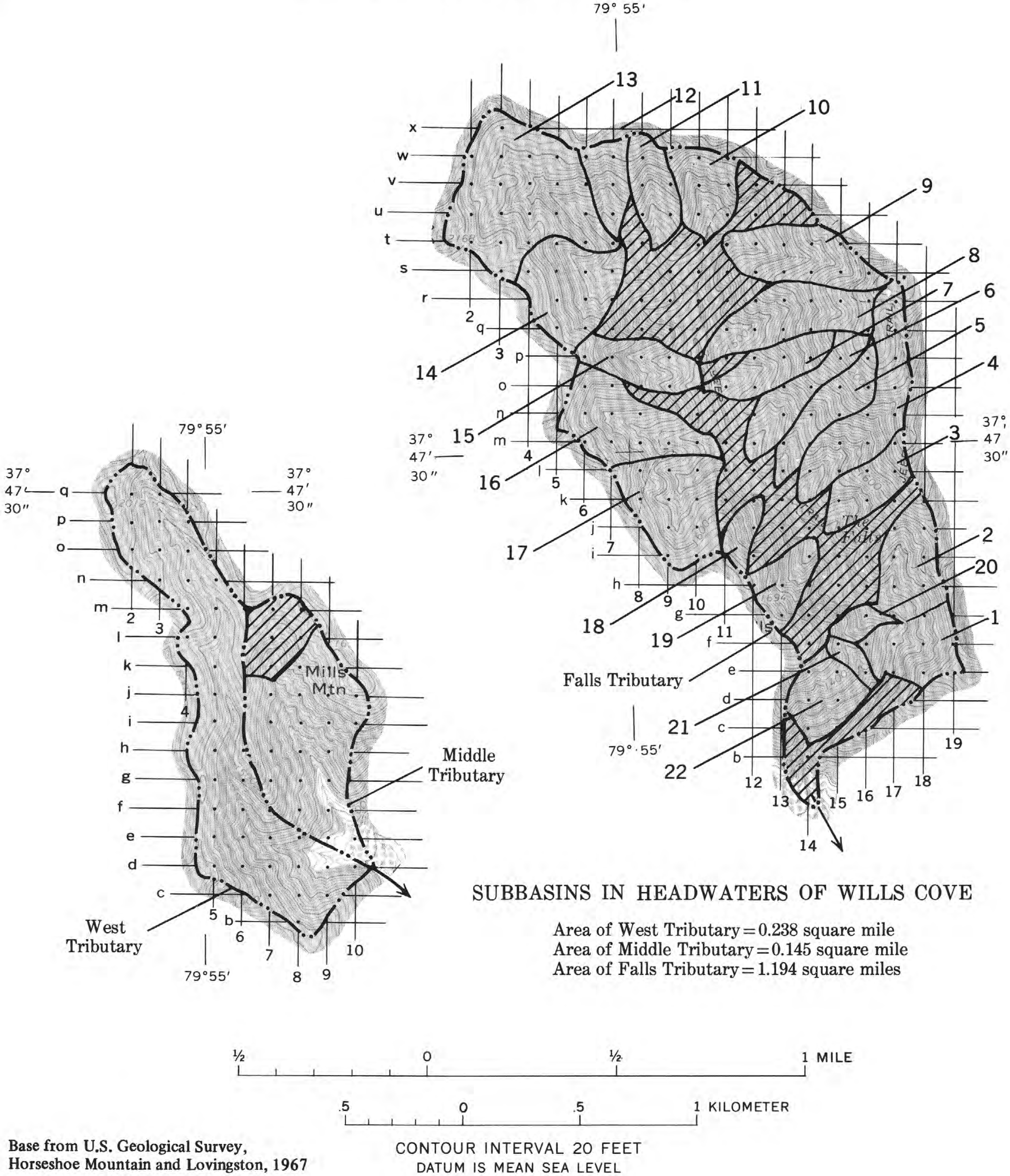
The terrain in Polly Wright Cove is virtually the same as in the other basins, but the curve in figure 5 shows a lesser part of the total drainage area with steep slopes for Polly Wright Cove because the map measurements included more of the flood plains of the basin.

Hillslope erosion appeared to be negligible or nonexistent in tributaries or subbasins having an average surface slope of about 35 percent or less.

STREAM LENGTH AND DRAINAGE AREA

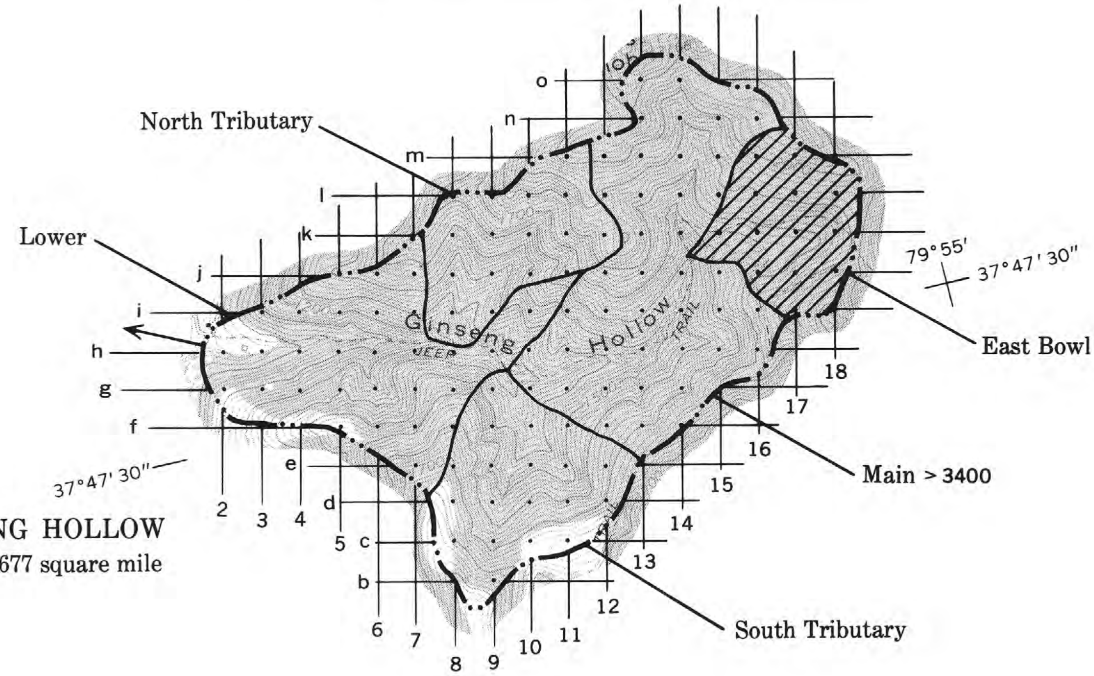
The time for debris eroded near the stream headwaters to arrive at a given downstream site depends in part on the distance along the stream channel which the debris must travel. Also, the amount of water and sediment which a stream carries depends partly on the stream's drainage area. Thus, stream length and drainage area are important factors in the geomorphology of a flood-damaged region.

Various geomorphic studies have established that stream length L is approximately related to drainage area A_d, where L is measured from topographic maps (scale 1:24,000 in this case) as horizontal distance from drainage divide along the channel of the longest stream above the station. Hack's data (1957,

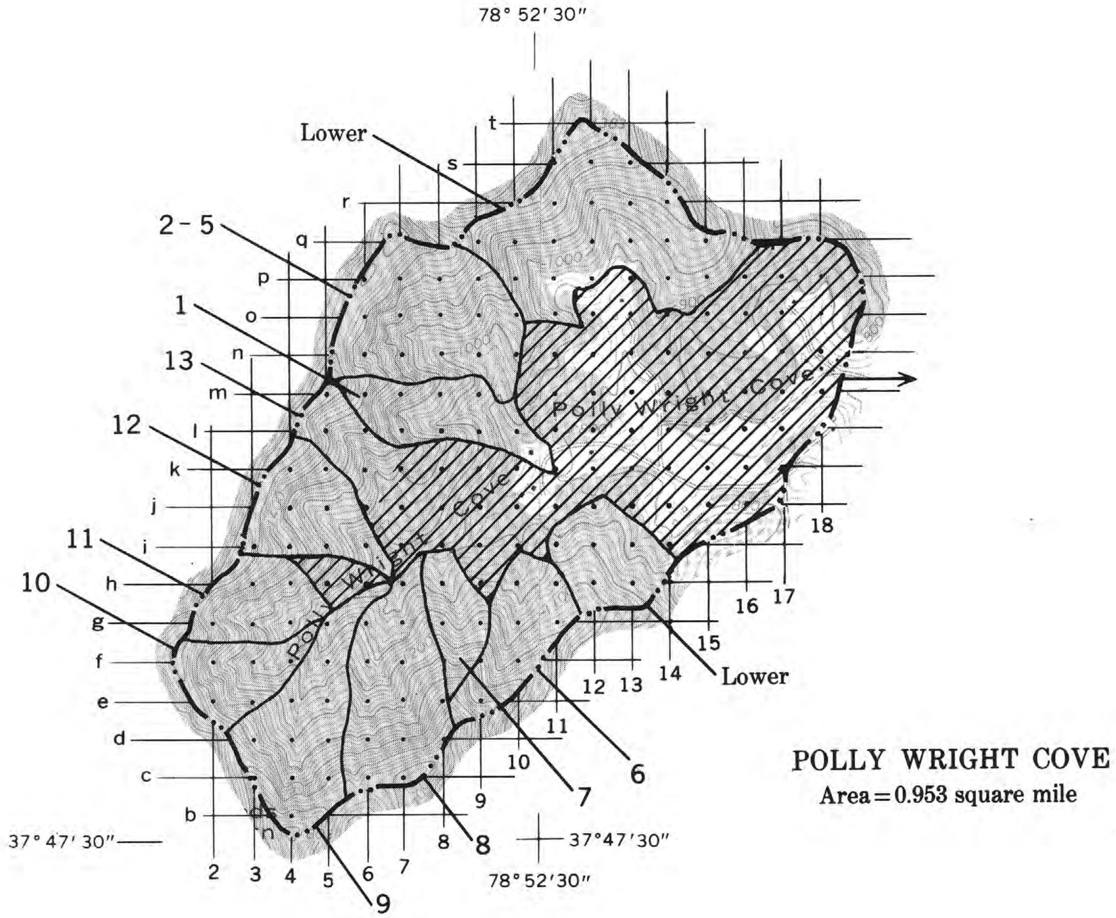


Base from U.S. Geological Survey,
Horseshoe Mountain and Lovingston, 1967

FIGURE 4.—Above and right. Contour maps of Ginseng Hollow, Polly Wright Cove and Wills Cove basins, showing locations of grid points (dots) used to determine average surface slope for tributaries and subbasins. Subbasins are outlined with solid lines. Arrow at mouth of each basin shows flow direction. Notable hillslope erosion lacking in patterned areas.



GINSENG HOLLOW
Area = 0.677 square mile



POLLY WRIGHT COVE
Area = 0.953 square mile

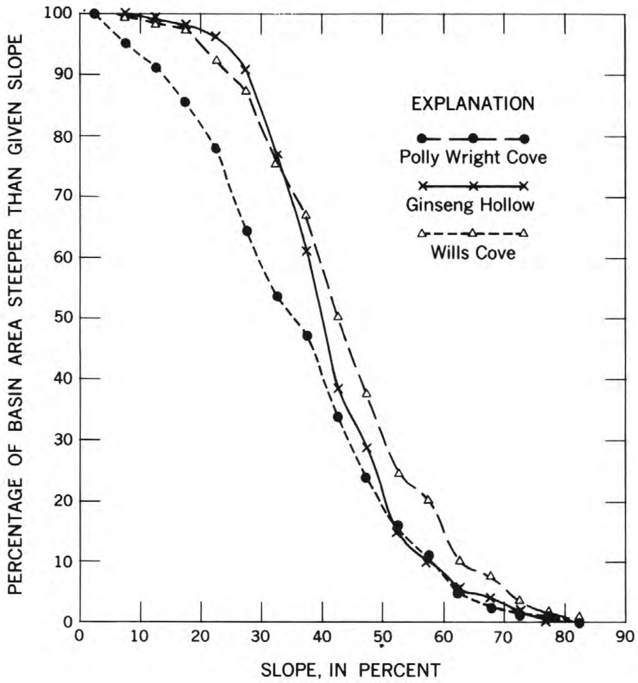


FIGURE 5.—Percentage of drainage-basin area steeper than given slope for Polly Wright Cove, Ginseng Hollow and headwaters of Wills Cove.

p. 64) for some 85 measurements in nearby Virginia and Maryland showed that if all of the points are viewed collectively, then $L \propto A_d^{0.6}$. Hack found this same general relation for 400 observations reported by Langbein (1947) for the northeastern United States. Two other areas which Hack tested—one in Arizona and one in South Dakota—also had power relations but with an exponent of 0.7. Mueller (1972) studied 65 large drainage basins (ranging from 5,000 to 5,000,000 sq mi) from various continents and reported the general relation $L \propto A_d^{0.466}$.

Figure 6 is a plot of stream length versus drainage area for three of the major streams studied here (see fig. 1). A general relationship in which L varies approximately with $A_d^{0.6}$ probably could be fitted to the complete group of points. But the scatter and range of data are such that other exponents might also apply. More interesting is the fact that the points for any one stream show little scatter about an eye-fitted line for the particular stream (values of L from the equations in figure 6 can be determined to within about ± 15 percent or better). The exponents of lines drawn for individual streams show considerable variation, ranging approximately from 0.4 to 0.7. Hack (1957) also found significant departures from his general relation $L \propto A_d^{0.6}$. Consequently, although general relations are often useful and interesting, some attention also should be given to the extent to which individual streams depart from the general relation.

A low exponent means a lesser rate at which stream length increases, for a given enlargement in drainage area. Conversely, for given increments of stream length, the stream with a lower exponent drains relatively larger areas as distance increases. Low exponents can reflect a lack of meandering (smaller increases in stream length) for given increases in drainage area, or they might suggest a relatively rapid enlargement of drainage area downstreamward. The latter is the reason for Dillard Creek's low exponent of 0.42.

LONGITUDINAL PROFILES OF STREAM CHANNELS

The longitudinal profie of a stream, a plot of distance versus elevation, shows how the channel slope changes with distance and is one of the basic characteristics of a stream channel.

TOPOGRAPHIC MAP DATA

Longitudinal stream-profiles of five prominent channels were measured from topographic maps (scale 1:24,000). The profiles extend for distances of 3–8 miles downstream from the drainage divide.

Figure 7 shows these longitudinal profiles (fall F versus horizontal distance L) on arithmetic scales. The characteristic concave-upward profiles are smooth and regular for three of the streams (Campbell Hollow, Rucker Run and Polly Wright), but the other two (Ginseng Hollow and Wills Cove) include a pronounced deviation or local steepening.

Many longitudinal profiles of streams and alluvial fans plot as straight lines on semilog paper. For some streams the elevation (ordinate) must be on the log scale, whereas for others the distance (abscissa) must be on the log scale. Krumbein (1937), Shulits (1941), Yatsu (1959), Leopold and Langbein (1962), and Fok (1971) dealt with the former type; their studies were concerned mainly with graded or well-developed rivers rather than mountain streams. Semilog plots of the profiles shown in figure 7, with elevation on the log scale, have a strong curvature, and no constant could be found (computed) to rectify the curves.

Hack (1957, p. 70; also Hack and Goodlett, 1960, p. 11) and Leopold and Langbein (1962, p. A9) dealt with the second type of plot, where elevations vary with the logarithm of distance. A diagram of this sort, an exponential or depletion function, describes the five profiles of figure 7 in a rather approximate way. The plotted points, not shown here, have a sinuous path, and a straight line fitted by eye predicts the altitude to within about ± 17 percent.

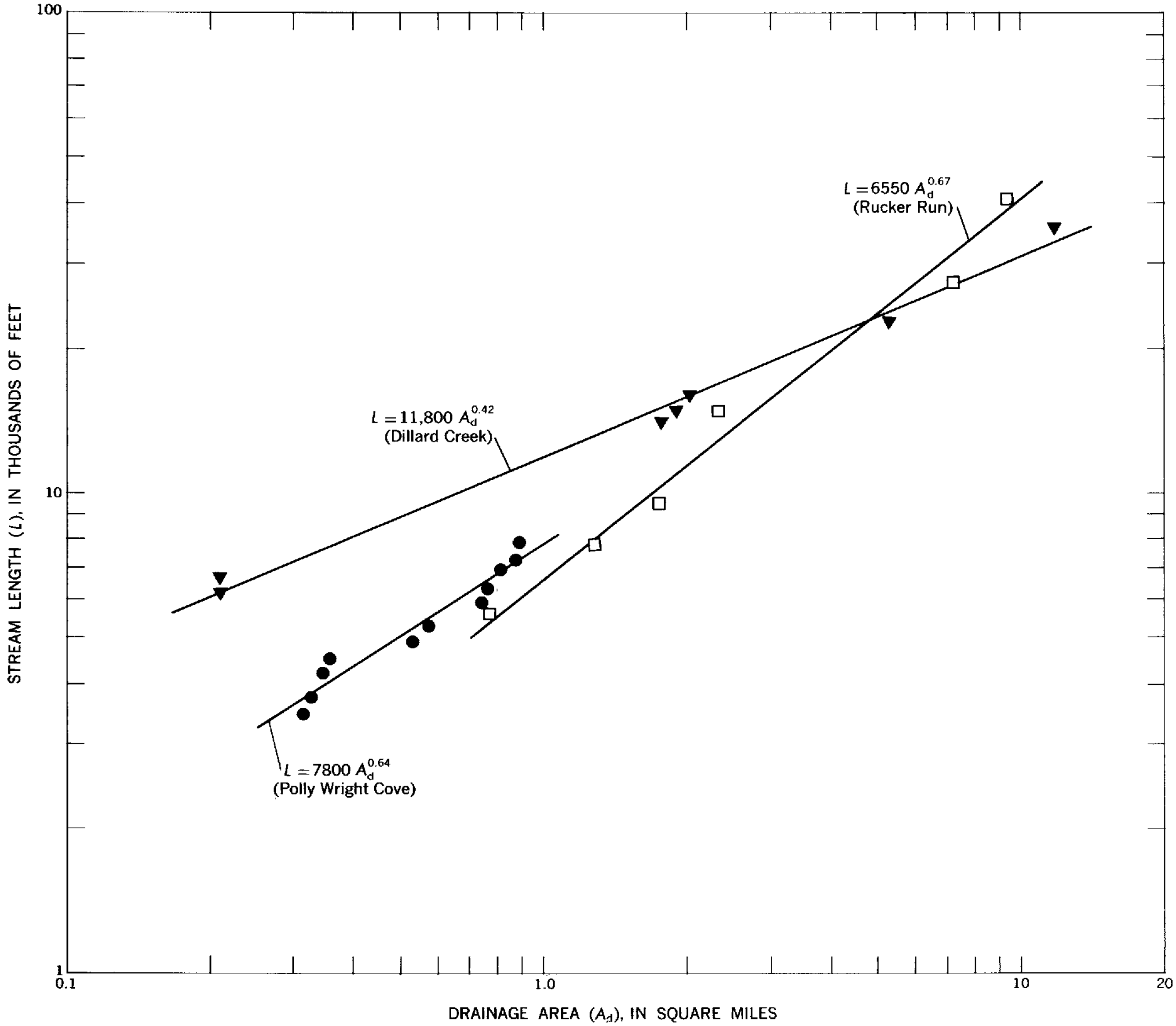


FIGURE 6.—Variation of stream length with drainage area for three Nelson County basins.

For the range of elevations involved with these stream profiles, the actual values of the elevations are approximately proportional to the logs of the elevations. Consequently a graph of elevation (not vertical fall) versus L on log paper, not included here, also shows an approximate straight-line relation. The alinement of points on such a logarithmic graph is slightly better than on the semilog plot. A straight line drawn by eye predicts the altitude to within about ±13 percent or better, with the possible exception of the point nearest the drainage divide. (Incidentally, plotting the fall F versus channel length L on logarithmic paper did not produce a straight line, nor could any constant be found to rectify the curve.)

The type of plot which best expresses the elevation or fall along the five streams is a hyperbolic equation, like the power law, and takes the form

$$F = L/(a + bL)$$

where a = the extrapolated value of the ordinate at $L = 0$ and b is the slope of the best-fit line on the graph. Rearranging the equation into the form $L/F = a + bL$ gives the type of diagram on which the data plot as a straight line (fig. 8). (An alternative way of rearranging the equation is $1/F = a(1/L) + b$, which indicates that a plot of $1/F$ versus $1/L$, not included here, also gives a straight line on arithmetic scales. This variant of the plot is not practical because much of the stream length is squeezed into a small section on the abscissa.)

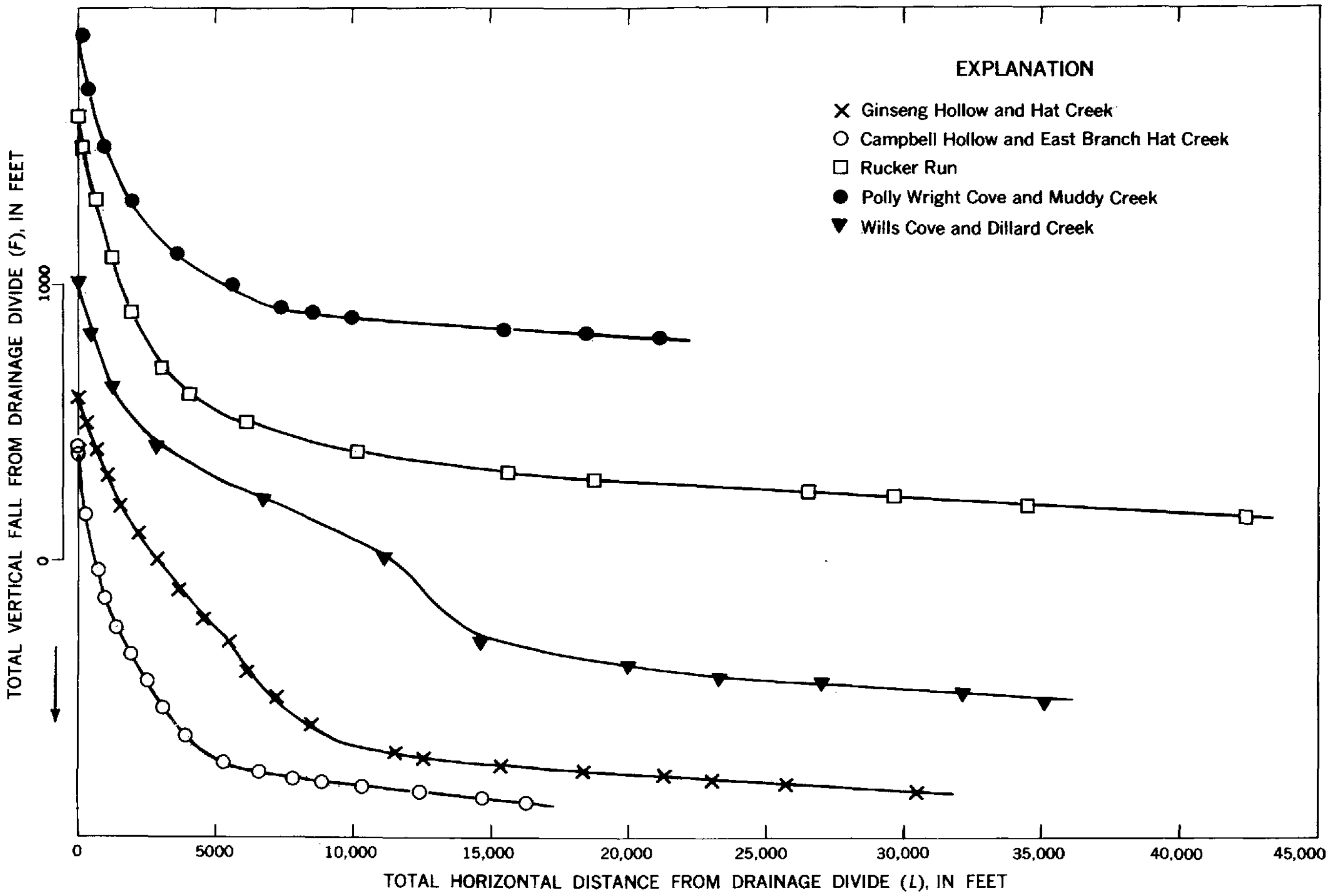


FIGURE 7.—Longitudinal profiles of stream channels, based on data from topographic maps. Vertical exaggeration is × 10.

One distinctive feature of this diagram is the striking linearity of data points. Second, the local steepening in Ginseng Hollow is quite apparent and divides the profiles into two segments. This local steepening was much less recognizable in the semilog and log plots. Last, the first data point on the Campbell profile (100 ft from the drainage divide) is markedly offset from the best-fit line. The ratio L/F for this point is higher than would be expected, so for this particular horizontal increment the fall is less than would be expected. Probably the concave profile is not as well-developed this close to the drainage divide.

After some algebraic manipulations, the equation $F=L/(a+bL)$ assumes a form very close to one of the equations which Hack (1957, p. 70, eq. 11) found for stream profiles in nearby areas of Virginia and Maryland. The pertinent Hack equation is for streams in which the average size of the bed particles decreased downstream, as is the case in the streams studied here. Written in logarithmic form, Hack's equation is

$$\log (F-C) = \log (k / (n+1)) + (n+1) \log L$$

where C is a constant of integration and k and n are also constants. The present equation, rearranged and put in logarithmic form, is

$$\log (F-\alpha) = \log \gamma - \log (L+\beta)$$

in which $\alpha=1/b$, $\beta=a/b$ and $\gamma=(-a/b^2)$. This is essentially Hack's equation with $n=-2$ except that the right-hand side has $L+\beta$, rather than L.

The coefficients a and b both reflect such features as the channel steepness and do not lend themselves to simple descriptive labels. Values of a in this study ranged from 1.1 to 3.1, and b-values ranged from 0.000600 to 0.000870. Eye-fitted lines predict the fall F to within about 9 percent error and usually to much greater accuracy.

The appearance of the local steepening in Ginseng Hollow on the L/F versus L diagram might seem to suggest that this type of plot is sensitive to local channel aberrations. This, however, is true only near the drainage divide, where the distance L is still rather short and F increases significantly. The apparent sensitivity virtually disappears after L has increased to about 2 or 3 miles, by which stage F usually increases by relatively minor increments.

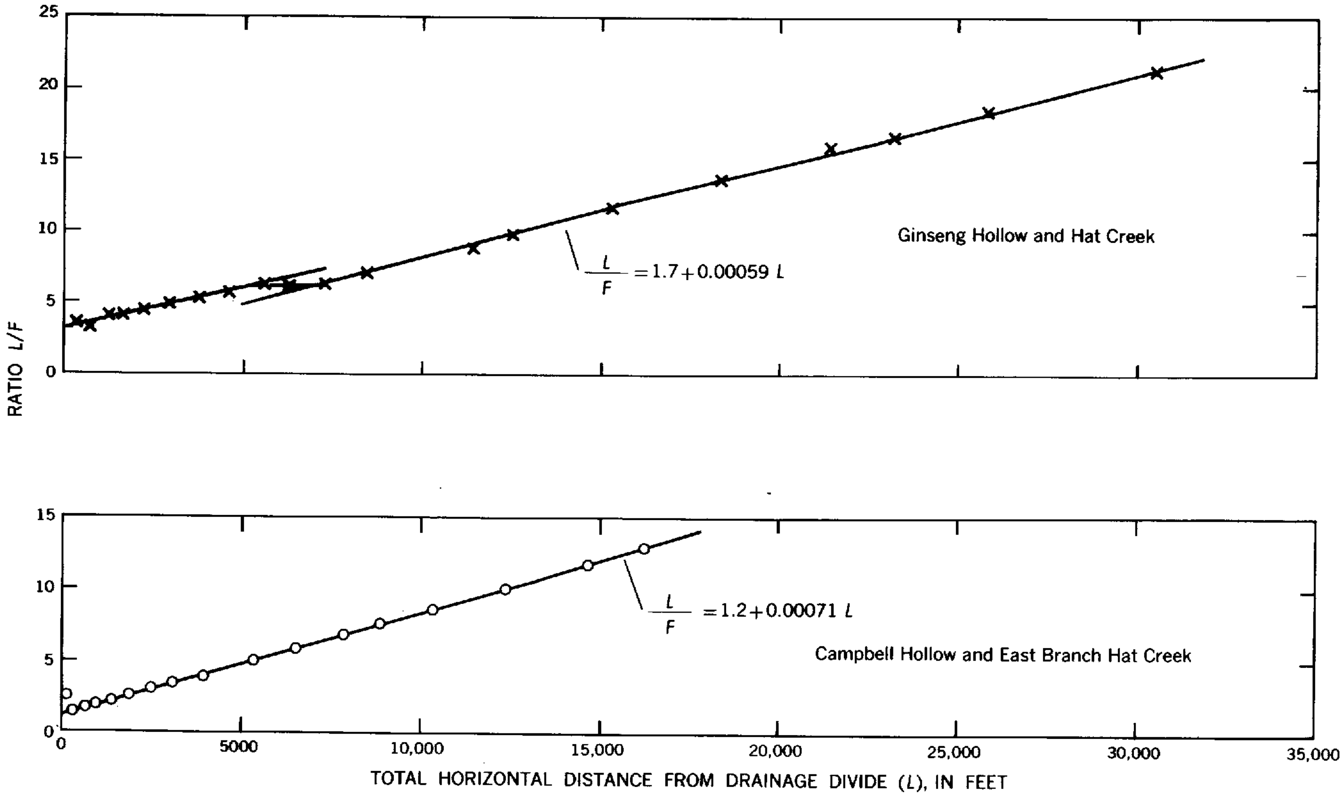


FIGURE 8.—Longitudinal channel profiles plotted in the form of L/F versus L (topographic map data).

FIELD MEASUREMENTS

On three of the channels—Ginseng, Polly Wright and Campbell—more detailed longitudinal profiles were measured in the field with a stadia rod and either a Zeiss level (for flatter reaches) or a hand level. Distances along the channel are virtually the same as horizontal distances and will be so treated in the following discussion. The distance and fall from the drainage divide to the uppermost channel station were taken from the topographic map.

These longitudinal profiles of mountain channels and, where present, their adjoining alluvial fans have the usual concave-upward shape when plotted on arithmetic scales (fig. 9). The local steepening in Ginseng Hollow, discussed earlier, appears in the upstream part of the Ginseng profile. The profiles tend to consist of straight-line segments, a feature which Bull (1964) pointed out for alluvial fans in California. The present field data show that this segmented characteristic applies to the channel upstream from a fan, in addition to the fan itself. Furthermore, the profiles show no sudden change in slope at a fan apex, and the location of the apex cannot be found from studying the profile. This is also a feature which Bull (1964, p. 101) noted for his profiles in California. And Denny (1965, p. 55) remarked that the mountain channels and alluvial fans which he studied in California and Nevada have a smooth profile with no break in slope at the fan apex. As Morisawa (1968, p. 94) noted, the primary cause of fan deposition therefore is not a sudden flattening of channel gradient. Instead, such deposition most likely results from the sudden spreading-out of the flow as it escapes the confining mountain channel.

Profiles on semilog and log paper, not included here, also consist of straight-line segments. The number of segments decreases progressively from arithmetic to semilog to log plots. As with the topographic-map data, a straight line on the log diagram is a better approximation of the longitudinal profiles than a straight line on semilog paper. The L/F versus L relation (fig. 10) describes the data most accurately.

The equations for obtaining the vertical fall F from the drainage divide, in feet, for a given horizontal distance L from the drainage divide, in feet, are:

Ginseng-Bryant:	$F=L/(2.52+0.00050L)$
Polly Wright:	$F=L/(1.36+0.00075L)$
Campbell:	$F=L/(1.20+0.00065L)$

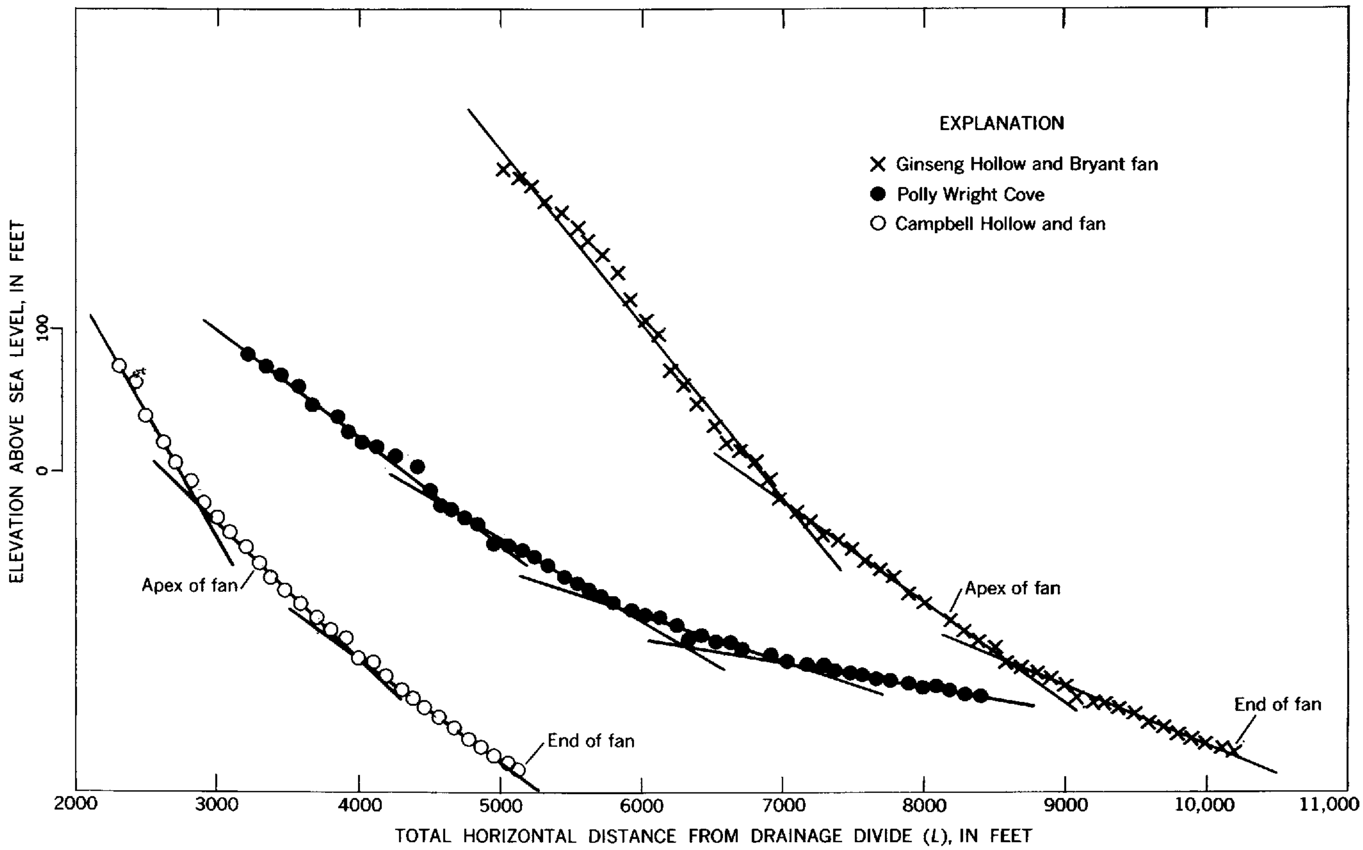


FIGURE 9.—Longitudinal profiles of mountain channels and, where present, adjoining alluvial fans. Data from field measurements. Vertical exaggeration is × 10.

These equations were obtained graphically from figure 10, in which the best-fit lines have been drawn by eye. Except for the local steepened reach in Ginseng Hollow, the equations predict the total fall from the drainage divide to any station on the stream channel to within ±2.5 percent.

In summary, this section has presented measured values of drainage areas, stream lengths and the steepness of the terrain, together with some relationships between these factors. A mathematical expression of the longitudinal profile has been found for several of the main streams in the study area. Inasmuch as some kinds of erosion described in the following section are typical of mountainous regions, future studies using more data may discover some mathematical relation between the geomorphological characteristics of the drainage basins, on the one hand, and the amount and type of erosion expected to result from a given quantity of rainfall. Similarly, when more data are available it may be possible to relate the geomorphological features to the downstream amount and pattern of deposition. Some tentative steps toward these goals are taken in later sections of this report.

EROSION

DEBRIS AVALANCHES

Measurements and estimates of amounts of erosion presented later in this report suggest that nearly half of all the storm-eroded sediment came from downslope-trending strips on the hillsides. Many scars (figs. 11 and 12) testify to the magnitude of this hillslope degradation. The nature of these scars and their association with the heavy rainfall indicate that they were caused by a type of mass-movement which Sharpe (1938, p. 61) called a debris avalanche. A debris avalanche is a rapid down-hill flowage of soil, rock, trees and other vegetation (if present) and water. According to Sharpe, "the typical debris-avalanche has a long and relatively narrow track, occurs on a steep mountain slope or hillside in a humid climate, and is almost invariably preceded by heavy rains." In Sharpe's classification debris avalanches differ from landslides in that landslides involve very little water. The Nelson County avalanches included soil, trees and other plants, water, and usually rocks ranging from gravel to boulders up to 10 feet in intermediate diameter.

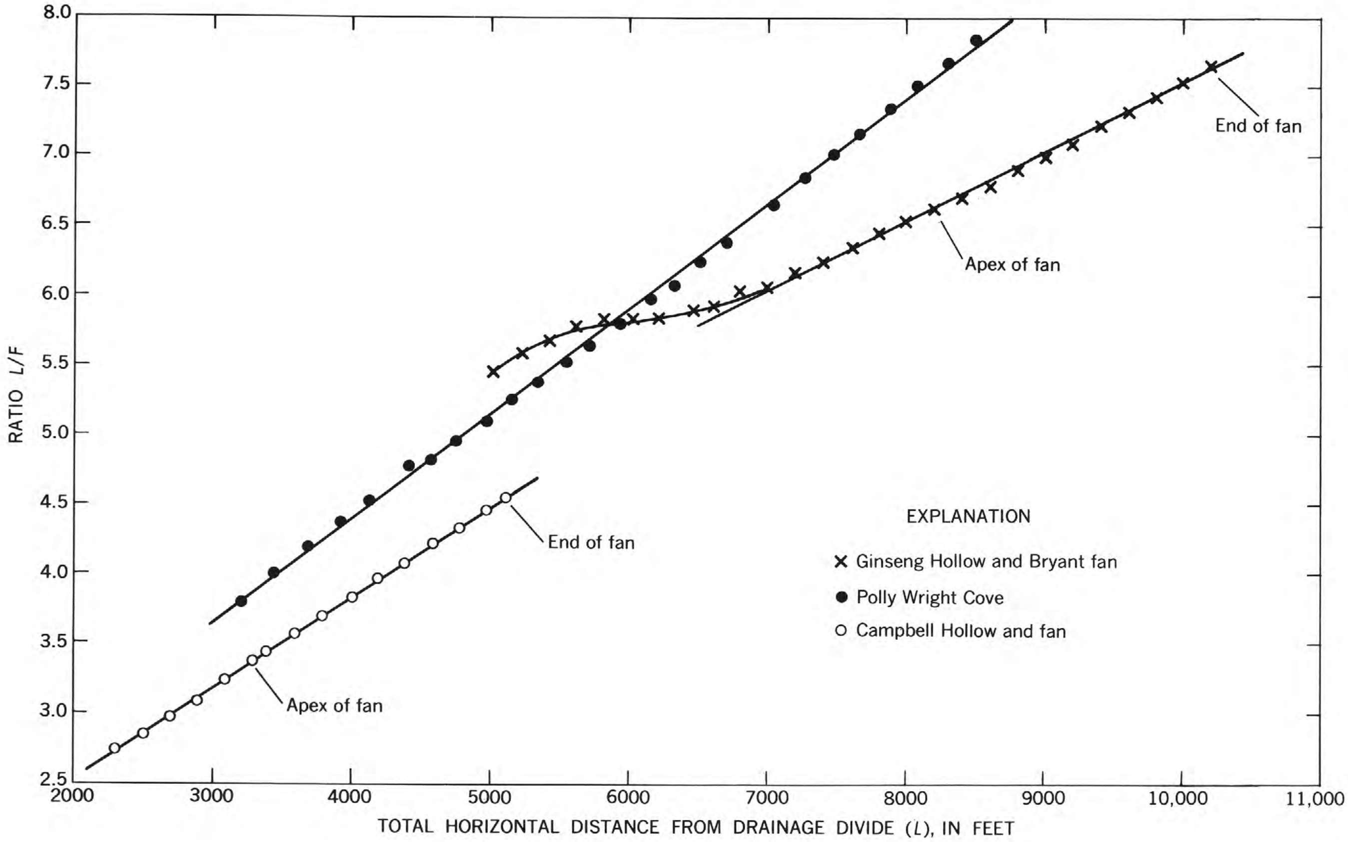


FIGURE 10.—Longitudinal profiles of figure 9 (mountain channels and, if present, adjoining alluvial fans), plotted in the form of L/F versus L. Data from field measurements. Every other point plotted.

A previous report (Williams and Guy, 1971) analyzed various aspects of the 1969 Nelson County debris avalanches.

Hack and Goodlett (1960) described in considerable detail the important role of debris avalanches in the overall adjustment among slope processes, mountain form, and vegetation for a similar central-Appalachian region in Virginia and West Virginia.

Eyewitnesses or survivors of debris avalanches say that the entire event occurs very quickly. Observations during the present storm were obscured because of the darkness. Mr. S. O. Mawyer of Woods Mill was almost swept away in a small avalanche near his home and only saved himself by grabbing a bush that was growing just beyond the slide border. He said the ground started oozing slowly downhill for a fraction of a second, and then the entire section of the hillside suddenly slid quickly down the slope, accompanied by a loud noise. Mr. B. R. Floyd, whose Lovingston, Va., home (a few hundred yards downstream from the base of a long, high mountain) was severely damaged by a 3,000-foot-long avalanche (fig. 13), estimated the duration of the event in the vicinity of his home to be about 3 minutes. Mr. W. E. Davies, a geologist for the U.S. Geological Survey, saw a debris avalanche in a 1949 West Virginia storm and testified (oral commun., 1970) that the whole strip of hillside started moving at about the same instant. These subjective descriptions give some indication of the short duration of a debris avalanche.

Rapp (1963, p. 198–200) quoted an eyewitness description of debris avalanches in western Norway: "A mass of earth, boulders, trees and water moves down the slope and a new slide track is formed * * * The river (at the base of the hill) is filled with a porridge of earth which flows downstream, mixed with a crowd of naked birch stems, twisting and whirling * * * New slides are coming down. It looks like a wave of water that squeezes earth and trees out of the ground and back again. The trees fall down immediately (with their bases pointing downslope, according to Rapp). Then they are pushed together with the earth and boulders on the way downslope, so they reach the river naked, without twigs and bark. Water sprays out in small cascades from the moving earth."

People living in or near the mountains in Nelson

FIGURE 11.—Erosion and deposition caused by the torrential rainfall, Davis Creek area, Nelson County, Va. Scars on hillsides mark locations of debris avalanches. Stream flood plain shows both erosion and deposition. (Photograph by Ed Roseberry.)

County were unanimous in emphasizing the awful roar they heard associated with both the debris avalanches and the swollen streamflow. The noise was variously described as sounding like a squadron of airplanes warming up for takeoff; or continuous discharges of dynamite, one charge right after the other, sometimes overlapping; or many cannonballs rolling along inside a bowling alley, with cracking, popping, and explosions.

The slides occurred at intervals ranging from "one right after the other" or "every few minutes" to "every once in a while" (Simpson and Simpson, 1970, pp. 8–9). The accounts of local residents suggest that the avalanching was more frequent toward the end of the storm. Some slides occurred after the rain had abated to a sprinkle or mist, at about 0400 hours on August 20.

At least the larger debris avalanches caused a very noticeable quivering or trembling of the earth's surface for a radius of several miles.

There is no way to determine how many of the 125 people who lost their lives in Nelson County were killed by debris avalanches. A number of missing houses, particularly in the Davis Creek basin, originally stood in small hollows which were scoured by avalanches. Other lives were lost when buildings and property hundreds of feet downstream from hillslopes were ravaged by avalanche water and debris. Dr. J. H. Gamble of Lovingston, who was in charge of identifying and examining the flood victims, found that almost none of the dead showed evidence of drowning. "The vast majority were just multiple fractures—rib . . ., spine . . ., large bone . . ., skull fractures—from rocks and trees and so forth. Most died from massive 'blunt force' injuries." (Simpson and Simpson, 1970, p. 270). Such testimony shows that sediment damage during floods can be very important in terms of lives as well as property.

SOURCE MATERIAL

The soil type on the mountainsides, according to the U.S. Department of Agriculture classification, is of the order inceptisol—a young soil that presumably has formed rather quickly, mostly from alteration of parent materials (here, the underlying crystalline rocks) and which lacks horizons of marked accumulation of clay, iron oxide, and aluminum oxide. The subdivisions within the inceptisol order would be ochrept and dystrochrept.

The soil usually is about 1–3 ft thick on the steeper parts of the terrain. Logically, the thickness of the profile should vary inversely with the long-term erosion rate and (or) directly with the weathering rate of the rocks. Most profiles on the hillslopes contain some larger rock fragments, including a few of boulder size.

Two or 3 inches of dark-colored organic matter cap the soil profile. Below this organic layer the soil color is mostly red-brown. The texture is that of a soil matrix, rather uniform in size-composition with depth, containing a few larger fragments ranging up to several feet in diameter. Soil samples from the side walls of 10 scars—three in Polly Wright Cove, five in Wills Cove, and two in Ginseng Hollow—were analyzed for particle size. Such samples were small (1 or 2 kilograms) and did not include particles larger than about 25 mm (millimeters). There appears to be no significant difference in particle size among the 10 soil samples. The average for the 10 samples was 15 percent gravel, 48 percent sand, 31 percent silt, and 6 percent clay-sized particles by weight. The median size ranged from 0.08 to 0.28 mm. The larger fragments occur at diverse levels within the soil profile and usually do not show any pronounced trend of becoming larger with depth. Some of these large fragments are on the top surface and probably move downslope by creep. No representative size-frequency distribution was obtained for these large rocks.

GENERAL DESCRIPTION OF AVALANCHE SCARS

Typical avalanche scars range from about 200–800 feet in horizontal length, 25 to 75 feet in width and 1 to 3 feet in depth. Some scars are as short as 20 feet or as long as 1,000 feet, as wide as 200 feet, and as deep as 15 or 20 feet. Average gradients are steep, ranging in most cases from about 0.30 ft per ft (foot per foot) (16 degrees) to 0.80 ft per ft (39 degrees). The most common average gradients were 0.50–0.60 ft per ft. Based on 12 typical scars that were measured in detail, the average amount of sediment removed in a single debris avalance was about 88,000 cubic feet or about 3500–4500 tons.

From the upslope tip, every debris avalanche continued all the way down the hillside. Most scars become slightly deeper with distance downslope. Widths tend to increase downslope, as do avalanche tracks in other areas (Swanston, 1969); however, there are many exceptions where the width stays about constant or decreases (fig. 12*A*).

The heads of individual scars can be almost anywhere on a hillside, except that not one is located at the crest of a hill. Some avalanches involved only the foot of a hillslope, while others extend nearly to the crest. Most scars originate in the middle third of the slope between the ridge crest and the channel.

From a top view the break at the head of the scar can be curved, irregular, or even a virtually straight line, normal to the downslope direction.

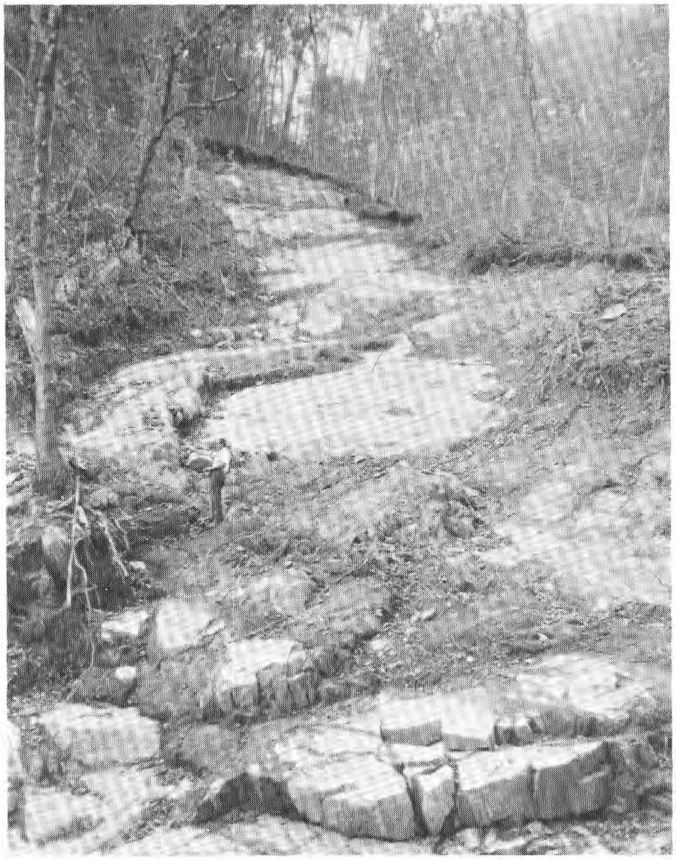
Almost all avalanches removed the entire vegetal cover in their path, including large trees. In many cases the bedrock was exposed along the full length of the scar. On other scars a foot or two of soil still covers the bedrock, particularly in the lower reaches. Some of the avalanches—probably those that occurred near the end of the storm—removed only the vegetation and part of the soil, so that no bedrock was exposed. Many scars eroded to bedrock occur next to scars which still have a partial soil cover (fig. 12*C*).

The break between most slide areas and the adjacent unaffected ground is clean and sharp, especially at the top of the scar. Many roots of the vegetation which had been growing in the eroded soil were broken and exposed. Most of these roots point downslope.

Mud deposits around the base of tree-trunks along the edges of some scars suggest that the top surface of the moving avalanche was as much as 3 feet higher than the normal ground surface.

Nearly all the avalanches in the headwater areas entered first- or second-order streams at the base of the hillslope. These streams begin at the head of a mountain ravine which itself experienced debris avalanching or at least severe channel enlargement during the storm. The streams wind through the mountains for distances ranging from a few hundred feet to as much as several miles, before emerging into the broad intermontane valleys.

Along such mountain-stream channels the banks opposite the scars rarely had deposits from the avalanches, nor was any material left in the channel itself. Thus at some time between the avalanche and the end of the flood, the stream removed nearly all



A

B

C

FIGURE 12 (above and right).—Typical scars left by debris avalanches on hillslopes: *A-C*, headwaters of Wills Cove; *D*, drainage basin upstream from Campbell fan; *E*, Ginseng Hollow.

D

E

the dislodged material. Few, if any, of the avalanches therefore occurred after the flood.

Avalanches on the sides of the broader intermontane valleys continued beyond the base of the hillside for hundreds of feet and in most cases reached the flooded zone in the center of the valley. After the flood subsided, the path of these avalanches was characterized by an incised channel and by sediment deposition along each side of the channel (fig. 14).

The huge volume of soil and rock eroded by debris avalanches is discussed in the section on "Sediment yield."

GEOMETRY OF SCARS

METHODS OF STUDY

We measured the longitudinal profiles, cross profiles, widths, and depths of 12 scars which to the eye appeared to be typical and representative of the various types. For seven of these, the tools were a hand level, tape, and surveyor's rod. From the tops of these seven scars to the base of the hillside, we measured distance down the scar centerline with the tape and read elevations at intervals of 10–100 feet (usually about 40–50 feet), depending on the steepness of the slope. We estimated or measured widths at intervals of 20–100 feet down the scar by extending the 25-foot rod from one edge of the scar toward the opposite edge. At a few stations where the width was not determined in this manner, we simply estimated the total width by eye. We estimated erosion in the center of the scar at all stations either by eye or by holding the rod on one edge of the scar, pointing the other tip toward the opposite edge and assuming that the rod represented the original surface.

The other five scars were mapped with an engineer's transit and surveyor's rod in sufficient detail

FIGURE 13.—Scar of major avalanche that penetrated Lovingston, about 0300, August 20, 1969. (Photograph by Ed Roseberry.)

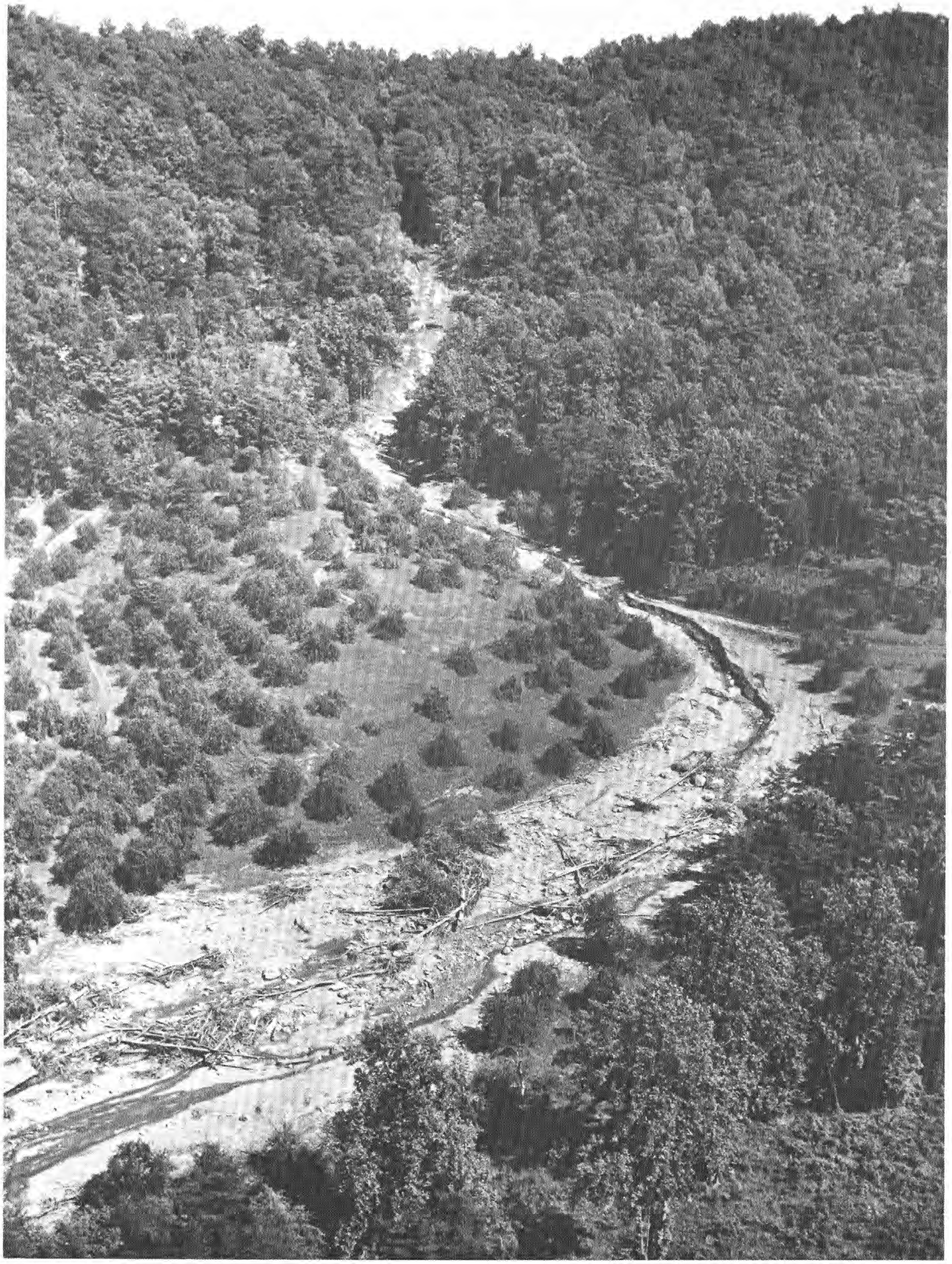
A

FIGURE 14 (above and right).—Debris avalanches along sides of intermontane valleys. *A*, East Branch of Hat Creek. Creek is at lower left foreground. Note deposition of logs, boulders and other particles along route from avalanche to creek. *B*, Stevens Cove. Note damage to orchard. (Photographs courtesy of Virginia Division of Mineral Resources.)

that 10–foot contours could be drawn. With this method the areal spacing of measurements was about the same as with the hand-level method except that additional readings were taken on the nearby undisturbed ground surface, both above the head of the scar and on either side. Figure 15 presents maps of typical scars and locations of cross sections. No depth-of-erosion estimates were made in the field for scars mapped by transit; these depths instead were estimated from the plotted cross profiles.

Elevations and distances from the drainage divide to the head of the avalanche scar were taken from the topographic map after we plotted the total horizontal scar distance, as measured in the field, on the map. The vertical distance from base of hillside (usually taken as the channel edge) to top of scar, as measured in the field, agreed very well with the distance indicated by the contour lines on the map; so for longitudinal-profile purposes the field and map measurements could be combined with no significant error.

A profile measured along the scar centerline in



B

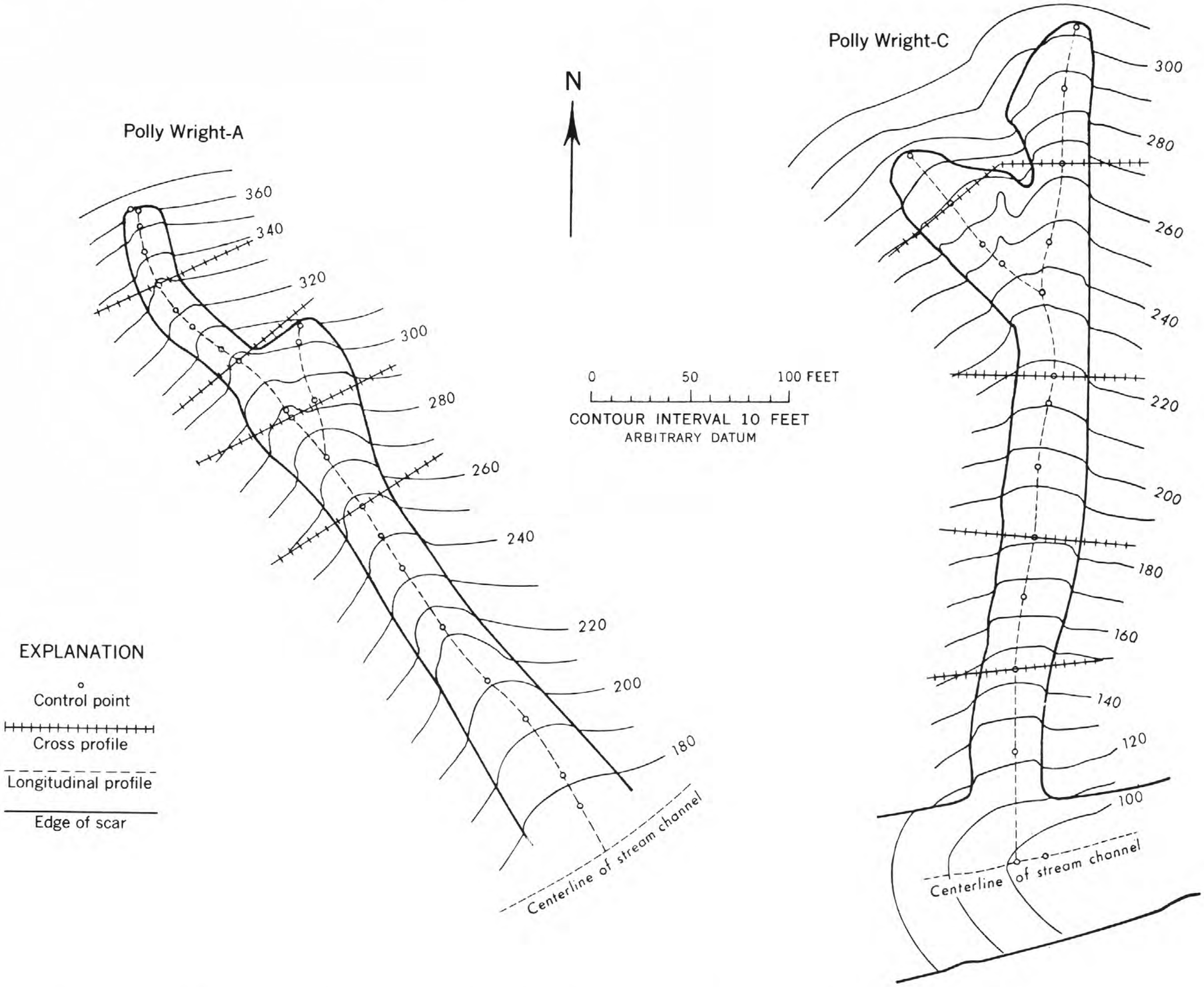


Figure 15.—Contour maps of two avalanche scars, Polly Wright Cove, showing locations of cross and longitudinal profiles.

most cases does not exactly represent the longitudinal profile before the avalanche because of the missing soil and rock in the scar. However, the overall lengths (ranging from 560 to as much as 1,600 feet) and vertical distances (240–599 feet) of the hillsides studied are large compared to the estimated depth of soil removed (generally less than about 5 feet). By adding to the measured elevations the thickness of soil and rock removed, as estimated in the field, we were able to reconstruct the approximate pre-storm profile for seven of the scars. These reconstructed profiles closely approximated those measured in the field. The only difference of any significance might be near the base of those hillslopes from which a particularly large amount of sediment was removed. This potential error might affect the apparent degree of concavity of that region of the hillside. The post-storm profiles as used here probably are quite close to the original profiles, in most cases. The avalanches certainly did not affect the profiles as far as the definition of convex and concave zones on the hillside is concerned.

LONGITUDINAL PROFILES

Figure 16 shows the 12 scar and hillside profiles, all of which have been standardized so that the units of length and elevation are from zero to 1.0. The abscissa is a sliding scale which we shifted systematically for plotting purposes, in order to put all of the

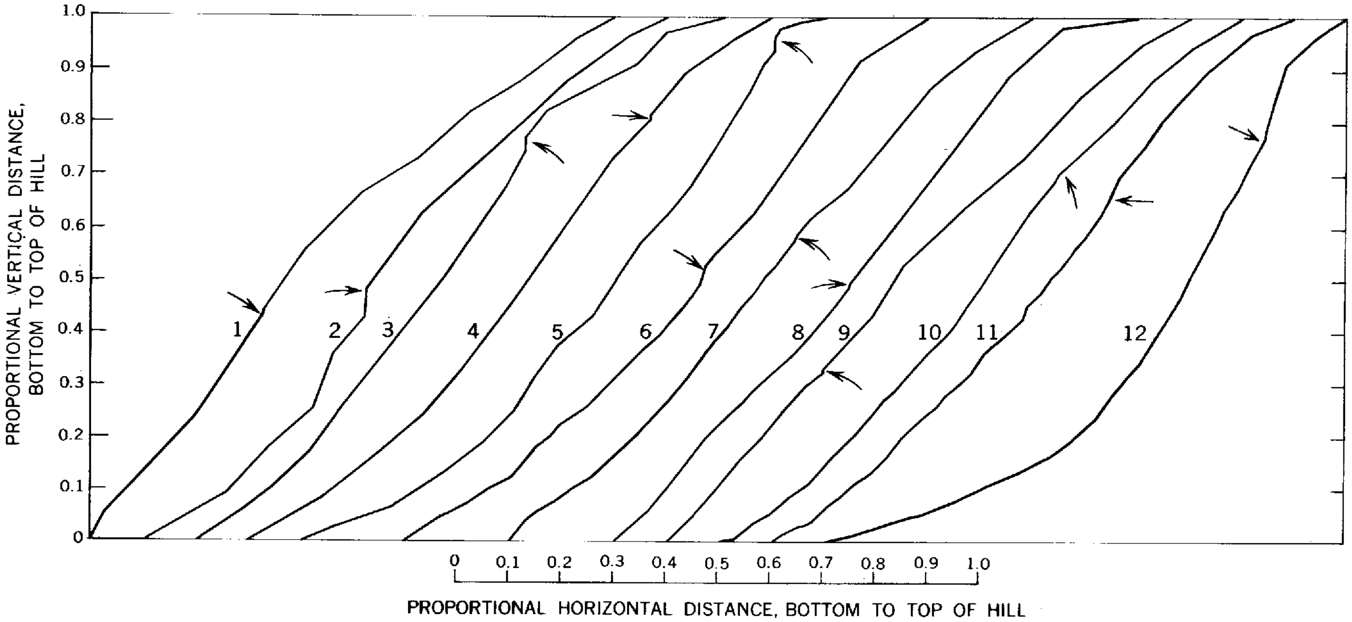


Figure 16.—Longitudinal profiles of hillslopes, including avalanche scars. Vertical and horizontal distances have been standardized to extend from 0 to 1.0. Table 2 gives actual scar and hillside dimensions. Arrow indicates position of head of avalanche scar. See table 2 for profile numbers and scar designations.

profiles on one graph. An arrow on each profile marks the head of the avalanche scar. Table 2 gives the actual scar dimensions.

The profiles in figure 16 show a variety of shapes, some details of which will be described in the paragraphs below. The reader should look at each profile, from base to hillside, and note (a) the general hillside steepness in the region of the head of the scar (arrow), and (b) the elevation of the head of the avalanche relative to the vertical distance from the bottom to the top of the hill. The scars generally tend to begin in the zone where the local gradient is steepest. In these profiles the steepest gradient is located from about 0.3 to 0.95 of the horizontal distance and from about 0.3 to 0.95 of the vertical distance from base to hilltop. The head of the scar occurred on the average at 0.59 of the horizontal distance and 0.62 of the vertical distance from base to hilltop.

Table 2.—*Hillside and scar dimensions and erosion volumes for measured avalanche scars*

Profile number	Scar designation	Hillside dimensions: Length (ft)	Hillside dimensions: Fall (ft)	Scar dimensions: Length (ft)	Scar dimensions: Fall (ft)	Erosion volume (1,000 ft³)
1	Ginseng—2,020	980	435	320	190	18
2	Ginseng—East Br	620	240	260	120	67
3	Ginseng—4,740	900	330	570	250	81
4	Ginseng—Middle	646	310	500	250	144
5	Ginseng—800	830	397	760	390	86
6[1]	Polly Wright—A	690	375	400	200	43
7	Polly Wright—B	560	300	300	170	52
8[1]	Polly Wright—C	930	420	400	200	45
9	Wills—5,850	880	505	260	170	32
10	Wills—1,300	660	385	420	270	100
11	Wills—1,100	1,200	600	870	390	162
12	Wills—7,550	1,620	565	860	380	233

[1] Figure 15 shows contour map of scars.

Hillslopes typically have a convex upper part and a concave lower part. In some cases a middle section of nearly constant gradient separates these two zones; otherwise, the convex and concave portions grade directly into one another. White (1966) has shown that convex, straight, and concave profiles can be most readily defined by plotting local hillslope gradient as a function of horizontal slope distance, on log scales. On such a diagram the convex profile near the crest of a hill in many instances plots as a straight line trending upward (positive exponent), because the local gradient increases with slope distance. A horizontal line on the graph indicates that the local gradient stays constant with distance away from the crest: that is, the hillslope is straight. The concave profile over the lower regions of a hillside means that the local gradient decreases with horizontal distance from the crest, and this type of profile commonly plots on log paper as a straight line sloping downward. All the slopes which White studied had smooth, regular profiles with prominently convex upper regions and concave lower regions.

To examine the 12 profiles in more detail we plotted the data on log paper using White's method and drew lines of best fit by eye. Figure 17 contains four typical plots. Local gradient for any station was reckoned from the arithmetic midpoints between the

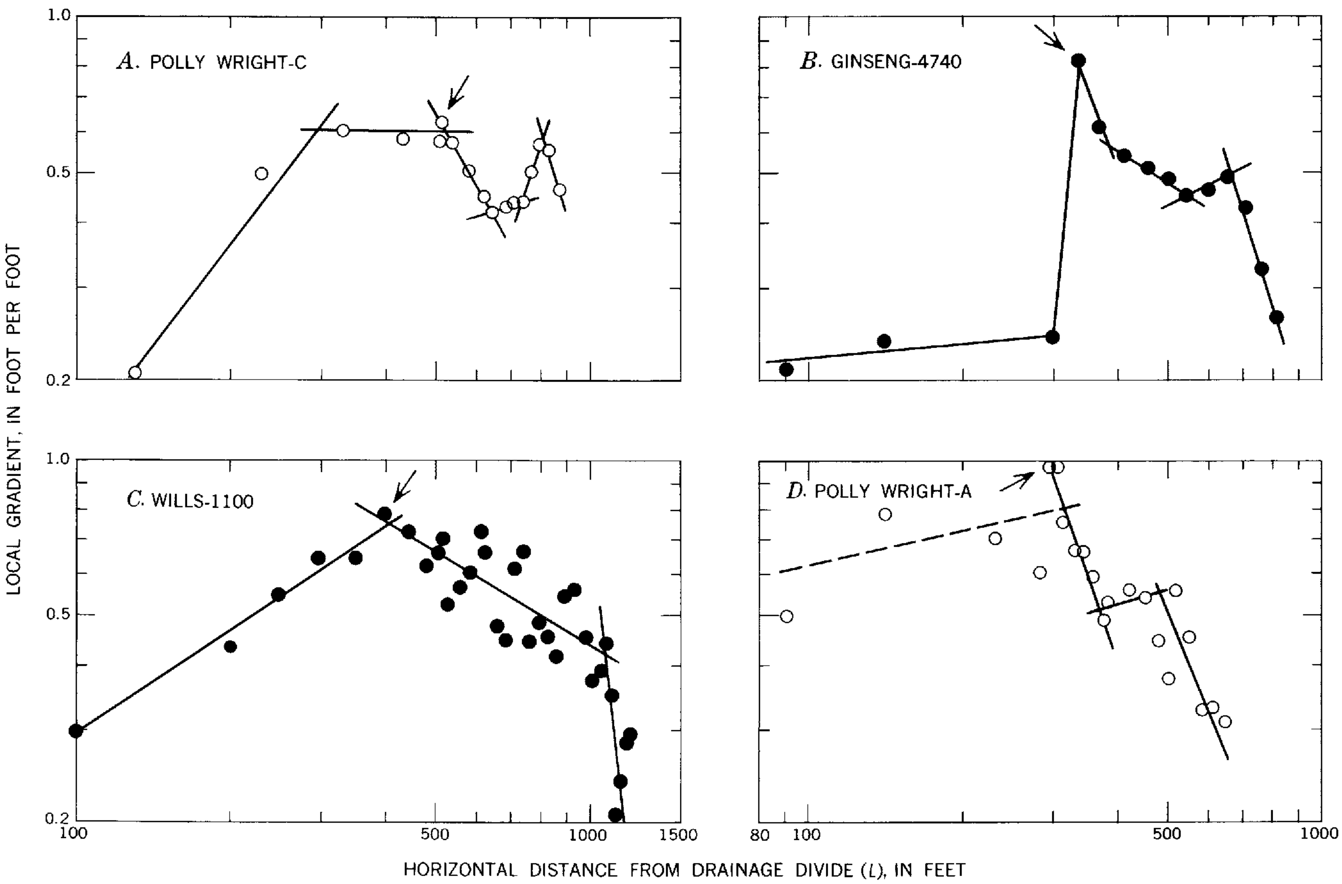


FIGURE 17.—Variation in local hillslope gradient with distance from hilltop, for four typical hillsides on which debris avalanches occurred. Arrow marks head of avalanche scar.

station of interest and the stations immediately above and below. The corresponding value on the abscissa is cumulative horizontal distance L from the drainage divide to the station. The graphs presented a wide variety of slope-types, as might be surmised from figure 16. Some showed enough scatter that various lines, straight or curved, might be fitted to the points, and the difference between convex and straight or between straight and concave segments of the profiles sometimes became a subjective decision.

What features do all the profiles have in common? First, all have a convex zone beginning at the top of the mountain. This upper convex zone varies widely in length—from the hilltop to about 0.1 or up to as much as 0.7 of the horizontal distance to the base of the hill. Some profiles showed two zones in this upper convex region, corresponding to two straight lines (both trending upward but at different rates) on the gradient-versus-length graph (fig. 17*B*).

Another common feature was that, with one exception, the profiles have a concave zone at the base of the hillside. This zone may include as little as about 0.1 or as much as 0.8 of the entire slope length. Also, the concave zone often consists of two different sections rather than one, as indicated by two downward-trending straight lines on the graph (for example, fig. 17*C*).

The profile pattern between the upper convex and lower concave zones varies. In five cases these two segments merged, with no intermediate zone (for example, fig. 17*C*). Two other profiles showed a straight section between the convex upper slope and the concave lower region. In the remaining five profiles the intermediate zone showed various combinations of convex, straight and concave segments, with no consistent pattern (fig. 17*A,B,D*).

The local gradient-versus-length plots verify the deduction that the top of the scar is generally on the steepest part of the hillslope. The plots also show this point in relation to the convex and concave zones. The persistent relation apparent from a study of these gradient-length graphs is that the avalanches in at least 9 of the 12 cases began at the lower end of the upper convex zone on the hillside, that is, *at the junction where the convex upper zone* *merges with the concave or straight section immediately below.* This would in fact be the steepest point on the hillside. More scar data should be analyzed to strengthen this conclusion, but the 12 scars do include different slope lengths, heights, profile shapes and degrees of erosion down their paths. In spite of these and other differences, there is a pronounced tendency for the head of the scar to be at the steepest region on the hillside.

For the same hillslopes a previous report (Williams and Guy, 1971) examined gradient-length graphs on which the origin of the profile was taken as the top of the avalanche scar. Because those graphs omitted the upper (convex) part of the hillside and began with the station several feet down from the start of the scar, the graphs indicated that the avalanches tended to form in the concave part of the hillside. It was therefore not evident that the convex upper slope of most profiles happened to terminate about at the top of the scar. Thus, more useful and accurate information is obtained by starting the profile plot at the crest of the hill.

Another pertinent point in regard to the origin of local-gradient plots is that because the abscissa is horizontal distance on a log scale, the exponent or slope of a best-fit line will vary widely depending on the origin, and consequently the magnitude, of the horizontal measurement. Although local gradient is fixed for any station on the hillside, the plotted points can be spread horizontally to cover several log cycles or compressed horizontally into a very short band, depending on the abscissa-values being plotted. White (1966) discussed other major problems associated with the sensitivity of such diagrams to choice of origin.

CROSS PROFILES

Transverse measurements of avalanche scars were made in the field; however, because the field profiles were not obtained in true straight lines across the scars and because the steep hillslopes would magnify the error due to any such deviations, the cross profiles presented here (fig. 18) were obtained from contour maps of the kind shown in figure 15. The section was marked with a straight line on the contour map and the elevation interpolated between contours at 5-foot intervals, starting at the deepest portion of the scar and proceeding to each side in turn.

The shapes and dimensions of the cross profiles vary considerably. Most scars are broad relative to the depth. Near the top of a scar, the profiles tend to be rather flat. Toward the base of a hill, the profiles have a variety of shapes—flattish, for example,

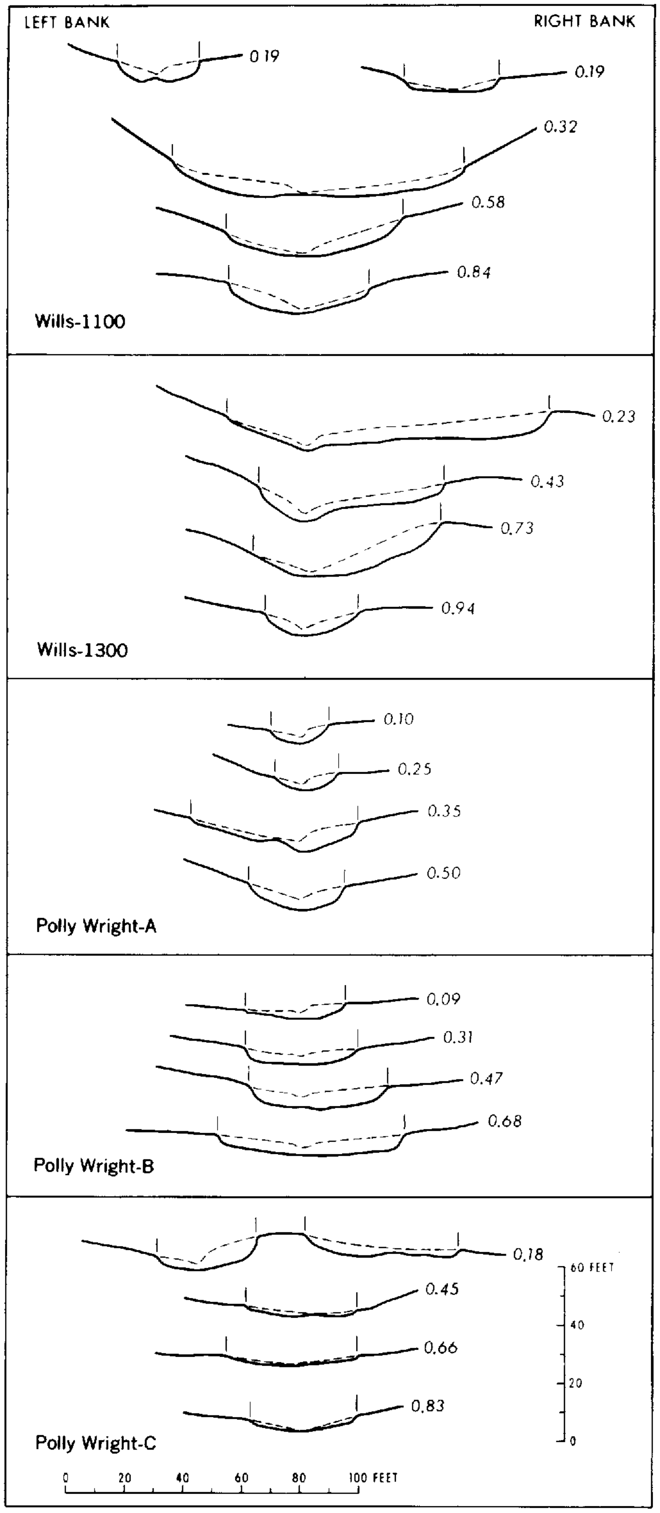


FIGURE 18.—Typical cross profiles of avalanche scars, drawn looking downslope. Number at edge of each profile indicates proportional distance of profile section down hillslope from start of scar. Vertical tick marks show edge of scar. Dashed line=estimated pre-storm ground surface.

where the soil depth was shallow and bedrock is exposed (fig. 18—Polly Wright *B*), or more U-shaped where the soil was relatively deep (fig. 18—Polly Wright *A*). Flaccus (1958) found V-shaped channels to be typical of the lower parts of avalanche scars in the White Mountains of New Hampshire.

The profiles also show typical trends of scar width, with distance downslope.

The scar depths indicated by the cross profiles in figure 18 seem to remain about constant or to increase downslope, depending on the soil thickness.

POSSIBLE CAUSES OF AVALANCHES IN NELSON COUNTY

The torrential rainfall undoubtedly was the primary cause of the debris-avalanches. But why did avalanches occur down certain strips of hillside and not on adjacent parts of the same slope or on other hillsides only 50 yards or less away? One possible reason is an uneven distribution of rainfall. A very severely eroded stream channel commonly occurred within a few hundred yards of a channel which appeared to have carried only a small flow of water. The most likely conclusion is that the rainfall intensity was highly irregular, both in time at a given spot and in area. Schwarz (1970) in fact affirmed the complex nature of the rainfall for this extreme event. He reported that a U. S. Weather Bureau survey team, investigating the various rainfall reports, found that for each of four observations of over 20 inches total rainfall, less than 6 inches was measured at a location less than 5 miles away. Huff (1967) states that multicellular patterns of precipitation cause highly variable intensities and quantities near the centers of areas covered by heavy storms.

Other factors which authors have mentioned as possible influences on avalanching are steepness of hillslope, vegetation type and density, kind of bedrock, attitude of stratified or jointed bedrock, erosion or bombardment of the base of a slope by a debris-laden stream, soil texture, orientation (aspect) of hillslope, length of hillslope, soil depth, susceptibility of soil to infiltration, ability of soil to transmit water (hydraulic conductivity), the initial presence of depressions or troughs along the hillslope, bolts of lightning, the vibrations of heavy thunder, and the uprooting of trees due to strong winds.

Although detailed measurements were not made, vegetation type and density appear to be essentially constant over the scarred area of the county, even when differences in orientation and steepness of slope and possibly in soil depth are considered.

Nearly all the bedrock of the area is rather impervious. From exposures along scars and stream channels, it appears that the avalanches were more common over massive crystalline rocks than over schists, but the former seems to be the predominant rock type anyway. There is thus no evidence that the type of bedrock influenced the likelihood of avalanching. In some cases more than one rock type is exposed along a scar. Some exposed bedrock is jointed, whereas other exposures have no joints or cleavage. Whether jointing increases the likelihood of avalanching could not be determined in our study.

The possible influence of soil texture on the ability of soil to conduct water and to be susceptible to infiltration cannot be sufficiently evaluated in a study initiated after the avalanches have occurred. Soil analyses at various stations on a hillside, before and after the avalanches, would be the best way to find any consistent associations. The same is true of the internal shear strength of any layers in the soil profile. Rapp (1963, p. 203) indicated that the main slides at Ulvådal, western Norway, occurred along planes mostly at or near the bottom of the podzolic layer and the top of the fine-grained substratum in the illuvial layer (the B horizon or layer accumulating clay for hundreds of years). Such a layer was not readily noticeable along the avalanche walls in Nelson County.

No large trees were found uprooted at the heads of scars. Some scars began immediately downslope from huge boulders, and the heads of other scars were many times wider than the zone which tree roots would disrupt. Although the role of uprooted trees in starting an avalanche could not be assessed, this factor at most was probably of minor importance.

Some big slides were heard after the thunder and lightning had stopped around 0400 August 20. Also, the lightning during the storm was mainly the horizontal type rather than vertical bolts which might have struck the ground. These considerations suggest that thunder and lightning probably were not major factors in the initiation of debris avalanches.

The remaining possibilities which could be associated with the Nelson County avalanches are stream action at the base of the slope, depressions down the hillslope, orientation of the hillside, the steepness of the slope, the horizontal length of the hillside and the soil depth. These factors are discussed in the following pages.

THEORETICAL CONSIDERATIONS

Immediately prior to an avalanche, slope stability is in a delicate balance between shear force and shear strength in the underlying materials. In soil mechanics, the ratio of these two forces is used to express a factor of safety that estimates the danger of sudden slope failure:

$$\text{Safety Factor} = \frac{\text{sum of shear forces on critical surface}}{\text{shear strength along the surface}}$$

A calculated factor of safety slightly greater than 1 in some cases may mean the possibility of slow yielding or creep, instead of a slope failure.

The mode of deformation can be planar or rotational. Except for some pockets of deep soil, the most common types in Nelson County was planar, involving a relatively thin layer of "loose" soils or products of weathering over an inclined bedrock surface. Where evidence of rotational deformation was found on deep soils, it was in some cases evident that a larger planar deformation had occurred immediately downslope.

Table 3 summarizes the various factors contributing to instability of earth slopes. Of the factors contributing to high shear stress, the weight of the rainwater is considered by the authors to be the most important in Nelson County. Bank-cutting by streams is discussed in the paragraphs below. The shear strength of the mantle probably was lowered mainly by changes in intergranular forces due to pore water.

STREAM ACTION AT THE BASE OF THE HILLSLOPE

Streams reportedly can cause avalanching by undercutting the hillside (Wenner, 1951; Scott and Gravlee, 1968; Rice and others, 1969). The mantle along the base of a slope does provide some support for the material uphill from it, so a slope would become less stable if the base of the hillside were eroded.

What is the field evidence for or against erosion of the base of the hillslope causing avalanching in Nelson County? The evidence in favor is that a stream channel trending approximately perpendicular to the direction of the avalanche had eroded its banks for as much as 20 feet along the base of some avalanche scars. However, it is possible that, instead of bearing a cause-and-effect relation, channel erosion and debris avalanches are both independent products of the torrential rainstorm.

Against erosion at the base of the mountain triggering avalanching is, first, the important fact that many debris avalanches occurred on hillsides which were not touched by stream floodwaters (fig. 14). Many avalanches, in fact, travelled overland for hundreds of feet (fig. 14) before moving into the stream channel, as for example, in the Hat Creek basin and along highway U.S. 29. Second, some of the avalanches were over a thousand feet long and only about 25–75 feet wide. It is unlikely that a few feet of erosion along the bottom of the hillside could trigger such long, narrow slides, unless the avalanching progressed upslope bit by bit. Testimony

TABLE 3.—*Factors contributing to instability of earth slopes*

[From Varnes, 1958]

Factors that contribute to high shear stress	Factors that contribute to low shear strength
A. Removal of lateral support	A. Initial state
1. Erosion—bank cutting by streams and rivers	1. Composition—inherently weak materials
2. Human agencies—cuts, canals, pits, and so forth	2. Texture—loose soils, metastable grain structures
B. Surcharge	3. Gross structure—faults, jointing, bedding planes, varving, and so forth
1. Natural agencies—weight of snow and ice and rainwater	B. Changes due to weathering and other physico-chemical reactions
2. Human agencies—fills, buildings, and so forth	1. Frost action and thermal expansion
C. Transitory earth stresses—earthquakes	2. Hydration of clay minerals
D. Regional tilting	3. Drying and cracking
E. Removal of underlying support	4. Leaching
1. Subaerial weathering—solutioning by ground water	C. Changes in intergranular forces due to pore water
2. Subterranean erosion—piping	1. Buoyancy in saturated state
3. Human agencies—mining	2. Loss in capillary tension upon saturation
F. Lateral pressures	3. Seepage pressure of percolating ground water
1. Water in vertical cracks	D. Changes in structure
2. Freezing water in cracks	1. Fissuring of preconsolidated clays due to release of lateral restraint
3. Swelling	2. Grain structure collapse upon disturbance
4. Root wedging	

presented earlier, however, indicated that the whole strip probably came down as a single event. These considerations suggest that stream erosion (or undercutting at the base of a hillside) probably was not a major cause of the avalanching, though it might locally have been a factor.

A second theory (Kuhaida, 1971) regarding stream action is that avalanching can be induced by vibrations in the bedrock as a result of bombardment of the slope-base by stream-transported debris. This process would apply more commonly to slopes on the outside of meander bends, where bombardment would be more direct. Most streams flow at right angles to the hillside, so there is some question as to the intensity, as well as the role, of any vibrations produced by the sediment moving in the stream. Also, as mentioned, many hillsides on which avalanches occurred were hundreds of feet away from a stream. Thus, while bombardment of a slope base could conceivably produce vibrations strong enough to trigger an avalanche, this probably was not a major factor in the 1969 flood.

The avalanches themselves seem to have produced the strongest vibrations during the storm. As mentioned above, people miles from the slide areas intermittently felt tremors, occurring at the same time as a loud rumble, which they later attributed to avalanches (Simpson and Simpson, 1970, p. 9). However, if vibrations caused debris avalanches, the latter probably would have occurred in groups rather than being irregularly spaced through time.

DEPRESSIONS OR TROUGHS ON THE HILLSIDE

Downslope-trending depressions or grooves on a hillside collect water, and because they become saturated sooner than the rest of the slope, they are likely places for avalanching. Field observations and photographs confirmed that many debris avalanches did indeed take place where indentations or incipient channels already existed on the hillside. In addition, cross sections of the scar and the adjacent unaffected ground, drawn from the transit surveys, commonly showed that the projected original soil surface over the scar definitely occupied a lower strip (fig. 18). Furthermore, bedrock exposed along some scars showed a weathered strip down the middle of the scar where the color and surface texture were noticeably different from the adjacent freshly exposed bedrock. This suggests that such strips had served as small incipient hillslope channels for many years before the storm. Finally, moss that was about 5 or more years old was growing on some of the exposed bedrock in the moist depression along the center of a few scars. Moss likes moisture, and moisture tends to collect in a depression or trough. (A study of the growth rate of moss on newly-exposed rock could help date previous debris avalanches.)

In general, 85 percent or more of the debris avalanches seem to have occurred along a previously existing depression in the hillside.

In their study of a neighboring Appalachian region, Hack and Goodlett (1960, p. 44) observed that "most of the chutes (scars) occupy former depressions or groove-like areas, and the impression is inescapable that the chutes are indeed incipient hollows or channelways that were partly obliterated during the passage of time by falling blocks and mass movement. They are, at rare intervals of time, flushed out and deepened by heavy runoff and the avalanching of debris." Thus Hack and Goodlett felt that the hillside depressions where avalanches tend to occur are themselves the scars of former avalanches or of similar rapid erosional processes. They further mention (p. 56) that the avalanches represent a headwater extension of the drainage network, that is, an increase in the drainage density. Figure 11 shows this feature.

Swanston (1969) found that debris avalanches in southeast Alaska in many cases occur along local drainage concentrations down the hillside.

ORIENTATION OF HILLSIDE

The aspect or orientation of the hillside could determine (1) the amount of sun the slope normally receives, which in turn influences the amount and type of vegetation, the initial (prestorm) moisture content of the soil, and consequently the depth and character of the soil profile; and (2) the angle of the attack and therefore the amount of rain on the slope, in view of the wind direction during the storm. For example, Tricart (1960) in the French Alps, Pippan (1969) in the Austrian Alps, Diseker and Richardson (1962) in Georgia (USA), and Rice, Corbett, and Bailey (1969) in southern California all found erosion or mass movement to be more common on slopes of a certain aspect. The preferred aspect, however, was not necessarily consistent among these studies. Flaccus (1958), on the other hand, concluded that the frequency of debris avalanches he studied in New Hampshire had no relation to the direction in which the hillsides faced.

In this study the amount and type of vegetation did not appear to vary significantly with different hillslope orientation. Visual examination along scar edges revealed no apparent differences in the soil profiles on slopes of different aspects. The angle of raindrop impact for a given rainfall intensity had little or no effect on sheet erosion because the ground surface was protected by an umbrella of trees, as well as by a mat of decaying leaves and low-growing vegetation. The angle of impact could, however, affect the amount of water striking the hillslope, in that more rain strikes a given area as the angle of rainfall becomes more normal to the hillside. Thus the chief potential importance of hillslope orientation probably was in regard to (a) prestorm moisture content of the ground, as influenced by the sun, and (b) the amount of water which the hillslope received during the storm, as influenced by the wind. The intensity and direction of winds blowing at ground level can vary dramatically during severe thunderstorms.

Counts were made of the number of scars occurring on hillsides facing each of the eight major compass directions (N., NW., W. and so forth). This was done by walking along selected drainage channels and marking, on a topographic map (scale 1:24,000), the location and extent of each scar. The orientation and average gradient (base to head of scar) were subsequently measured from the map. The amount of territory covered ranged from 9,000 to 16,000 feet of distance along the stream channel for each of the eight categories of hillside aspect, and many different stream channels were involved. (A given channel or reach serves two hillslope orientations, simultaneously.) The results (table 4) were expressed in terms of average number of scars per 1,000 feet of mountain stream channel (base of hillside), for each different category of approximately constant hillside (scar) gradient and aspect. A few scars for slopes less than 0.30 and more than 0.79 are not listed. Most of this work was done in the headwaters of Ginseng Hollow, Polly Wright Cove and Wills Cove. However, scars for some combinations of aspect and gradient (mainly on slopes less than 0.40 and greater than 0.70 ft per ft) were sparse, and additional observations were made in Fortunes Cove (eight scars) and the Davis Creek basin (16 scars) to supplement some of the deficient classes.

TABLE 4.—*Debris avalanche scars per 1,000 feet*

[Figures refer to average number of scars per 1,000 ft of reach inspected along stream channel or base of hillside, for a given hillside orientation A given channel distance or reach serves two orientations simultaneously. A few scars for slopes less than 0.30 and more than 0.79 are not included]

Aspect of hillside	Without respect to slope (1)	With respect to slope (ft per ft) 0.30-0.39 (2)	0.40–0.49 (3)	0.50–0.59 (4)	0.60–0.69 (5)	0.70–0.79 (6)	Total (7)
N	3.4	4.0	2.1	3.0	3.6	5.4	18.1
NE	2.8	1.6	.6	4.6	2.8	(1)	9.6
E	2.4	2.0	.7	4.9	3.4	5.2	16 2
SE	1.5	(1)	.5	2.3	.7	(1)	3.5
S	1.1	1.4	2.9	.6	.6	(1)	5.5
SW	1.3	.8	1.6	1.4	2.0	(1)	5.8
W	1.4	1.0	.3	.3	(1)	(1)	1.6
NW	1.0	0	.8	.9	1.4	1.0	4.1

[1] No avalanche scar observed for this slope and aspect.

The data in table 4, while sufficiently representative for the following analysis, may be incomplete or faulty in several ways. Although 186 scars were recorded, involving some 103,000 feet measured along mountain stream channels, the number of scars for any one aspect and slope category ranged from 0 to 13, averaging about 3. The total distance along the main stream for any one aspect and slope category ranged from 0 to 6,300 feet, averaging about 2,300 feet. In a few cases the amounts examined may not be enough to give a true picture of the avalanche frequency distribution. Also, not every combination of aspect and gradient was present to any significant extent in the study area. Furthermore, in some cases the determination of zones of constant slope and hillside aspect from the topographic map involved some subjectivity. Finally, plotting and measuring avalanche scars on the map entails some error. From the slopes of the 12 scars measured in the field, the maximum error in measuring average scar slopes from the map is judged likely to be about ±0.10 ft per ft, but usually the error is less than 0.05 ft per ft. In spite of these possible errors, the data should be reliable enough to indicate general trends.

Column 1 of table 4 shows the scar frequency for different compass orientations, without respect to the steepness of the hillside. If steepness and scar dimensions are not considered, then column 1 suggests that hillslopes facing north, northeast, and east experienced several times as many avalanches as slopes facing most other directions. Columns 2–6 give the number of scars for each slope category and aspect. Although the data are not as conclusive as might be desired, slopes facing north, northeast, and east still tend to have more scars, for approximately constant hillside gradient. Hence with most other factors virtually constant, the susceptibility of a hillside to debris avalanches in the study area tended to be associated with the aspect of the hillside. Slopes facing north, northeast, and east suffered the greatest number of avalanches, probably because the absence of direct sunshine left these slopes with a greater pre-storm moisture content and (or) the wind drove a greater amount of rainfall onto these slopes during the storm.

HILLSIDE GRADIENTS

Avalanching, soil slippage and soil erosion should occur more readily on steeper hillslopes, as reported by Rice, Corbett and Bailey (1969) and Zingg (1940). This is because the force required to start a particle moving down a slope decreases as the slope steepens.

The horizontal lines of table 4 show the frequency of avalanching with increasing gradient, for a constant compass direction. On the basis of these data alone, and contrary to what might be expected, one cannot conclude that avalanching in Nelson County occurred more often on steeper slopes. Some hillside aspects (N., SW.) show this relation, but others (SE., S., W.) definitely do not.

If it is assumed that the data in table 4 are sufficiently representative, steeper slopes, beyond a minimum of about 0.30 ft per ft (below which very few avalanches occurred), do not seem directly related to greater frequency of avalanching, for this storm. Unlikely as this seems, there may be some undetected factor or combination of factors contributing to this conclusion. One possibility is that the gradient at the head of the scar may be the pertinent slope, rather than the average gradient as measured from head to base of scar. Another possibility is that the steeper slopes, say greater than about 0.70 ft per ft, do not occupy a large enough distance along the stream channel or a large enough hillside area to give a statistically meaningful frequency of avalanching.

LENGTH OF HILLSLOPE

Longer hillslopes understandably can provide greater quantities of sediment than short slopes (Zingg, 1940). In addition, longer hillsides may enhance the likelihood of avalanching because the downhill movement of water produces more water on and in the soil with distance downslope. Data on observed scars and topographic map measurements of hillslope length were used to determine if debris avalanching was more common and more extensive on longer hillslopes.

The fieldwork consisted of marking on the topographic map the locations of avalanche scars and estimating their lengths by eye. Only scars having north, northeast, and east aspects were considered because these aspects suffered more avalanching than others. Gradient was ignored, on the basis of the tentative finding that avalanching occurred with about equal frequency for all slope categories between 0.30 and 0.80 ft per ft.

Topographic-map analysis consisted of measuring the horizontal length of the hillslope, from ridge crest to stream channel, at each scar location, and measuring the distance along the stream channel, for each category of hillslope length. Columns 1 and 3 of table 5 show the categories of hillside length and the total channel distance inspected for each category. The distances in column 3 are converted to percentages in column 4. (Hillslope lengths less than 200 ft represented only 2 percent of the drainage area and had essentially no scars.)

In table 5 the number of observed scars for each category of hillslope length (col. 7) does not truly indicate any influence of slope length because all slope lengths were not equally present. For example, 28 avalanches occurred on hillsides 400–500 feet long, while only seven scars were found on hillsides longer than 1,000 feet; however, slopes 1,000 feet long were scarce, relative to those in the 400–599-foot class. The number and (or) length of observed slides in each category of hillslope length therefore should be compared to the percentage of valley distance along the mountain stream channel (col. 4) and also to the relative amount of drainage area (col. 5) for each category. Because we inspected nearly all channels in which the sideslopes had north, northeast, and east aspects, column 4 in table 5 is

TABLE 5.—*Effect of hillslope length and drainage area on frequency and amount of scarring*

(1) Category of hillslope length (ft)	(2) Midpoint-average length of hillside (ft)	(3) Valley distance inspected (ft)	(4) Percent of total valley distance	(5) Area of sample = (2) (3) / 43,560 (acres)	(6) Percent of total sample area represented by each category of hillslope length	(7) Number of scars observed	(8) Total length of all scars[1] (ft)	(9) Relative number of scars per acre = (7)/(5)	(10) Average length of scars per acre = (8)/(5) (ft)
200–399	300	27,800	28	192	14.0	20	3,540	0.104	18.4
400–599	500	29,900	30	344	25.1	28	8,560	.081	24.9
600–799	700	23,700	23	381	27.7	19	7,850	.050	20.6
800–999	900	12,400	12	256	18.7	14	7,430	.055	29.0
≧1,000	1200	7,200	7	199	14.5	7	4,960	.035	24.9
Total		101,000	100	1,372	100.0	88	32,340		

[1] Length of individual scars estimated during field inspections and summed for each hillslope category.

representative of the entire drainage area of the basins studied. Drainage area, as used in table 5, is the product of the mean hillslope length (col. 2) and the valley distance for the particular category (col. 3), in units of acres. The percentage of the total area represented by each category (col. 6) was then determined from column 5.

The number of observed scars (col. 7) is roughly proportional to the percentage distance along the mountain stream channel (col. 4), for a given category. Therefore the number of scars is essentially independent of the length of hillside, that is, the length of the hillside probably had little or no influence on the likelihood of avalanching.

In regard to the effect of drainage area on the frequency and amount (length) of scars, column 9 in table 5 lists the relative number of scars per acre and column 10 gives the average total length of scars per acre. Figure 19 shows how these factors vary with hillslope length. The relationships in figure 19 can be reduced to a straight line by using the square root of the hillslope length, but this would not improve their accuracy. (See Wischmeier and Smith, 1965, p. 9.) While hillslope length is not an important factor in the number of slides reaching a unit length of stream valley, it is an important factor in the number per unit of drainage area and in the total length of scars per unit drainage area. As length of hillslope increased, the number of scars per acre decreased (fig. 19*A*) and the total length of scars per acre increased (fig. 19*B*). For example, about 10 scars occurred per 100 acres on 300-foot slopes in contrast to only about three avalanches per 100 acres on 1,200-foot slopes (fig. 19*A*). On the other hand, the total length of scarring was about 1,600 feet per 100 acres on 300-foot slopes and 2,800 feet per 100 acres on 1,200-foot slopes.

These general conclusions regarding slope length can be more readily visualized in a drawing of two hypothetical sections of hillslope (fig. 20). Both sections have the same drainage area, but section *B* is 2.5 times the slope length of *A*. The unit distance along the stream channel in figure 20 is distance *XY*. The number of avalanche scars encountered in walking along this unit reach is the same for both sections (five for this example), so that hillslope length is not a factor in the frequency of avalanching per unit distance along the stream channel. However, on the basis of unit drainage area, section *A* has 12 scars whereas *B* has only five scars. This shows how more avalanches occurred on the short than on long slopes, for a given drainage area (col. 9 of table 5 and fig. 19*A*). Lastly, summing separately the scar

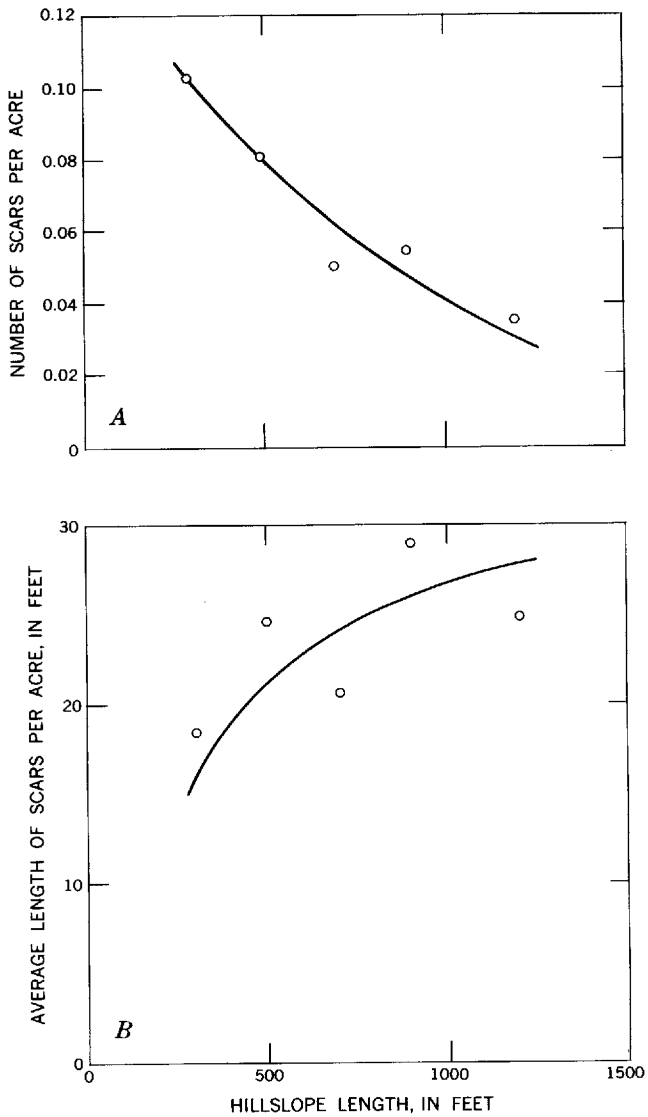


FIGURE 19.—Relation of hillslope length to number of scars per acre and to average distance of scarring per acre, for hillsides having north, northeast, and east aspects.

lengths in sections *A* and *B* shows that the scars in *B* amount to a greater total length. Since the acreage of *A* and *B* are equal, the average length or distance of avalanching per acre is greater on long than on short slopes (col. 10 of table 5; fig. 19*B*). The latter conclusion actually holds for unit distance along the stream channel, as well as for unit drainage area.

SOIL DEPTH

The influence of soil depth on avalanching could not be determined. Most of the avalanches began in soil depths of about three feet or less, although minor soil slips were common near the bottoms of hillslopes

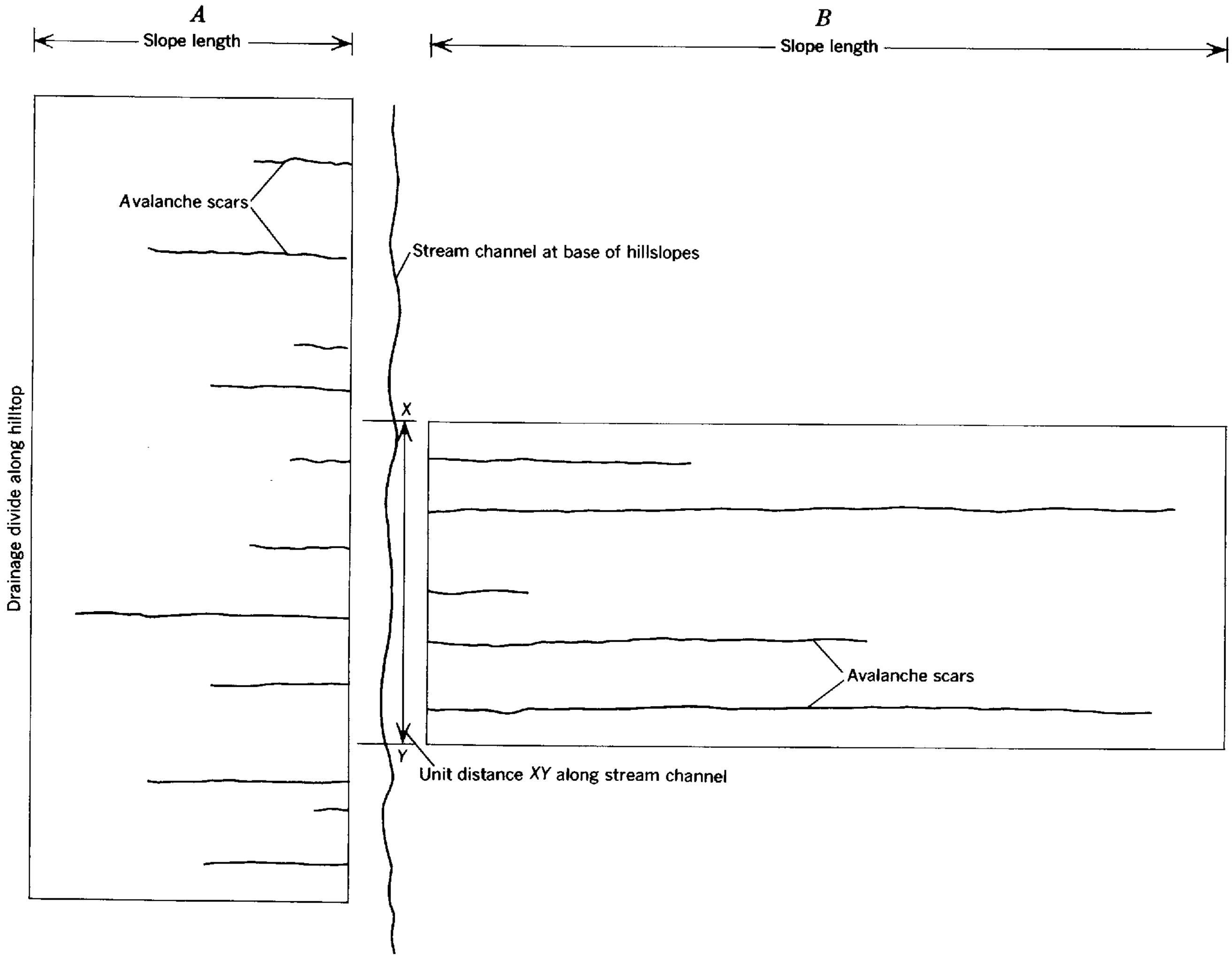


FIGURE 20.—Top (plan) view of two hypothetical drainage areas, *A* and *B*, showing influence of hillslope length and drainage area on frequency and length of avalanching. Areas of *A* and *B* are equal, but the hillslope length of *B* is 2.5 times that of *A*. Though the length of scarring is greater on *B*, the number of scars is greater on *A*.

where the soil depth was unknown and where the slip may have been triggered by undercutting.

Soil depth may affect avalanching, in that water can filter downward for greater distances in deep soil, whereas a shallow soil would quickly become saturated. The potential for soil slippage increases with increasing saturation, which causes a greater bulk weight. Thus from this viewpoint, one might expect more avalanching in shallower soils, other factors being equal.

On the other hand, the shear force would be proportional to the depth of the soil and water. Deeper soils, which also could contain more water, therefore would be subjected to a greater shear force. There would also be more possibilities for planes of weak shear strength with deep soils. These considerations suggest that avalanching might occur more readily on deep soils, if other factors are equal.

If the effect of greater shear over deeper soils is important, deep-soil avalanches might tend to occur on flatter slopes than those in which shallow-soil avalanches occur. Such a depth effect could explain why we found as many avalanches on relatively flat slopes (0.30–0.39 ft/ft) as on steep slopes (0.70–0.79 ft/ft).

SUMMARY

The head of a debris avalanche tended to be at the steepest point on the hillside, that is, at the junction between the upper convex slope and the straight or concave section. The only factors associated with greater frequency of debris avalanching for this storm were depressions on the hillside, aspect of hillside, and length of hillside per unit drainage area. More avalanches occurred on slopes facing north, northeast, and east than on slopes facing other directions. Some factors, such as soil depth and various soil properties, may affect the likelihood of avalanching, but their importance could not be assessed. Other variables (for example, average steepness of hillslope within the range of 30–80 percent gradient) did not appear to be related to the occurrence of an avalanche. Hillslope length did not affect the number of slides reaching a unit length of stream channel; however, on a unit-drainage-area basis more scars occurred on short than on long slopes, whereas the length of scarring was less on short slopes than on long hillslopes. All the above conclusions depend on the uncertain assumption that the rainfall intensity, although probably varying in time and area, did not affect any of the possibly relevant factors.

EFFECT OF DEBRIS AVALANCHES ON STREAMFLOW

The sediment deposited by a debris avalanche or landslide can temporarily block or severely impede the flow along a stream channel. Famous examples of this phenomenon are the 1925 slide on the Gros Ventre River in Wyoming and the 1959 landslide into the canyon of the Madison River near Yellowstone National Park. In Nelson County channel blockage probably occurred just upstream from a sharp bend in the headwaters of Ginseng Hollow (Guy, 1971). The evidence for channel blockage here is the very big difference between the maximum water discharge attributable to precipitation rate alone and the much higher discharge which actually flowed through the bend as measured by indirect methods. The large discharge through the bend partly reflects the release of a volume of water stored upstream of the avalanche as it was sliding across the channel and partly reflects the volume of the slide material that was added to the stream discharge.

An estimate of the channel discharge which would result from the maximum rate of rainfall was made in the same manner described earlier in the report, using the rational formula. With 25 inches per hour as the maximum probable rainfall intensity, the measured drainage area as 20 acres, and a coefficient of 0.90, the maximum discharge in the channel attributable to rainfall would be about 450 cfs.

An actual peak flow of about 8,000 cfs in the channel was determined from topographic field measurement at section B–B' (see fig. 21). Because these measurements were obtained by hand level, stadia rod and pacing, the results are not precise and are only approximations. The required measurements at section B–B' were (1) the superelevation Δh, that is, the vertical distance between the high-water marks on the inside and outside of the bend, (2) the cross-

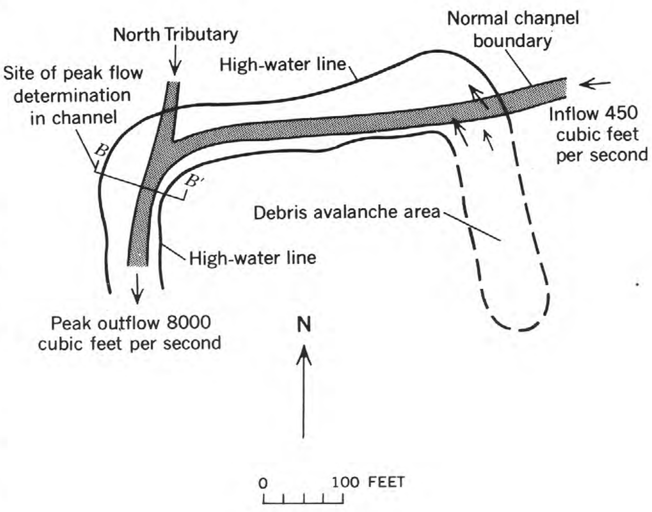


FIGURE 21.—Sketch of debris avalanche zone, normal channel boundaries and high-water lines left by peak flow, northeast branch of Ginseng Hollow.

sectional flow area A, (3) the outside and inside radii of the bend (r_o and r_i, respectively), (4) the average radius of curvature $r_c = (r_o + r_i)/2$ and (5) the water-surface width W.

The approximate mean flow velocity V in the bend can be computed from an equation transformed from Chow (1959, p. 448):

$$V = \left(\frac{\Delta h g r_c}{W} \right)^{1/2}$$

where g=acceleration due to gravity. Inserting the measured values gives

$$V = \left(\frac{13.5 \times 32.2 \times 80}{80} \right)^{1/2} = 21 \text{ ft per sec.}$$

A different method of computing V is through the use of Grashof's theory as given by Schoklitsch (1937, p. 151):

$$\Delta h = 2.30 \frac{V^2}{g} \log \frac{r_o}{r_i}$$

Rearranging the equation and inserting the field data yields

$$V = \left(\frac{13.5 \times 32.2}{2.30 \times \log \dfrac{120}{40}} \right)^{1/2} = 20 \text{ ft per sec}$$

which differs by less than 5 percent from Chow's more approximate equation. The computed peak discharge Q is then $Q = AV \approx (390)(20) \approx 8{,}000$ cfs.

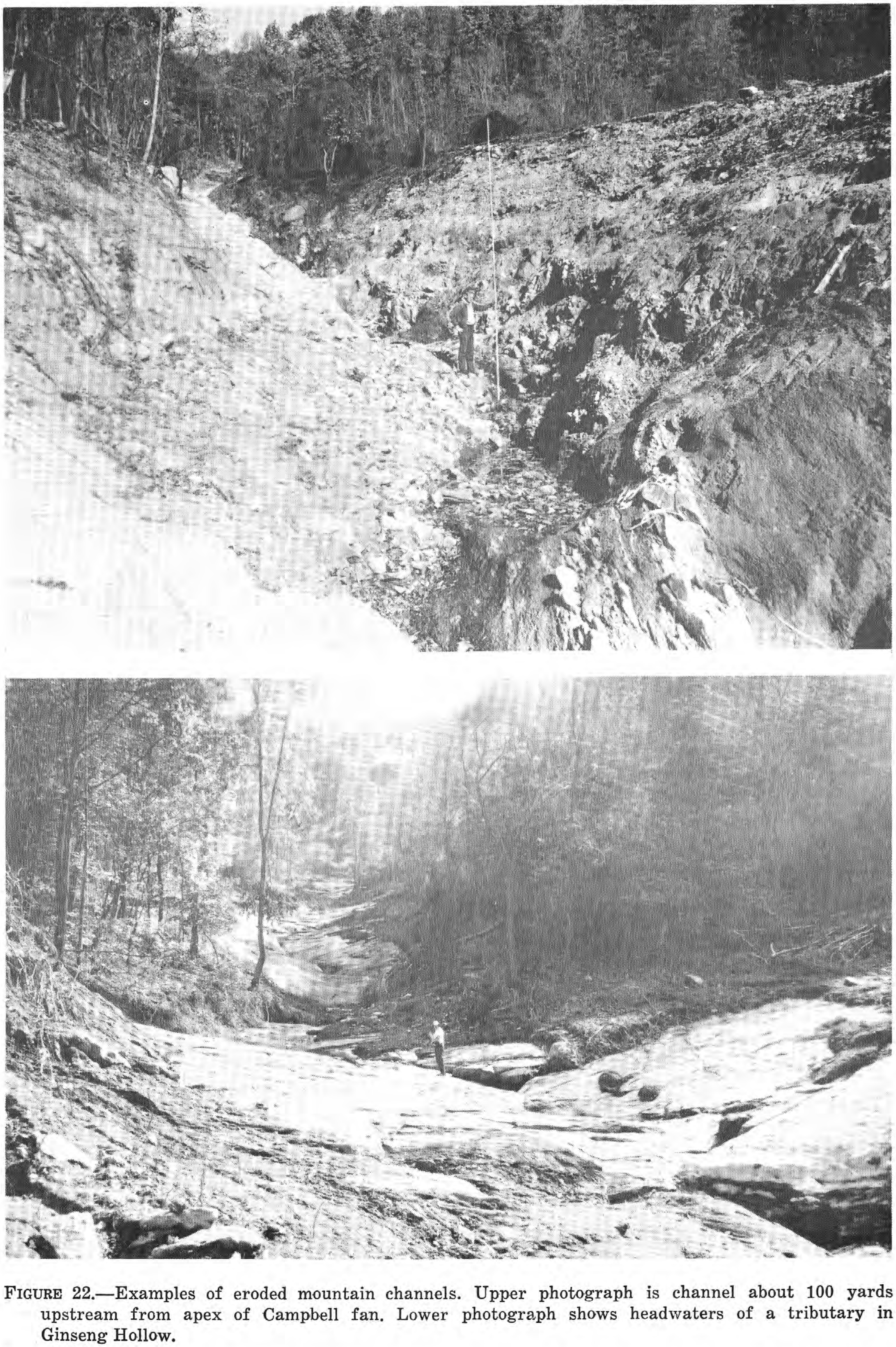


FIGURE 22.—Examples of eroded mountain channels. Upper photograph is channel about 100 yards upstream from apex of Campbell fan. Lower photograph shows headwaters of a tributary in Ginseng Hollow.

An error analysis of this calculated 8,000 cfs should consider the effects of the measuring techniques used, nonsteady flow, irregular velocity distributions, irregularly shaped channel boundaries, imperfect circular shape of the bend, the effect of sediment and other debris in the flow, and the possibility that some channel enlargement may have occurred after the peak flow. Also, a tributary from the north entered the channel just upstream from the bend (fig. 19); however, near the junction the high-water marks on this north tributary were much lower than those on the study reach; so the flow from the north tributary probably was stored behind the rushing water-sediment mixture and did not contribute to the peak flow in the bend. The many indeterminates, plus the imprecise basic data measurements, could amount to an error of about ±3,000 cfs in the computed maximum discharge. Nevertheless, the peak discharge definitely was many times (about 18 times, as estimated here) the discharge reasonably attributable to an assumed very high rate of rainfall runoff. The most likely explanation for the high flow rate is that the 57,000 cubic-foot avalanche moved through the reach in only a few seconds, together with water which the slide had temporarily blocked in the stream channel.

This case exemplifies one way in which a debris avalanche could cause serious damage to man and structures, especially in a small drainage area. By temporarily disrupting normal runoff, a hillside avalanche in the headwaters can trigger a chain of events culminating in a tremendous surge of sediment-laden water traveling at a very high velocity. The large difference in discharge between a relatively steady flood flow, though it may be at record heights, and a big surge of water and debris can mean the difference between minor damage and devastation.

STREAM CHANNELS

Mountain streams in Nelson County commonly range from 1 to 10 feet wide. During the flood the stream channels were enlarged by severe erosion and the removal of trees and shrubs (fig. 22). The exact extent of such erosion could not be precisely determined, as the dimensions and location of the channel before the flood in most reaches were uncertain. However, estimates and in some cases measurements were made of the amount of channel erosion for headwater streams, as discussed under "Sediment Yield."

Figure 23 shows a channel which formerly was only a few inches deep and narrow enough for a man to step over. The magnitude of stream erosion in this photograph was not uncommon in the mountainous parts of the study area.

In some upstream reaches of mountain channels the bedrock on the side slopes restricted the bank erosion. In other areas bank erosion was evident where vegetation such as trees and crops still clung by a few tenacious roots to alluvium along the edges of stream banks. Many stretches of old logging roads in the mountains and of paved highways in the downstream valleys were partly or completely washed away (fig. 24). Farmers along Indian Creek and near the mountain ravine upstream from the Campbell alluvial fan testified that the new channels are at least twice as wide as the preflood channels.

Two factors restricted the downcutting of stream beds. On the steeper mountain channels the resistant bedrock was exposed either prior to the flood or soon after the flood started, due to the thin mantle. Downstream in the broader valleys the overall gradient was much flatter, and this limited the amount of downcutting which the stream could accomplish while still maintaining a slope steep enough to satisfy the hydraulic and sediment-transport requirements.

Farther downstream, severe scour commonly oc-

FIGURE 23.—Channel carved by the flood in the yard of S. K. Wills, at the head of Wills Cove. Before the flood this channel was but a few inches deep and so narrow that a man could easily step over it.

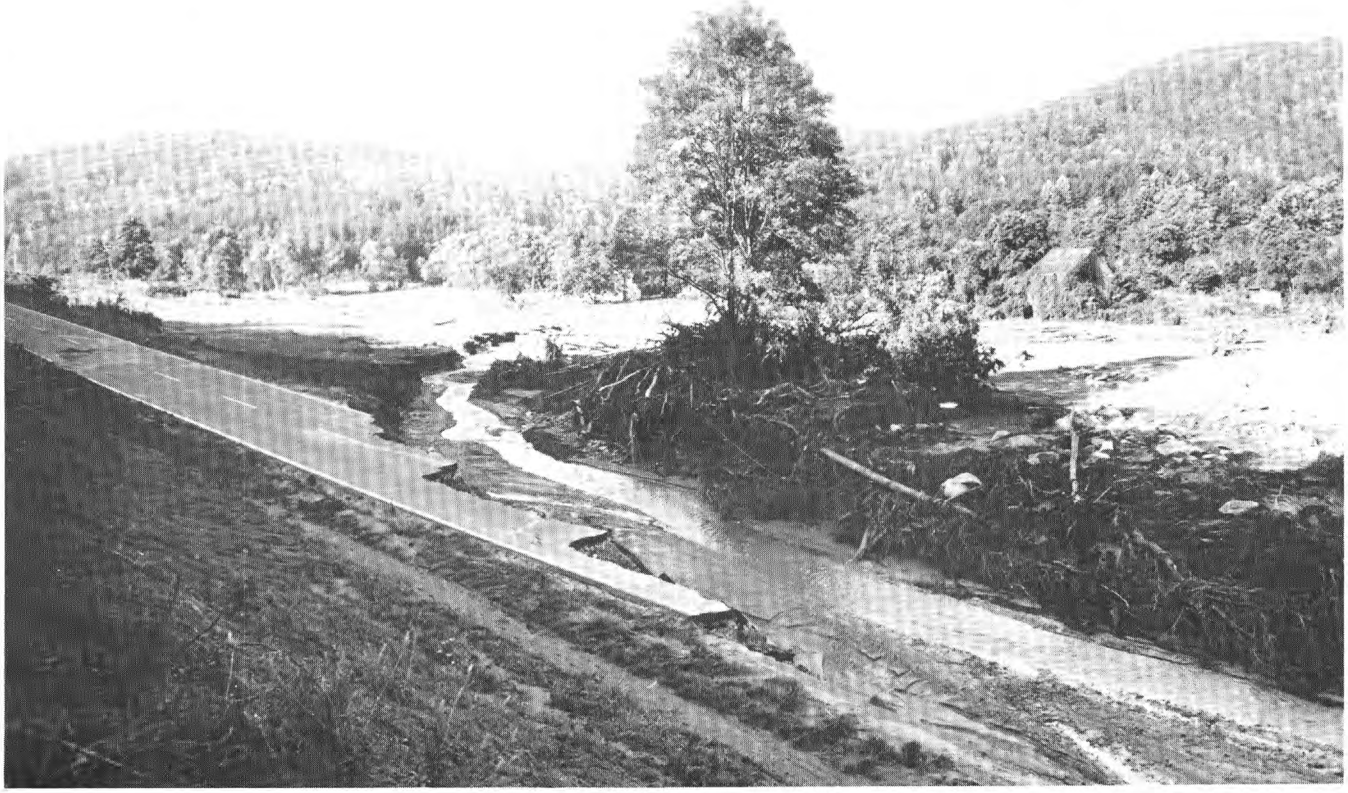


FIGURE 24.—Channel erosion along highways. (Upper photograph courtesy of Virginia Division of Mineral Resources; lower photograph courtesy of Virginia Department of Highways.)

FIGURE 25.—Scour of stream banks under a railroad trestle—Tye River at Norwood, Va. James River in background. Bridge completely washed out in foreground. Note also floodplain deposition. (Photograph courtesy of Virginia Department of Highways.)

curred next to bridges and railway trestles (fig. 25). In some localities as much as 20 feet of the original stream bank was eroded, on each side of the stream. The many trees carried by the raging streams probably contributed to this. The trees caught on bridge piers and collected into huge debris jams that blocked parts of the normal stream channel. Blockage of parts of the flow cross section could cause faster velocities in the unobstructed sections, thus increasing the stream's erosive power. The jams also encouraged overbank flooding which in some cases resulted in erosion of the surface of the floodplain.

W. N. Whitehead (Simpson and Simpson, 1970, p. 150) happened to observe the process by which a roadway was eroded next to a bridge on Route 151 at Tye River.

"The river was up and passing over the roadway. As it cascaded down the east (downriver) side of the fill, it began to erode this, and we watched it as it kept digging back, with the road collapsing bit by bit. The water ate its way back, from the downriver side, all the way through the fill. Once it got back thin enough the whole thing broke, and the water just charged through. Then the gap between the bridge and the edge of the road started getting wider and wider. This was occurring on each side of the bridge."

After the water broke through, the gaps accommodated some of the flow and the level dropped. At its peak, the water had been "well over the bridge railings."

Another type of channel erosion in some downstream areas was the caving or slumping of freshly deposited sediment into the stream, along the banks (fig. 25). This erosion probably occurred as the flood was receding and involved sediment laid down earlier by the same flood. The amount of sediment eroded by bank caving is difficult to estimate.

Channel erosion, then, was certainly quite severe in some places; however, other channels apparently underwent little or no erosion. Thus the amount of channel erosion varied widely, even within areas as small as an acre.

Bank erosion frequently exposed old alluvial deposits. Some of these are 30 feet thick or more, with particles ranging from sand to boulders in size. Stratification is sometimes evident, but in other cases all particles are completely mixed in a vertical section. Occasionally some lenses of imbricated particles appear. These alluvial deposits probably resulted from ancient floods.

Along any given reach, a stream could exhibit deposition as well as erosion (fig. 26), but on balance more erosion occurred in the headlands, whereas deposition predominated in the flatter and wider downstream valleys. Both erosion and deposition occurred, in about equal amounts, in the apex region of alluvial fans.

SEDIMENT YIELD

VOLUME OF AVALANCHE EROSION

The volume of sediment removed by a debris avalanche could not be measured precisely, and all

A

B

C

FIGURE 26.—Channels showing both erosion and deposition. *A*, South Fork of Rockfish River. *B*, Ginseng Hollow. Note men for scale. *C*, Small tributary to Hat Creek.

of the data presented here are based on field estimates of the prestorm ground configurations.

Table 2 lists the estimated erosion volumes for the 12 scars measured in detail. The amount of sediment removed from these typical scars ranged from about 18,000 cubic feet to 233,000 cubic feet. The average of the 12 is 88,000 cubic feet. The 12 examples selected do not include the largest and smallest scars in the entire storm area.

Figure 27 shows both the amount of erosion and the compass orientation for these 12 scars. If the orientations and volumes of the scars shown in figure 27 are representative of the entire drainage areas, the diagram strengthens the earlier conclusion that hillsides facing north through east were most prone to sliding, not only in frequency of avalanches but also in total amount eroded.

Field inspection was made of all of the upper part of the Wills Cove drainage area (shown in figure 4), all of Ginseng Hollow, and about 86 percent of the Polly Wright drainage area. Visual estimates were made of the length, average width and average depth of each scar encountered. Column 2 of table 6 lists the total estimated amounts of avalanche erosion for the various tributaries and subbasins. The avalanches in Wills Cove produced about 2.60 million cubic feet, those in Ginseng Hollow about 1.44 million cubic feet, and those in Polly Wright Cove about 1.46 million cubic feet of sediment. The neighboring Davis Creek, Fortunes Cove and adjacent areas were also severely affected by avalanches; so the total amount of sediment eroded by this process was considerable.

Column 5 of the table shows that the volume of avalanche erosion per square mile ranged from about 1.5 to 2.1 million cubic feet, with the greatest amounts originating in Ginseng Hollow.

A very crude relation exists (fig. 28) between the average slope of the ground surface, as determined from the the topographic maps (table 1), and the volume of sediment eroded in debris avalanches. The scatter prevents the determination of a reliable regression curve, but the trend is from no erosion at 32–35 percent average slope to 3 or 4 million cubic feet per square mile at 50 percent average slope.

SEDIMENT YIELD FROM STREAM CHANNELS

As a part of the detailed studies made in Wills Cove, Ginseng Hollow and Polly Wright Cove, estimates were made of the net volume of channel erosion in many of the tributaries and subbasins. These estimates were based on visual field inspection of the complete lengths of most tributaries and on the results of 18 surveyed channel cross sections which compared prestorm and poststorm profiles. About $2\frac{1}{2}$ years before the storm the U.S. Geological Survey had taken aerial photographs of the entire area for topographic map purposes. Using a Kelsh stereoscopic plotting instrument with these photographs, we measured on the photographs 18 prestorm channel cross sections at locations which, judging from the aerial photographs, would be easiest to locate in the field. We then went to the field with a transit and level rod and measured the post-storm cross section at these 18 sites.

Although the poststorm profiles could be measured quite accurately, the final erosion estimates include several possible errors. The 1967 photographs may not exactly represent the condition of the channels just before the 1969 flood, but any channel erosion

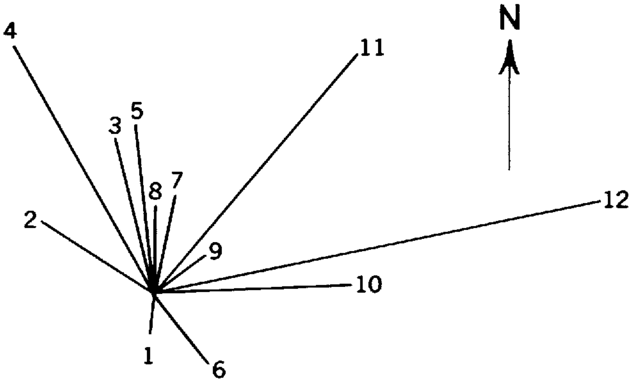


FIGURE 27.—Orientation and relative volume of the 12 avalanches that were measured in detail. Scale: 1 inch=100,000 cubic feet.

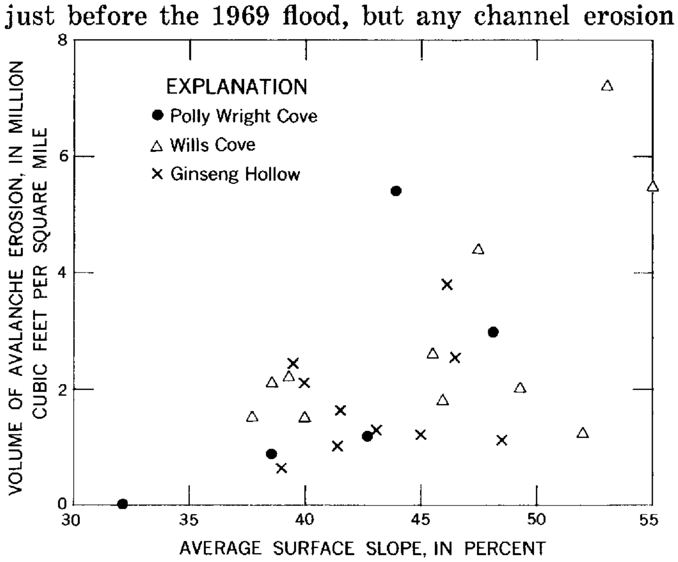


FIGURE 28.—Volume of avalanche erosion with respect to average surface slope.

TABLE 6.—*Estimated net sediment yields and storm average transport rates*

[Estimates are based on field inspection of all of the Wills Cove (upper part) and Ginseng Hollow drainage areas and about 85 percent of the Polly Wright Cove basin. See figure 4 for designation of drainage basins indicated in column 1]

Drainage basin	Net yield (10^3 ft^3)			Net yield per square mile (10^6 ft^3)			Average denudation (inches)	Weight of sediment eroded (10^4 tons)	Storm-average sediment-transport rate	
	Avalanche	Channel	Total	Avalanche	Channel	Total			(10^4 tons per day)	(10^3 lb per sec)
(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)
Wills Cove, upper part										
West tributary	908	873	1,781	3 8	3.7	7.5	3.2	104	357	83
Middle tributary	379	403	782	2.6	2.8	5.4	2.3	46	158	37
Falls tributary:										
Subbasins 1	316	158	474	7.9	3.9	11.8	5.1	27	93	22
2	52	23	75	1.4	.6	2.0	.9	4	14	3.2
3	2	4	6	.07	.14	.2	.1	.3	1	0.2
4	1	75	76	.01	1.1	1.1	.5	5	17	4.0
5	0	68	68	0	1.5	1.5	.6	4	14	3.2
6	28	56	84	.9	1.9	2.8	1.2	5	17	4.0
7	1	−1	0	.04	−.04	0	0	0	0	0
8	136	106	242	2.6	2.0	4.6	2.0	14	47	11.1
9	65	108	173	1.4	2.3	3.7	1.6	10	34	7.9
10	32	48	81	1.0	1.5	2.5	1.1	4	14	3.2
11	32	131	163	1.2	4.9	6.1	2.6	10	34	7.9
12	50	86	136	2.1	3.6	5.7	2.5	8	27	6 3
13	65	20	85	.6	.18	.8	.3	5	17	4.0
14	98	120	218	1.6	2.0	3.6	1.5	13	45	10.3
15	98	9	107	3.1	.3	3.4	1.5	6	21	4.8
16	60	81	141	1.0	1.3	2.3	1.0	8	27	6.4
17	30	20	50	.5	.3	.8	.3	3	10	2.4
18	120	0	120	8.6	0	8.6	3.7	7	24	5.6
19	45	96	141	2.0	4.4	6 4	2.8	8	27	6.4
20	4	7	11	.7	1.2	1.9	.8	5	17	4.0
21	0	24	24	0	3.0	3.0	1.3	2	7	1.6
22	81	42	123	2.1	1.1	3.2	1.7	7	24	5.6
Subtotal (1–22)	1,316	1,282	2,598	---	---	---	--	155	531	124.1
Noncontributing	0	0	0	---	---	---	--	0	---	----
Main channel	----	965	965	---	---	---	--	60	206	48
Total, Falls tributary	1,316	2,247	3,563					215	737	172.1
Average, Falls tributary	----	----	----	1.1	1.9	3.0	1.3			
Total, Wills Cove	2,603	3,523	6,126					365		
Average, Wills Cove	----	----	----	1.7	2.2	3.9	1.4			
Ginseng Hollow										
Lower subbasin	523	712	1,235	3.0	4.1	7.1	3.1	73	250	58
North tributary	104	152	256	1 2	1.8	3.0	1.3	15	51	12
South tributary	616	526	1,142	5.4	5.6	11.0	4.7	67	230	53
Main above sta. 3400	197	303	500	.9	1.4	2.3	1.0	30	103	24
East bowl (noncontributing)	0	0	0	0	0	0	0	0	0	0
Total, Ginseng Hollow	1,440	1,693	3,133	---	---	---	--	185	634	147
Average, Ginseng Hollow	----	----	----	2.1	2.5	4.6	2.0			
Polly Wright Cove										
Subbasins 1	85	114	199	2.0	2.7	4.7	2.0	12	41	9
2–5	180	194	374	2.1	2.3	4.4	1.9	22	75	17
6	80	211	291	2.2	5.7	7.8	3.4	18	62	14
7	40	90	130	1.5	3.5	5.0	2.2	8	27	6
8	133	268	401	1.9	3.9	5.8	2.5	24	82	19
9	156	230	787	2.6	3.8	6.4	2.8	23	79	18
10	248	86	334	7.2	2.5	9.7	4.2	19	65	15
11	204	105	309	5.5	2.8	8.3	3.6	18	62	14
12	187	128	315	4.4	3.0	7.4	3.2	18	62	14
13	28	92	120	1.2	3.8	5.0	2.2	7	24	6
Total, subbasins 1–13	1,341	1,518	2,859	---	---	---	--	169	---	----
Lower	115	123	238	1.5	1.6	3.1	.9	14	48	11
Noncontributing	0	0	0	0	0	0	0	0	0	0
Total, Polly Wright Cove	1,456	1,641	3,097	---	---	---	--	183		
Average, Polly Wright Cove	----	----	----	1.5	1.7	3 2	1.4			

during this 2½-year period probably was minor. The plotting-instrument technique, probably one of the smallest sources of error, was at best accurate to within about ±1 foot in elevation. Horizontal dis tances, measured from the plotted Kelsh elevation points with an engineer's scale, were accurate to within 1 percent. The major problem was locating in the field the specific site that had been measured on the photographs. (The photographic cross sections had to be measured first because of the limited choice of suitable locations identifiable from the photographs.) Numerous abandoned logging roads along the ravines helped pinpoint the selected sites, but several of the sections were difficult to locate in the field because some of the roads, old cabins, and other landmarks noted on the photographs were washed away during the flood. The various sources of error could amount to about 25 to 50 percent in the amount of erosion attributed to any site. In addition, the visual estimates, the method used to measure poststorm profiles in most reaches, involve an unknown amount of error.

While observations were being made of channel erosion, attention was also given to any significant channel deposition. Such deposition was notable and persistent along some parts of the main channel in the Falls tributary of Wills Cove and in Ginseng Hollow. Deposition occurred mainly in reaches of flatter gradient and less confined valley walls where stream velocity probably decreased. The figures for net sediment yield from channels (cols. 3 and 6, table 6) represent the estimated erosion minus any deposition.

Channel erosion, as near as can be determined from these volume estimates, accounted for more than half of the total sediment erosion (col. 3, table 6). The amounts of yield from individual tributaries were as much as 5.7 million cubic feet per square mile (col. 6, table 6). The average yield for the three basins was 2.2, 2.5 and 1.7 million cubic feet per square mile for Wills Cove, Ginseng Hollow and Polly Wright Cove, respectively. The lower average for Polly Wright reflects the effect of a rather large part of the basin on which little or no erosion occurred.

Net channel erosion apparently was slightly greater than avalanche erosion, as far as volume of erosion per square mile is concerned (cols. 5 and 6 of table 6).

TOTAL SEDIMENT YIELD AND AVERAGE DENUDATION

If the amounts of yield from the separate subbasins or tributaries are added, the estimated total yield from the headwaters of Wills Cove is 6.1 million cu ft, from Ginseng Hollow about 3.1 million cu ft, and from Polly Wright Cove about 3.1 million cu ft. These volumes represent 3.9, 4.6 and 3.3 million cu ft per sq mi for the Wills (upper part), Ginseng and Polly Wright basins, respectively.

If each of these amounts were imagined to be removed uniformly from the entire area of the respective drainage basin, the denudation from Wills Cove averaged about 1.4 inches, from Ginseng Hollow about 2.0 inches, and from Polly Wright Cove about 1.4 inches. The average denudation for an individual tributary or subbasin (col. 8 of table 6) ranged from little or nothing to about 5.1 inches.

Judson and Ritter (1964) give estimates indicating that for central Virginia the average denudation due to sediment removal by suspended load and bedload is in the range of about 0.6–1.4 inches *per thousand years*. The 1.4–2.0 inches removed in one storm for the drainage basins studied here show an extent to which individual basins and rare events can depart from the long-term average.

ESTIMATED SEDIMENT-TRANSPORT RATES

Although no measurements of sediment-transport rate were made during the flood, a very rough idea of the likely average rate at the mouths of some tributaries can be obtained from the estimated total sediment yield and the duration of the storm. As transport rate customarily is expressed in weight per unit time, the estimated volumes of sediment yield must be transformed into weights. This transformation (col. 9 of table 6) was made by using 110 lb per cu ft (pounds per cubic foot) for the avalanche soils and 125 lb per cu ft for the coarser channel material. Seven hours was the approximate time during which the sediment moved in these headwater areas. The storm-average sediment transport rates (cols. 10 and 11 of table 6) ranged from very little at the mouths of some subbasins to as much as about 7,370,000 tons per day (172,000 lb per sec) at the mouth of the Falls tributary in Wills Cove. These estimated rates are for the full stream width. Since the stream at the mouth of the Falls tributary was about 100 feet wide during the storm, the estimated rate of sediment movement at that section may have averaged about 1,700 lb per sec per ft of width for the 7-hour period. (By way of comparison, laboratory flumes rarely exceed 1 or 2 lb per sec per ft width.) Such magnitudes reflect the large quantities of rock and earth unleashed by debris-avalanches and severe channel erosion. Because of this irregular rate of sediment introduction, the transport rate at any one place must have varied considerably with time.

DEPOSITIONAL FEATURES

The floodwaters laid down extensive deposits of sediment, often to the detriment of man and his property. In some places the sediment came to rest very close to its source, as at the foot of a hillslope or on an alluvial fan at the mouth of a mountain channel. At the other extreme, some of the small particles eroded from the mountainsides undoubtedly moved all the way to Chesapeake Bay or the Atlantic Ocean.

Five kinds of deposits were left after the flood: (1) debris avalanche deposits at the base of hillslopes; (2) mountain-channel deposits, usually in sediment patches scattered along the newly eroded channel, but occasionally as large debris piles behind a temporary dam; (3) alluvial fans where a narrow mountain channel emerged into a relatively broad intermontane valley; (4) deltas at the junction of a stream and major highway, where backwater formed during the flood due to plugging of a culvert; and (5) vertical accretion deposits on floodplains.

Only rarely were debris avalanche deposits left at the base of the hillside. Some of these particles were probably too big to be moved by the floodwaters (fig. 29). In other cases the slide probably did not occur until after the peak flow, when the flow in the ravine was no longer sufficient to carry away the debris. Where no stream channel flowed past the hillside, the momentum of the avalanche nearly always transported most of the material away from the base of the slope.

FIGURE 29.—Boulders which were dislodged in a debris avalanche and which probably were too large to be moved by the floodwaters (Ginseng Hollow). Man on rock shows scale.

CHANNEL DEPOSITS

Some eroded material was left here and there on the bed of the mountain channel after traveling a distance ranging from a few feet to nearly a mile. These channel deposits were either laid down by the recession flow (since the peak flow probably could have moved the particle sizes involved) or trapped behind dams such as huge rocks or log jams (fig. 30), in which case the deposition could have occurred at various stages of the flood.

Deposits are sparse along the enlarged mountain channels and generally absent near the heads of the streams. As a stream approached its outlet into the broad valley, the scattered deposits on some channels—those which became wider and flatter—became more common, thicker (as much as 5 ft), and locally continuous. Downstream reaches of mountain channels which did not become appreciably wider and flatter usually had no sediment patches.

Particle sizes in channel deposits range from silt to boulders 10 feet or more in intermediate diameter (figs. 31 and 26). Some of these sediments were measured to determine their size-frequency distribution, but this subject is discussed later in the report because the deposits studied for grain-size distribution were continuous with downstream alluvial fan and (or) floodplain deposits.

Most of the channel deposits were unsorted. Others

FIGURE 30.—Upstream view of typical log jam in Ginseng Hollow. Rocks of all sizes accumulate behind such dams. Normal water flow, about 3 feet wide, can be seen in foreground just to right of man.

A

B

C

D

FIGURE 31.—Range of particle sizes commonly present in upstream deposits. *A*, Ginseng Hollow. *B*, small ravine near Shaeffer Hollow. *C*, *D*, Polly Wright Cove. Except for *B*, the deposits can be interpreted as either floodplain or mountain-channel deposits. Note normal water flow in left-center foreground of photograph *A*.

showed thin lenses or layers of well-sorted pebbles or cobbles, often in an imbricated pattern. Such well-sorted deposits in a few places formed minor ridges along the edge of the channel, extending for about 15–30 feet downstream parallel to the stream direction. The ridges may be remnants of a deposit that filled the channel bottom at some stage during the flood, with the rocks from the central zone being removed by recession flow. The ridges definitely were formed after the peak flow because they were on the channel bottom well within the maximum cross-sectional flow area, and the peak flow would have easily removed them.

Boulders as well as cobblestones accumulated in imbricated patterns, as figure 32 shows. Such large particles, although definitely moved by the flood, probably were deposited during or soon after the peak flow.

Debris piles behind dams in mountain channels are as much as 200 ft long and extend all or most of the way across the channel (up to 100 or even 200 ft). These wedge-shaped deposits are as much as 20 ft deep at the downstream face. They show little or no stratification or preferred orientation of particles.

Deposits in the normal stream channel downstream from the mountains were under water at the time of the investigation. These submerged sediments were not studied because they probably were altered by the postflood streamflow to the extent that they did not represent flood deposits. Valley sediments located on either side of the normal stream are here treated as flood-plain deposits, except for the alluvial fans and highway deltas.

FIGURE 32.—Downstream view of imbricated boulders left in a stream channel in Edes Hollow.

SEDIMENT SAMPLING, COMPUTATION OF SIZE-FREQUENCY DISTRIBUTIONS

Except for the scattered debris piles in the mountain channels the sediment deposits were inspected for grain-size characteristics and other features.

For various reasons it was usually impossible to define a sedimentation unit (a thickness of sediment deposited under virtually constant physical conditions). This fact, plus practical considerations, dictated that in most cases only the surface material was examined in regard to particle sizes.

The sediments laid down more than a mile or two downstream from the alluvial fans or the head of the main valley were mostly sandy. These fine-grained deposits were sampled in a manner described below and were analyzed by sieving.

In the upstream areas near the head of the valley a typical flood deposit commonly included particles ranging from fine silt to large boulders (fig. 31). These deposits were measured by the "pebble-count" method (Wolman, 1954). We stretched a measuring tape perpendicular to the stream axis across the full width of the deposit and recorded the intermediate axis of the surface particle lying under every two-foot tape increment. Grains smaller than about 10 mm were listed simply as "sand" for convenience, and the intermediate axis of larger particles was measured with a meter stick. The number of individual particles counted ranged from about 30 to more than 300, depending on the width of the deposit. At 5-foot intervals across the deposit a "sand" sample was removed, except in those instances where no fine material happened to occur at a given 5-foot station. These samples represented the top inch or two of sediment. All such sand samples for a given section or reach were combined, and in the laboratory a single composite sieve analysis was made, reflecting an "average" size-frequency distribution for this finer material. If the sand sample contained about 10–15 percent less than 0.062 mm, a pipet analysis of this part was also performed. In these cases the results of the sieve and pipet analysis were directly combined and recomputed as one distribution, on the assumption that the sieve diameter is similar to the sedimentation diameter, within the silt and clay range.

Ideally one would prefer a single size-frequency distribution representing all the particle sizes present in the deposit. Several factors had to be considered to achieve this goal.

Particles for a sieve and pipet analysis are usually obtained with little thought as to possible size differences between grains exposed on the top surface of the deposit and those beneath the surface. The sample usually contains both. The actual grain size fortunately does not influence the quantity of each size sampled, because the opening of the container ordinarily is considerably larger than the grains. The analysis yields percentages by weight, for each size class. Particles in a pebble count or stone count, however, represent particles lying only on the top surface of the deposit and are selected with a definite bias, namely in proportion to the exposed area of the stone. Moreover, the recorded data give percentages on a number basis rather than a weight basis. All of these differences had to be resolved, at least as far as possible, in order to combine the sieve and pebble-count data into a single overall size-frequency distribution.

Kellerhals and Bray (1971) show that if the grain sizes in a deposit are randomly dispersed the number-percentages for surface grains are equal to the weight-percentages of a three-dimensional sample of the deposit, on the assumption that all particles have the same specific gravity. This means that for the sediments analyzed here no conversion factor was needed to transfer pebble-count (number) percentages to sieve (weight) percentages. On this basis we directly combined the pebble-count and sieve-analysis percentages into one continuous frequency distribution, performing in the process an adjustment for the relative proportions of large (pebble-counted) stones and finer (sieved) particles.

The resulting frequency-distribution is a weight-percent distribution representing all particles within the three-dimensional body of the deposit. This distribution, incidentally, is approximately the same as the percent of surface area each size class covers on the top of the deposit. It is not a weight percentage distribution of surface grains, that is, not the percentages that would result if all grains visible from a top view were plucked from the surface and weighed.

The assumptions involved are (a) homogenous dispersal of all grain sizes throughout the body of the deposit and (b) constant particle shape, specific gravity, orientation and packing. Naturally these assumptions are never completely true. Small grains are often less abundant on the surface, as they tend to hide in the voids or be winnowed away; however, we believe we sampled most or all of the pertinent flood sediments before any postflood erosion significantly altered the surface of the deposits.

A simple example will show the computing procedure. At a sampling site (cross section over the surface) on an alluvial fan 59 entries were recorded in a rock count: 14 large stones and 45 particles of "sand." Column 3 of table 7 shows the percentages by number for the large stones, and column 4 lists the percentages on a weight basis for the sieved grains. Directly combining these two distributions would imply that each of the two groups occupies 50 percent of the surface area of the deposits. However, the stone count showed 14 tallies in the large-stone range and 45 recordings of "sand." Large stones, in other words, covered only $^{14}/_{59}$ of the surface area of the deposit. The percentage in each size class for large stones was therefore multiplied by $^{14}/_{59}$ or 0.237, and the percentage in each size class of the sieve analysis was adjusted by a factor of $^{45}/_{59}$ or 0.763 (col. 5). This step produced a single weight-frequency distribution for the entire deposit at the sampling site, and the usual cumulative size-frequency curve could then be drawn (fig. 33).

The sampling at each pebble-count site also included a measurement of the intermediate axis of all of the largest stones within about 10 feet upstream and downstream of the tape section. The theory was that these boulders would be proportional to the stream competence, that is, to the stream's ability to transport sediment. They would represent a limiting particle size showing what range of sizes the flow

TABLE 7.—*Computational procedure of combining a sieve analysis with a pebble count*

[Location of sample: Bryant fan, station 1,800. Adjustment factors: 14/59 = 0.237 for large (pebble-counted) stones, 45/59 = 0 763 for fine (sieved) particles]

(1) Size class (mm)	(2) Number of stones	(3) Percentages within subgroups: large stones	(4) Percentages within subgroups: sieve-pipet analysis	(5) Percent by weight, entire range of sizes	(6) Cumulative percent finer than upper size limit
4,096–2,048	--				
2,048–1,024	--				
1,024–512	--				
512–256	1	7.1		1.7	100.0
256–128	1	7.1		1.7	98.3
128–64	2	14.8		3.4	96.6
64–32	6	42.9		10.2	93.2
32–16	4	28.6		6.8	83.0
	14	100.0			
16–8			14.2	10 8	76.2
8–4			11.4	8.7	65.4
4–2			14.3	10.9	56.7
2–1			19.0	14.5	45.8
1–0.5			15.8	12.1	31.3
0 5–0.25			12.2	9.3	19.2
0.25–0.125			7.1	5.4	9.9
0.125–0.062			3.2	2.4	4.5
0.062–0.031			2.8	2.1	2.1
			100.0	100.0	

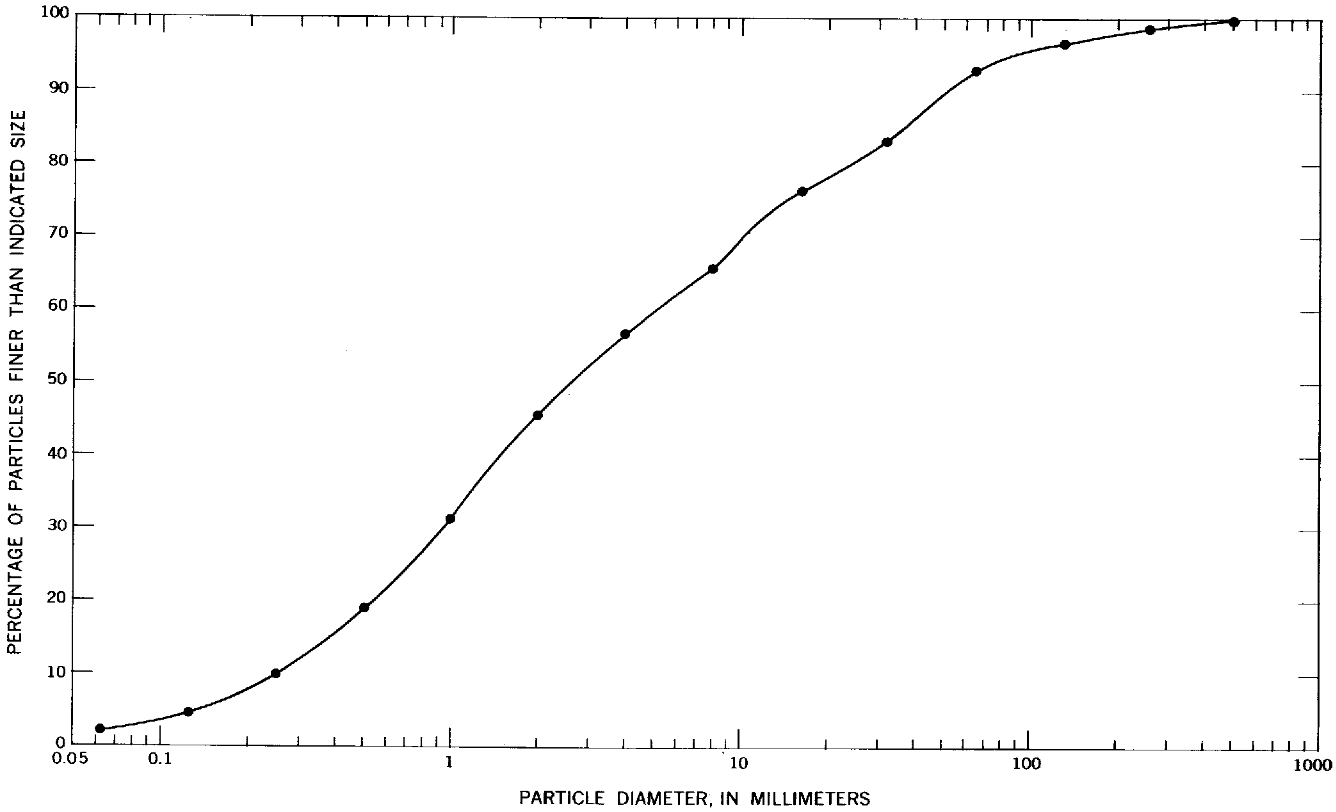


FIGURE 33.—Typical particle size-frequency distribution, based on percent by weight, of a deposit that has a wide range of particle sizes (Bryant fan, station 1800).

could and could not carry, in the region of the sampling site (on the assumption that all sizes were available for transport). Further reflection, however, suggests that the theory of competence is more complicated than might seem at first glance. First, the rate at which sediment entered the channel probably was unsteady, responding in part to the occurrence of debris avalanches on the hillsides upstream. There is in fact no way of knowing (1) whether the total volume or weight rate of sediment introduction was steady (it probably was not) or (2) which sizes of rocks entered the channel at a given stage during the flood. Besides these uncertainties regarding the sediment introduction, there are several aspects of the water discharge which could affect the stream's ability to transport sediment. The amount of water entering the channel, for example, varied with time in some unknown way because of the uneven distribution of rainfall intensity throughout the storm. Was the water discharge highest at the time the largest stone arrived at the head of the study reach? And did the discharge increase downstreamward, as is common, thus increasing the stream competence with distance? Against such a downstream increase in competence would be the flatter channel or floodplain gradient with distance, which would reduce the stream competence. In addition, the channel configuration varied with distance, at some places causing the flow to become narrow and deep and at other places wide and shallow. Such changes in flow cross section can affect the sediment transport. "Stream competence," then, actually involves a number of different factors, all of which might be important in determining the stream's ability to move the stones delivered to it. The largest stones finally deposited at a given section could have entered the channel at different times and could have been deposited and shifted a number of times during the flood. Thus the final areal distribution of large stones represents the combined effects of (1) some unknown rate of sediment introduction for each particle size throughout the flood and (2) varying degrees of stream competence at different times and locations.

In most cases it was possible to tell whether a given boulder had been transported and deposited by this particular flood. However, as a safety factor against any possible misinterpretation and freak stone occurrence, the data on the following pages represent the average size of the five largest stones at each site, within 10 feet of either side of the cross section line.

DEFINITIONS OF SIZE-DISTRIBUTION CHARACTERISTICS

Cumulative particle size-frequency curves were plotted on semilog paper as in figure 33, with percent finer than a given grain size on the ordinate (arithmetic scale). The following simple measures, easily and quickly obtained graphically, were adopted to describe the curves and are used in the subsequent discussion.

The average particle size d_{av} was estimated by averaging the grain sizes corresponding to the 10th, 50th, and 90th percentiles of the distribution (d_{10}, d_{50} and d_{90}, respectively). Since grain size is plotted on a log scale, d_{av} is the antilog of the quantity (log d_{10}+log d_{50}+log $_{90}$)/3.

The range, spread or sorting of particle sizes (S_o) is reflected by the percentile range d_{90}–d_{10}. Because of the log scale for grain sizes, this range is log d_{90}–log d_{10}, or log (d_{90}/d_{10}). The d_{90} and d_{10} values were selected because they cover the widest possible definable range common to all curves. This measure includes most of the distribution and is not affected by the actual sizes involved. S_o=0 means "perfect sorting" in the range between d_{90} and d_{10}, that is, all particles in this range have the same size. In such case d_{90}=d_{10}, the quotient is 1.0 and the log or sorting is zero. S_o values of 1.0, 2.0, 3.0, and so on mean that the percentile range d_{90}–d_{10} covers one, two and three log cycles of grain size, respectively.

The measure of asymmetry of the distribution or skewness *sk* is (log d_{10}+log d_{90}—2 log d_{50})/(log d_{90}—log d_{10}). This is similar to the relative skewness listed in Croxton and Cowden (1939, p. 254) but deals with a distribution of the logarithms of particle diameters. Possible values of *sk* fall within ±1.0, and in a symmetrical distribution *sk*=0. A distribution having a preponderance of small grains has a positive value of *sk*, whereas negative values indicate a preponderance of large particles.

The average size (geometric mean) of the five largest stones (d_{ls}) is the antilog of (sum of logs of intermediate diameters, divided by five). The geometric mean, rather than the arithmetic mean, is used because particle sizes customarily are treated on a logarithmic scale.

ALLUVIAL FANS

Upon escaping the confinement of the rather narrow mountain channels, several of the swollen streams deposited a large amount of sediment in the form of alluvial fans (figs. 34 and 35). Such alluvial fans or cones of debris seem to be common depositional features of floods in hilly areas (Eisenlohr, 1952; Jahns, 1947; Chawner, 1935). Fans actually did not form on many streams, as the ground configuration often did not favor the formation of fans. For example, the mountain channel may have widened considerably before reaching the broad valley, so that no sudden spreading out of the flow occurred. In other cases the channel leaving the mountains was deep enough to contain much of the flow for some distance beyond the mountain front, so that only typical flood-plain deposition resulted. Some drainage basins had very few debris avalanches in the headwaters, so the streams in those basins carried relatively little sediment.

Some of the more notable alluvial fans were formed at the home of R. L. Bryant at the mouth of the Ginseng Hollow (fig. 35), at the home of W. A. Campbell near Shaeffer Hollow, at the home of S. K. Wills at the head of Wills Cove, and at a downstream site on Cub Creek (fig. 34). Fans ranged in length from a few feet to nearly 0.4 mile. For example, the Bryant fan was about 2,000 feet long, the Campbell fan about 1,830 feet and the Wills partial fan about 850 feet. (Another stream eroded the downstream part of the Wills fan.)

The cascade of water, rock and rubble that formed an alluvial fan often caused considerable damage to buildings and property, because fans usually occupied areas of regular land use. At R. L. Bryant's home at the mouth of Ginseng Hollow, part of the house was torn away, including the section in which Mr. Bryant was sleeping at the time. (He rode 140 ft downstream on his mattress, along the edge of the flow, sailed in through the open door of his barn and spent the rest of the night sitting in the upper part of his barn—a lucky survivor.)

Alluvial fan sediments were devoid of any recognizable sedimentary features, such as bedding, lamination, crossbedding, mud cracks, imbrication, or other preferred orientation of particles. Except that many rocks seemed to be lying on their long and intermediate axes, the deposits looked as if they had been shaken in a giant mixer and dumped.

Also dumped on fan surfaces were many trees and branches, isolated or in piles. Isolated large trees tended to be parallel to the flow direction, although the roots could be pointing upstream or downstream. (This tendency could be a useful aid in determining flow direction in ancient sediments, if remains of

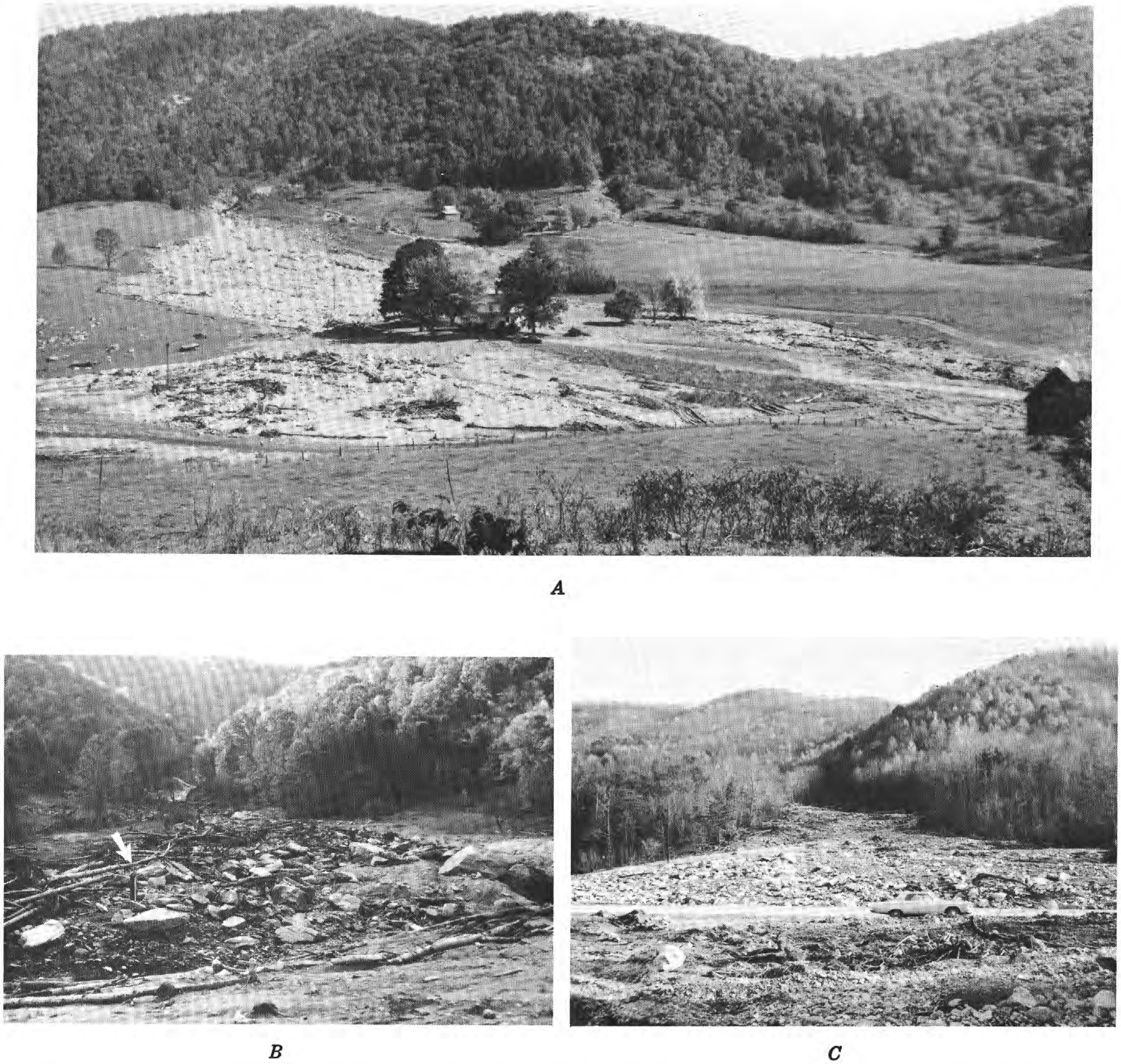


FIGURE 34.—Upstream views of alluvial fans which formed where the stream escaped the confinement of the mountain channel. *A*, Campbell fan. *B*, Wills partial fan (arrow points to man for scale). *C*, fan entering valley of Cub Creek about 2 miles down from headwaters.

trees could be found.) Trees in log jams tended to be oriented at oblique angles to the flow direction.

The flow depth over an alluvial fan should decrease proceeding downstream from the apex. There was very little evidence to indicate the flow depth over the Nelson County fans. From some high-water marks on a hillside along the left-bank edge of the Bryant fan, the peak flow depth about half way down the fan was judged to be about 3–4 feet.

In all cases the stream at the fan apex, and occasionally even further downstream on the fan, cut as much as about 10 feet vertically into old alluvial deposits and soil horizons (fig. 36). In fact the general ground configurations reveal many old alluvial fans where mountain channels enter the broad valleys in Nelson County. For example, all the fresh fan deposits discussed here were formed on old alluvial fans; so the 1969 storm was only one in a chain of

FIGURE 35.—Aerial view of Bryant fan and lower part of Ginseng Hollow (Photograph courtesy of Virginia Division of Mineral Resources).

similar major events in the geomorphic history of the area.

For a few weeks after the flood the creeks on the fan tended to show a braided pattern, especially toward the downstream part of the fan.

AMOUNT OF DEPOSITION

Sediment was transported to a fan over a period of time during which the discharge and sediment transport rate undoubtedly varied considerably. The

FIGURE 36.—Eroded zone at apex of Campbell fan.

deposition or growth of the fan therefore proceeded in stages. Due in part to the varying erosion and deposition with time at the apex of the fan, the flow arriving at the fan apex probably was diverted to slightly different directions during the flood, and the successive depositional stages probably manifested themselves as "tongues" of deposit which varied in areal location on the fan. As a result, the fan thickness varies from one spot to another.

We analyzed amounts of deposition on the Bryant and Campbell fans by setting up stations down the fan centerline. At a number of these stations the cross-sectional area of the deposit, in a plane normal to the fan centerline (that is, roughly normal to the flow direction) was determined by stretching a measuring tape over the full width of the fan and recording the depth of fill along this cross section. The depth varied from place to place, but usually it did not exceed 1–2 feet. Depth was determined on the basis of exposure of the original surface at some points or by digging through the deposit. Along any given cross section it was possible to define increments over which the depth of fill was approximately constant, and the many subareas were summed to get the total cross-sectional area of deposition, for each downstream station.

The sediment deposited on a fan should depend in part on the water discharge and fan slope. However, the problems of which type discharge (for example, storm-average or various instantaneous rates) and slope (local slope or average slope from fan apex to top) determine or indicate the final amount and distribution of the fan deposit are unresolved. These in fact would be interesting subjects for future research. As estimated earlier, the storm-average water discharge at the apex of the Bryant fan was roughly 1,050 cfs. For the apex of the Campbell fan the rational formula yields an estimate of only 134 cfs, owing to the much smaller drainage area (0.08 sq mi). Average fan slopes (fig. 9) were 0.045 ft per ft and 0.080 ft per ft for the Bryant and Campbell fans, respectively. The product of discharge times slope, roughly indicative of stream power, is thus more than about four times greater for the Bryant fan than for the Campbell fan. However, aside from the fact that the storm-average discharge and overall fan slope may not be the factors most closely related to the volume of the deposit, the sporadic nature of the amount and rate of sediment supply to the two fans makes comparisons difficult.

Figure 37 shows how the cross-sectional area of deposition varied with distance down two fans. The actual horizontal distance has been converted to proportional distance from apex to downstream tip of fan, in order to compare the two fans.

The apex or upstream region of an alluvial fan typically showed both erosion and deposition. On balance either process could predominate, depending on the width of the upstream mountain channel relative to the width of the fan apex. At the Bryant home the channel upstream was not very constricted (fig. 35), and some deposition occurred before the stream left the mountain channel. Consequently deposition predominated over erosion around the fan apex. The mountain channel upstream from the

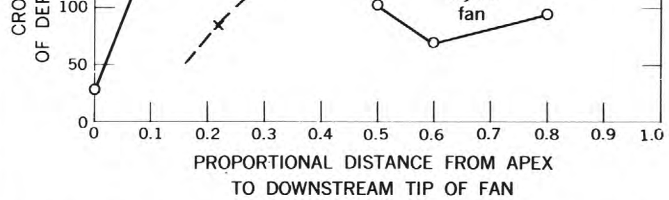


FIGURE 37.—Amount of deposition with distance down fan, for the Bryant and Campbell alluvial fans.

Campbell fan was deep and V-shaped (fig. 36), and the fan had net erosion for about the upstream 0.1 of the length of the fan.

On the basis of figure 37, the Bryant and Campbell fans each received about 300,000 cubic feet of sediment during the flood.

A striking similarity in the two curves of figure 37 is that the volume of deposition builds up abruptly from very little at the apex to a peak amount and then diminishes abruptly toward the downstream tip of the fan. On the Bryant fan the peak deposition occurs about 0.3–0.4 of the way down the fan, while on the Campbell fan the peak is about 0.5–0.7 of the way along the fan.

CROSS-FAN PROFILES

Cross-fan profiles are surface profiles measured normal to the long axis (centerline) of the fan, that is, measured approximately parallel to the mountain front. Such profiles normally would be convex owing to greater deposition in the middle part (Bull, 1964, p. 114).

Figure 38 contains cross-fan profiles for the Campbell and Bryant fans. The Bryant profiles are convex except for the first profile, which is 0.2 of the way down the fan. The original stream ran along the south (left bank) edge of this fan; so the topography is generally lower along that edge. In the downstream 0.4 of the fan length, the deposit splits into two prongs (fig. 35). The apex of the Campbell fan, as mentioned above, is mainly an erosional rather than a depositional surface. Significant deposition began around 0.17 of the way down the fan, but the profile for that station generally reflects erosional processes more than deposition. Cross profiles further down the fan are only slightly convex. In the downstream half of the fan, the main drainage is along the left edge; therefore, as with the Bryant fan, the surface is lower along that edge.

GRAIN-SIZE CHARACTERISTICS

On the Bryant, Campbell, and Wills fans particle-size analyses were made across the full width of the deposit, at various stations down the fan. Figure 39 is a graph of the various grain-size characteristics as a function of distance down the fan. The grain-size features at a given station down the fan centerline represent the particles over the full width of the fan, at that downstream station. Except for the large-rock measurements, the sampling stations on the Wills fan were too few in number (2) to warrant plotting.

Average grain sizes range from about 2 to 6 mm. Although one might expect the average grain size to decrease with distance down the fan (Chawner, 1935), such a trend is not very pronounced (fig. 39). The largest value of d_{av} occurs not at the apex of the fan but a short distance downstream—0.2 of the way from the apex to the lower end of the Bryant fan and about 0.4 to 0.5 of the way down on the Campbell fan. Deposition did not begin as near the apex on the Campbell as on the Bryant fan, as mentioned above.

The geometric mean of the five largest stones ranges from about 500 to 1,700 mm. On the Bryant fan the size of the largest stones generally shows a gradual decrease with distance of travel. Large stones on the Campbell fan increase in size over the upper 0.4 of the fan length, then gradually decrease in size with further distance down the fan. The trend in fact is quite similar to that of the average grain size for the Campbell fan, but this is not the case on the Bryant fan. Large stones on the Wills partial fan show no particular tendency to increase or decrease with distance, but this truncated fan was less than half the length of the other two. Pashinskiy (1964, p. 277) noted that on debris cones bordering the Psezuapse River, near the Caucasian Mountains, USSR, the coarsest material generally is located in the middle part of the cone. Lustig (1965, p. 145), on the other hand, found that large stones on California fans were most abundant near the apex region.

The largest stones on the Bryant fan are smaller than those on the Campbell and Wills fans. This is probably because the mountain ravines upstream from the latter fans were confined, and practically no channel deposition occurred. The Ginseng Hollow channel upstream from the Bryant fan, on the other hand, had a greater width/depth ratio and contained debris jams and channel deposits (fig. 26*B*, 30, 31*A*). These deposits included many of the large boulders which otherwise would have been dumped on the fan.

In general the alluvial fans are characterized by a wide range of particle sizes—from clay or silt up to boulders about 2 meters in intermediate diameter. The sorting, in other words, is quite poor and was probably affected by deposition of finer sediments at the end of the flood. The d_{90}–d_{10} range covers as much as 3.7 log cycles of grain size along most of the length of the fan (fig. 39). Sorting on the Campbell fan stays virtually constant at a value of about 3.4 along the entire study reach. On the Bryant fan the upstream 0.6 of the fan length also has virtually constant sorting, an S_o value of about 3.4 to 3.7, but

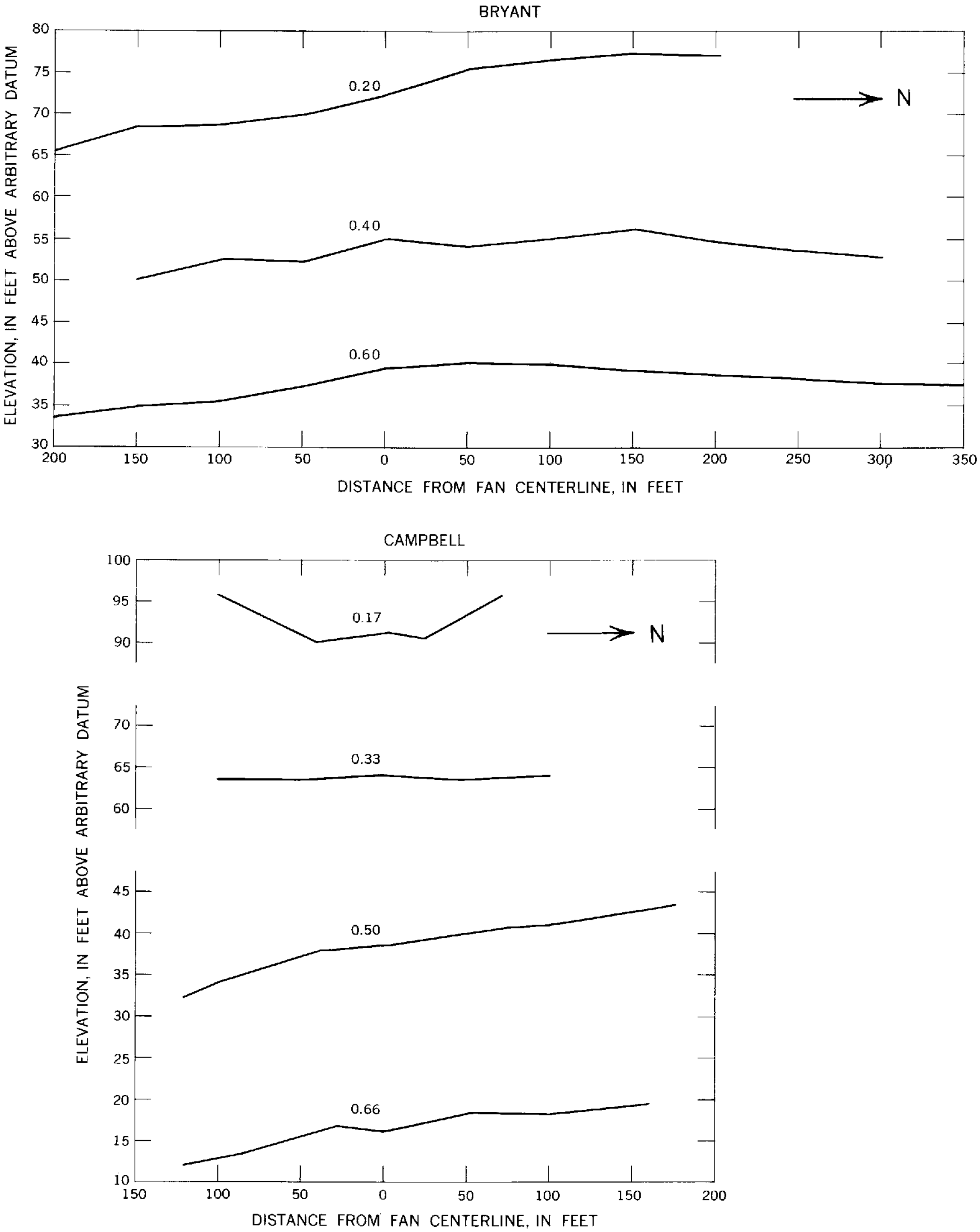


FIGURE 38.—Cross-fan profiles, looking down the fan, for the Bryant and Campbell alluvial fans. Number next to line of profile indicates proportional distance down fan, from apex to toe. Profiles reflect all pre-1969 deposition, in addition to Camille flood deposition.

the sorting improves with greater travel distance, so that at 0.9 of the way down the fan the d_{90}–d_{10} ranges encompasses only about 2.3 log cycles of the grain size—perhaps reflecting the fewer large stones.

All skewness values are positive, indicating a predominance of the smaller grain sizes. The weight of the smaller grains in the fan, in other words, is greater than that of the large boulders. With distance along the fan the skewness values change inversely with the average grain size (fig. 39). The

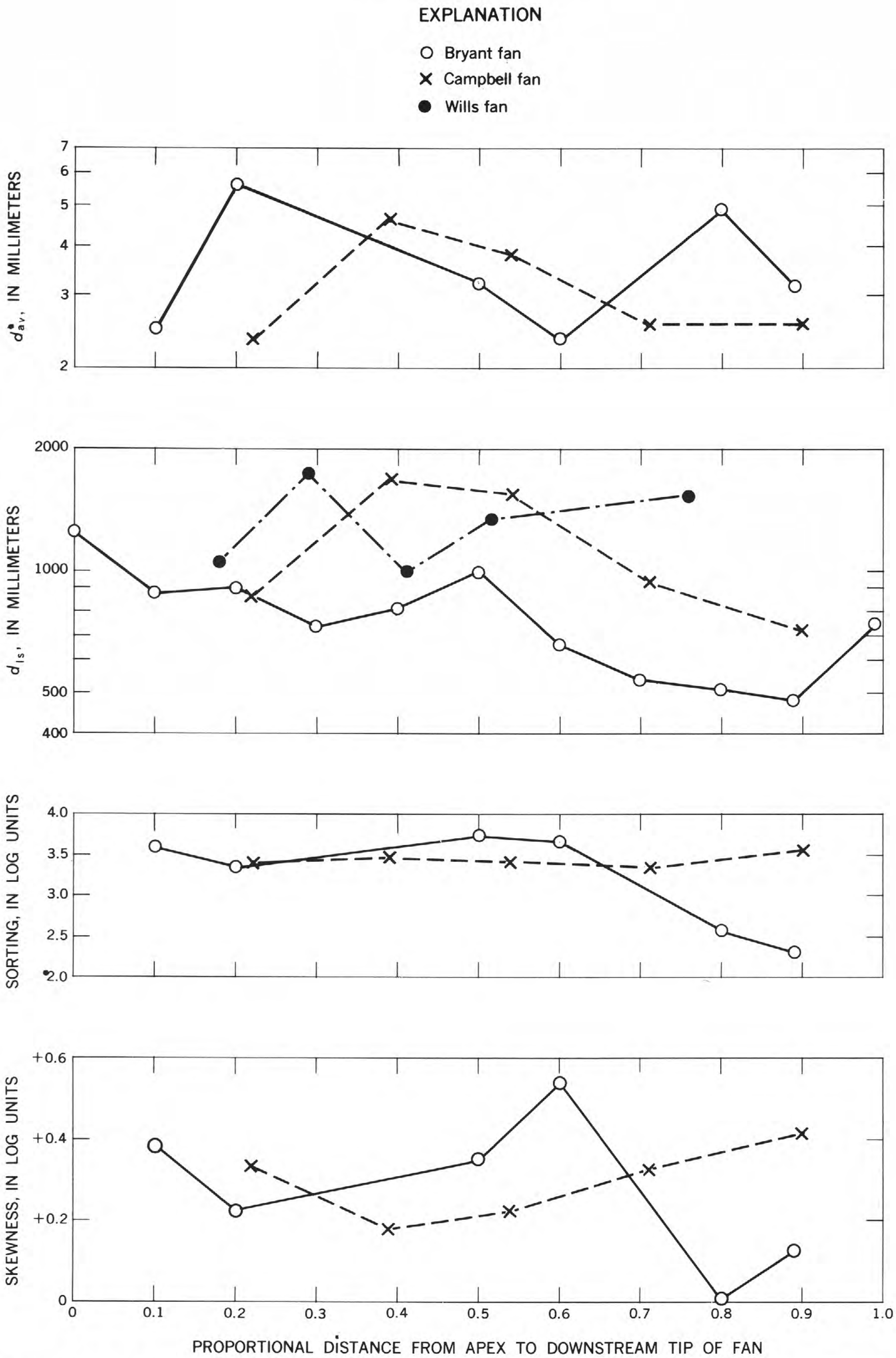


FIGURE 39.—Change in particle size-distribution characteristics with distance along alluvial fans.

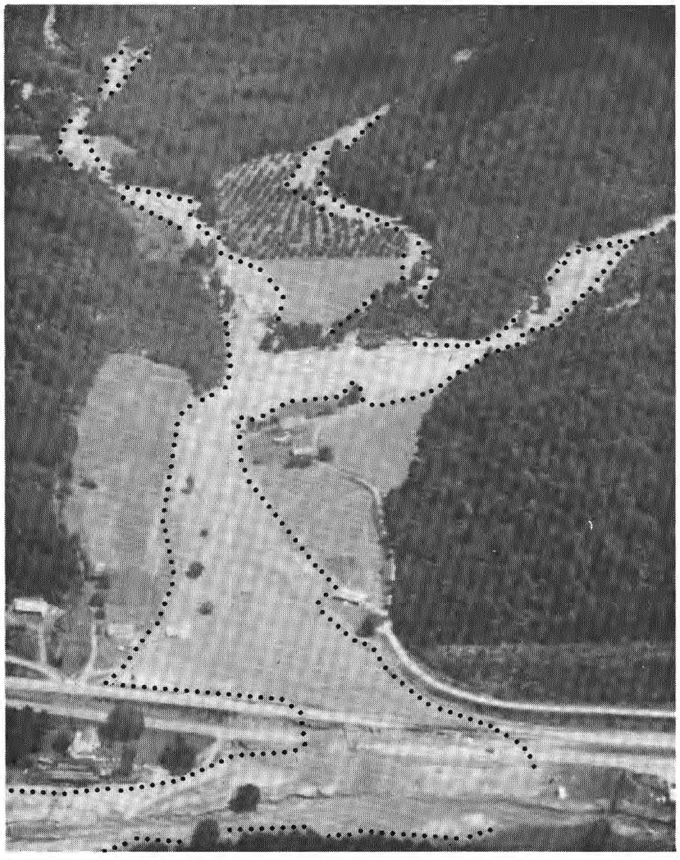
reason for this is that for a given range of grain sizes a decrease in d_{av} means the smaller grains are becoming relatively more abundant in terms of weight than the large boulders, so that *sk* takes on larger positive values. Conversely, given a distribution having a preponderance of smaller particles, an increase in d_{av} means that the larger stones are becoming more important is terms of weight. The distribution therefore tends to become more symmetrical, that is, the *sk* values tend toward zero. If the large stones were to become more abundant than the small grains, *sk* would be negative.

HIGHWAY DELTAS

Vast bodies of sediment accumulated in deposits herein called highway deltas (figs. 40 and 41). Passageways under small bridges or highways became clogged with vegetation and other debris, probably early in the flood. In some cases the raging stream scoured a new channel by eroding the highway next to the bridge, but in other cases the highway became an effective dam. A large body of water with a mean flow velocity considerably slower than that of the unaffected flow formed just upstream from the dam. The stream filled sediment into the ponded water until a sediment wedge or delta had build up to the road level, after which time the water carried the rest of its load across the highway and on to some farther destination.

Where present, the highway deltas are within about 4 miles of the stream headwaters. Culverts or bridges this close to the headwaters are rather small and became plugged fairly easily. Farther downstream the streams are larger and flowed under major bridges that washed out completely or that did not become plugged with debris, so that deltas did not form.

The dimensions of highway deltas are limited laterally by the valley or floodplain walls and vertically by the height of the roadway. Delta lengths often are difficult to define because the deposit gradually merges upstreamward with the regular floodplain deposits. One of the smallest deltas is on route 151 about a mile south of Brent Gap. The deposit is about 200 feet wide at the highway, 150 feet long, 10 feet deep at the thickest place on the downstream face, and gradually thins laterally. The resident whose house is only about 50 feet outside the edge of this deposit testified that all of this sediment arrived within about a 2-hour period. Other deltas (fig. 40) are as much as about 700 feet wide and 10 to 12 feet thick at the downstream face. Lengths are always



A

B

FIGURE 40.—Delta that formed where drainage from Edes and Melton Hollows meets highway U.S. 29. *A*, Overall upstream view looking up Melton Hollow. Edes Hollow forks to the right at middle of picture. U.S. 29, a divided four-lane highway, in foreground. Dotted linework indicates boundary between grass and sand areas. (Photograph by Ed Roseberry.) *B*, Closeup aerial view of downstream part of delta. An inhabited house and two smaller buildings disappeared from this part of the delta during the flood. House in center of photograph is nearly half buried in sand. Construction crew at work unplugging culvert. (Photograph courtesy of Virginia Department of Highways.)

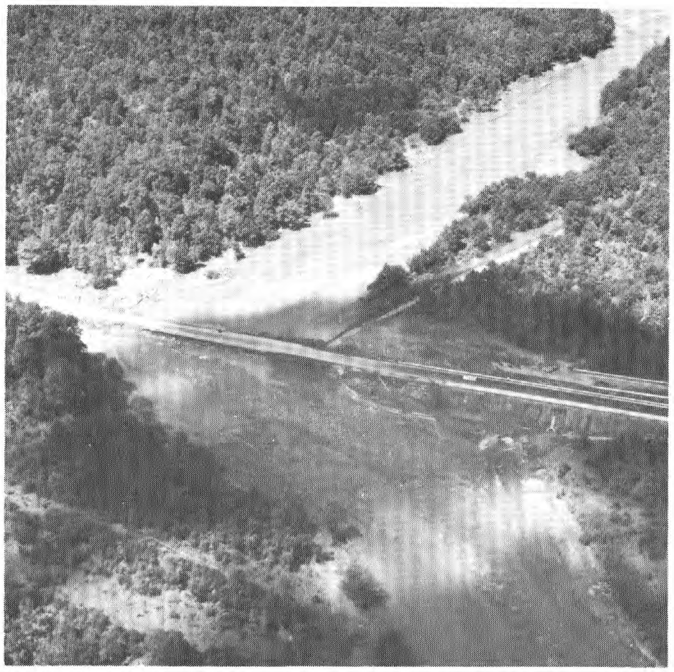


FIGURE 41.—Upstream view of delta deposited by Dillard Creek at highway U.S. 29. Flow is from upper right to lower left. One half of the four-lane highway is completely buried. (Photograph courtesy of Virginia Department of Highways.)

greater than maximum widths and can be defined conservatively as about 1,000 feet for the examples shown in figures 40 and 41.

The large delta that formed where Edes Hollow meets highway U. S. 29 (fig. 40) has an estimated cross-sectional area of 5,600 sq ft at the downstream face (about 700 feet wide and an average of 8 feet deep). If the volume is determined as approximately $\frac{1}{3}$ $\times$ area of downstream face $\times$ length, the delta contains an estimated 1.9 million cu ft of sediment. The drainage basin upstream is about 1.3 sq mi, of which about 0.9 sq mi experienced erosion that, from aerial photographic and field inspection, seemed as severe as that in Ginseng Hollow, Polly Wright Cove and Wills Cove. In the latter three basins the severe erosion contributed an average of about 4.9 million cu ft of sediment per square mile. If the avalanche and channel erosion in the Edes Hollow basin is similar to that in these three basins, Edes Hollow contributed about 4.4 million cu ft. The highway delta therefore contained an estimated 39 percent of the sediment eroded from the mountainous part of the basin. Of the remainder, some was deposited upstream and some was carried downstream from the delta.

Because the Edes Hollow delta is only about a mile from the drainage divide, the flooding and deposition in the delta probably did not last for more than 7 or 8 hours. A storm-average sediment-transport rate for the region of the delta can be estimated from the amount of sediment contained in the delta and the period of accumulation. Actually the postflood delta surface was filled to the lowest level of the highway (fig. 40*B*, upper left) and showed that some sediment definitely moved across the highway. Consequently, the volume of sediment left in the delta (about 1.9 million cu ft) divided by the period of deposition will produce a minimum estimate of the average transport rate. If 1.9 million cu ft accumulated in 8 hours, the storm-average sediment-transport rate for the apex of the delta was about 66 cfs, or 6,600 lb per sec (285,000 tons per day). Field notes indicate that the water surface in the vicinity of the apex of the delta was about 200 feet wide. The approximate storm-average sediment-transport rate at the apex of the delta therefore was about 33 lbs per sec per foot of channel width. Instantaneous rates may well have been higher.

The estimated average transport rate in any case represents just the delta region, rather than points upstream from the delta. The transport rate into the delta reflects the combined contributions of all upstream tributaries and debris avalanches, minus any deposition occurring upstream from the delta.

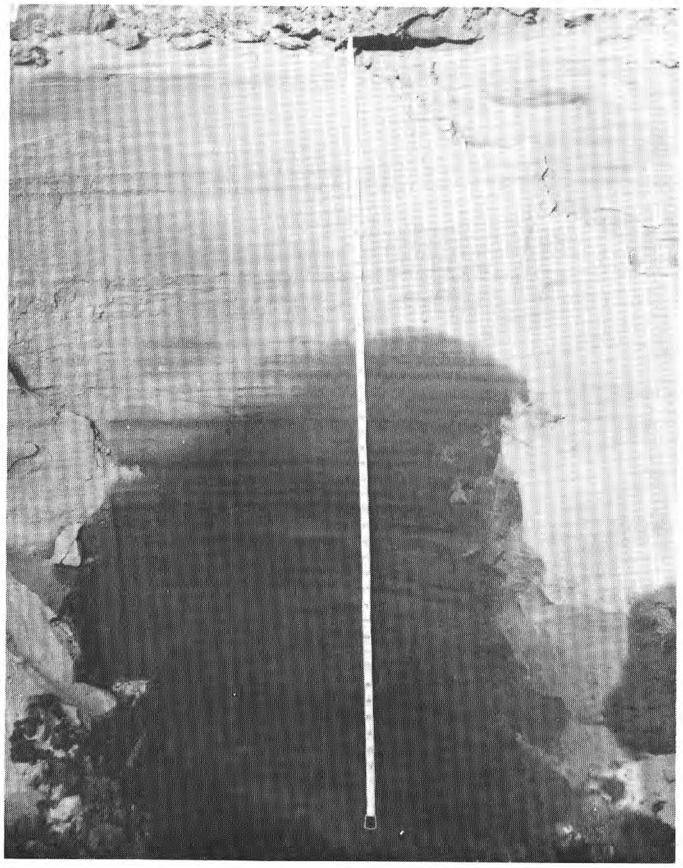
The large delta that formed where Dillard Creek meets highway U. S. 29 (fig. 41) has a cross-sectional area at the highway of about 3,150 sq. ft. (about 350 feet wide by an average of 9 feet deep). It gradually merged with flood-plain deposits upstream. If 1,000 feet were considered the length, the delta contained about 1 million cu ft of sediment. The Dillard Creek basin has 3.3 sq. mi. of mountainous area, 1.6 sq. mi. of which is in Wills Cove, where the estimated yield was 4.9 million cu ft per sq mi. We hiked extensively over the Fortunes Cove area (the other part of the Dillard Creek basin) and found it to be eroded to a similar degree as Wills Cove. If the Wills Cove erosion yield of 4.9 million cu ft per sq mi applies also to Fortunes Cove, the Dillard Creek highway delta trapped an estimated 22 percent of the sediment yield. This figure is somewhat less than the 39 percent obtained for the Edes Hollow delta because the Dillard delta is about 2 miles farther downstream from the sediment source, and much of the eroded sediment was deposited on flood plains upstream from the delta.

As the flood receded, the flow on the surface of highway deltas began meandering and often assumed a braided pattern (fig. 40*B*, and 41). Toward the end of the flood, the flow became very slow or even stopped over much of the delta, as indicated by the nature of the deposits, described below. When workmen subsequently unplugged the culverts, the deposits in the central parts of the deltas were usually eroded considerably as the stream worked its way down to the culvert level.

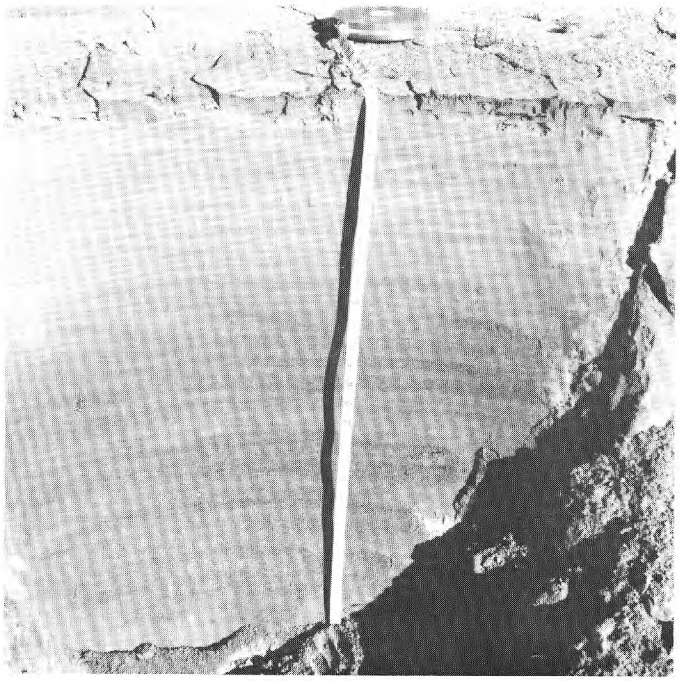
From the relative elevations of the highway and the floor of the delta, it appears that the water depth in which these deltaic deposits were formed ranged very approximately from 1 or 2 feet for the topmost sediments to 10 or 12 feet for those deposited earliest. The 10–12-foot estimate is a maximum depth; in all probability the water level rose as the sand laminae accumulated, so that the water depth during deposition was probably somewhat less than 10 feet.

SEDIMENTARY FEATURES

The sand deposits which make up nearly all of the deltaic sediments have distinct layers, approximately horizontal, ranging in thickness from a fraction of an inch to about 1 inch (fig. 42*A*). These laminae are present from the base of the deposit to within a few inches of the top surface, and exist over the entire area of the delta. Single laminae cannot always be traced very far laterally, and they are often nearly



A



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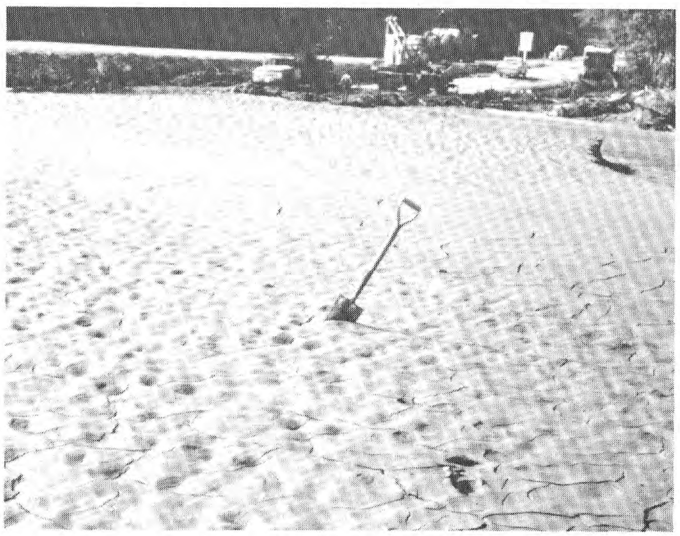


C

FIGURE 42 Above and right.—Sedimentary features of the Dillard Creek highway delta. *A*, Sand laminae. Section shown here is about 5 feet thick. Units on measuring tape are feet and tenths of a foot. *B*, Sand laminae capped by cross-bedded layer of fine sand, with mud layer on top surface. This sequence probably reflects diminishing water flow over the delta. Flow direction uncertain. *C*, Ripples and mud cracks.

inconspicuous in some areas and well defined in others.

In addition to color differences, laminae in many cases differ from one another in grain size. For example, although most laminae are medium and fine sand, some are noticeably coarse in texture. Lenses as much as 6 inches thick, consisting of gravel or pebbles, are present but are not very common. Such lenses, where present, are usually at or near the bottom of the deposit. Sometimes rocks up to 6 or 8 inches in diameter are found within a sand or gravel matrix. These larger particles can occur singly, on no common level, or they may occur in lenses which extend varying distances—five to several hundred feet in length.

The sand laminae in the delta on Dillard Creek at U.S. 29 (fig. 41) are capped by a 3–6-inch layer of fine sand that shows well developed small-scale cross-bedding (fig. 42*B*). These features are known to form only during hydraulically tranquil (subcritical) flow and slow sediment-transport rates. They probably are due to the braided streamflow that developed toward the end of the flood. Water depth at the time of deposition probably was on the order of one to three feet. Since the local flow direction in the braided network probably varied considerably from time to time, there is no assurance that a given exposure is cut parallel to the flow direction. The cross-stratification was not preserved in the center of the delta, perhaps because such strata in this region were eroded when the culvert was cleaned.

The top surface of most highway deltas was covered with a layer of mud (fig. 42*C*) that ranged in thickness from a fraction of an inch to about 1 foot. This mud layer in most cases is missing along the central part of the delta, so if originally present it probably was swept away when the culvert was unclogged. The mud probably was deposited by slow-moving or almost still waters, near the end of the flood.

Ripples and mud cracks are common surface features of highway deltas. Figure 42*C* shows mud cracks in flat areas as well as superimposed on ripple marks. The ripples may or may not be symmetrical, and it is quite difficult to determine from them the current direction at the time of deposition. Wavelengths of these ripple marks are commonly about 0.25–0.5 foot, and heights are about 0.05 foot or less.

In the days and weeks after the flood some settling of deltaic deposits, especially the small grains, occurred as water gradually escaped from the pore spaces. The final thickness of such deposits therefore is less, by some unknown factor, than the original thickness.

VERTICAL VARIATIONS IN GRAIN-SIZE CHARACTERISTICS

Sediment samples at selected vertical intervals were obtained at a total of four sites on the various deltaic deposits. Two locations were on the delta where Dillard Creek meets highway U. S. 29 (fig. 41); a third site was on the Edes Hollow–U.S. 29 highway delta (fig. 40); and the fourth was on a small delta (about 200 ft wide by 150 ft long by a maximum of 10 ft deep) in the headwaters of Hat Creek on State route 151. Samplings at the latter site and at one Dillard Creek site were on the front (downstream) face of the delta. The second location on the Dillard Creek delta was 200 feet upstream from the front face, about 0.2 of the way from the front face to the apex. On the Edes Hollow delta the sampling was about 50 feet upstream from the front face (about 0.05 of the distance from the front face to the apex). The sites were approximately on the center axis of the delta except for the spot 200 feet upstream on the Dillard delta. This site was about 20 feet from the left-bank edge of the delta and was chosen because of the good vertical exposure.

In all four vertical sections the deposited material generally became finer as the flood progressed, that is, the sediments became finer upward. Table 8 gives the pertinent size-frequency characteristics. Depths greater than 3.5 feet on State route 151 delta were not sampled, but field notes emphasize that the lower sediments were noticeably coarser than the overlying ones. At the Edes Hollow site the lower half of the deposit does not show a progressive coarsening with depth, but the upper half definitely does. Thus

TABLE 8.—*Vertical variations in grain-size characteristics in highway deltas*

Delta	Depth below top surface (ft)	d_{av} (mm)	S_o	sk
State route 151	0.5–1.5	0.039	1.16	−0.11
	2.5–3.5	.43	2.09	−.46
Edes Hollow	.5	.22	.55	−.12
	2.5	.36	.84	.01
	4.5	.60	.97	−.13
	6.5	.48	1.05	−.24
	8.5	.40	1.01	−.32
Dillard Creek–front face	Top surface	.09	.58	.02
	.5	.11	.55	−.21
	2.5	.18	.68	−.38
	4.5	.19	.79	−.42
	6.5	.24	.93	−.50
	8.5	.30	.88	−.45
Dillard Creek–200 ft upstream	Top surface	.01	1.47	−.58
	.3	.07	59	−.06
	1.0	.12	.71	−.17
	2.0	.24	.84	−.32

during the depositional phase of the flood the coarsest material dropped out earliest, with finer and finer sizes being deposited as the flood progressed. The rate at which the average grain size decreases varies considerably from one sampling location to the next.

The average grain size seems to depend primarily on how close the delta is to the sediment source (mountain hillslopes). The State route 151 delta was closest to the hills (about ¼ mile), and this site had more gravels and pebbles than the other locations. The Edes Hollow delta is about ½–1½ miles from the sediment sources and had smaller average grain sizes. Sediment in the Dillard delta probably had traveled 3–4 miles and thus was the smallest of the group. Changes in grain size with distance are discussed in more detail in the section on "Floodplain Deposits."

The first sediments deposited had the widest range of grain sizes, and as deposition continued the sediments became better sorted (table 8). The only exception to this general trend was the top mud layer at the site 200 feet upstream on the Dillard delta; because of the clay sizes in this sample, the size curve extends over a greater range. The mud layer at the face of the delta was washed away before it could be sampled, when highway crews unclogged the culvert.

The actual values of sorting and the rate at which the sediments became better sorted upward seem to depend primarily on how close the delta is to the sediment source (mountain hillslopes). The State route 151 delta was close to the hills (about ¼ mile) and consequently received a wider range of particle sizes (more of the gravels and pebbles). The sediments in this delta are the most poorly sorted. Field notes suggest that sorting improves upward most rapidly at this site. At the Edes Hollow delta, which is somewhat farther from its sediment source, sorting is much better than at State route 151 and shows less overall improvement upward. The sediments in the Dillard delta had traveled the farthest, have the best sorting, and improve upward at the slowest rate.

Skewness values are mostly negative, indicating a preponderance of the coarser grains over the fines (table 8). Sediments laid down first were the most skewed, and the size distributions generally became more and more symmetrical as the delta-building progressed. There are no noticeable trends regarding any effect of proximity of source material on rate of change of skewness.

To summarize the trends in size-frequency distribution with time: as the flood progressed, the sediments transported to and deposited in highway deltas became increasingly finer in size, better sorted, and more symmetrical in weight-frequency distribution. Most of these trends also appeared, but often were less pronounced, in the depth variations of flood-plain deposits, as discussed below.

FLOOD-PLAIN DEPOSITS

Flood-plain deposits (figs. 43, 44) were laid down on the gently sloping ground immediately adjacent to the normal stream channel. As used in this report, the term "flood-plain deposits" excludes particles left in mountain channels and, where present, alluvial fans. Thus flood-plain deposits begin at or just below the mountain front at the head of the cove and continue, in some places intermittently, for many miles downstreamward.

The volume and grain sizes of flood-plain deposits depend to some extent on local valley topography. Where the valley narrows, the mean flow velocity probably increased and less material was deposited. Conversely, more sediment was deposited in wider and flatter sections of a valley. In choosing the study areas we tried to avoid either of these extremes. Some reaches underwent both erosion and deposition.

Very little flood-plain deposition occurred in valleys which had few debris avalanches in the headwaters.

Flood-plain deposits are as much as 5 feet thick but generally are from about 0.2 foot to 3 feet. As one would expect, at any downstream location they tend to be thickest toward the center of the valley and thinner near the edges. Flood-plain widths range from a few feet to about 500 feet. The floods were nearly everywhere many times wider than the original channel.

SEDIMENTARY FEATURES

Flood-plain deposits within a mile or so of the headwaters and avalanche areas have no discernible sedimentary structures. Rather, they are quite similar to alluvial fan deposits in that all particle sizes, ranging from silt to boulders, seem to have been dumped all over the flood plain (figs. 43, 44*A*). Large rocks and gravel in flood-plain deposits usually did not accumulate in any distinctive way, such as in bars or splays.

Farther downstream the deposits consist mostly of sand. From three sites studied in detail—one each on Rucker Run, Dillard Creek and Muddy Creek—and from other general observations, it appears that such sandy flood-plain deposits in Nelson County have several characteristics.

The laminations are a noticeable feature of nearly all sandy flood-plain deposits (fig. 45). These lami-

FIGURE 43.—Upstream view of typical valley flood-plain deposits, headwaters of Davis Creek. Note damage to orchard. (Photograph courtesy of Virginia Division of Mineral Resources.)

nae are generally about 0.01 foot thick or less. Sometimes they appear to extend over the whole flood-plain width, but in other cases a single layer can be traced only for about 5 feet before it pinches out between other laminae. In addition to color differences, some laminae are distinguishable by a slightly coarser texture. Also, some layers are different in that they do not remain in the same plane as most of the others; instead, these errant laminae may deviate or dip by as much as 0.2 foot over a 2-foot horizontal distance. They seem to be isolated, however, and are rather rare.

The laminated sand often contains an occasional 1–2-inch pebble.

The exposure near the center of the valley on Rucker Run (fig. 45) has conspicuous pockets or lenses of varying textures. The pockets usually include coarse sand and may have gravel and a few pebbles. Normal laminae, on the other hand, seemed to consist mainly of medium and fine sand. As figure 45 shows, the top and bottom surfaces of the pockets are quite irregular and can rise or fall by as much as a foot or so over about a 5-foot horizontal distance in the approximate direction of flow.

No crossbedding was observed at any of the three flood-plain sites studied in detail.

Ripple marks are common surface features of the sandy flood-plain deposits and evidently formed as

A

B

C

FIGURE 44.—Common flood-plain deposits. *A*, Rucker Run about 0.3 mile downstream from head of valley. Circle around man shows scale. *B*, Rucker Run about 0.75 mile upstream from highway U.S. 29. *C*, Muddy and Davis Creeks about 0.2 mile south of Woods Mill, looking upstream (Photograph courtesy of Virginia Department of Highways.)

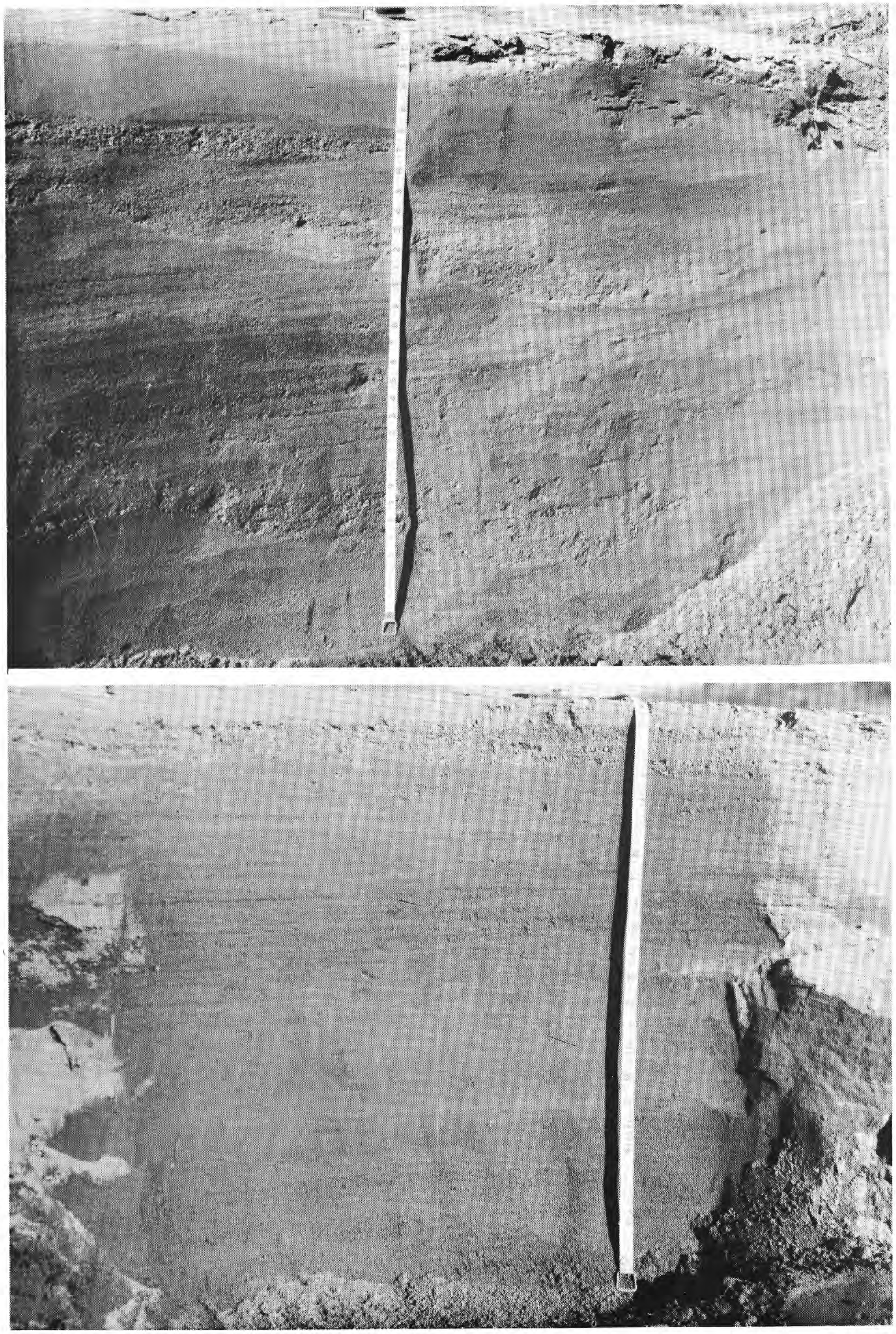


FIGURE 45.—Sandy flood-plain deposits showing laminae, pockets, lenses and coarse particles (Rucker Run about 0.75 mi upstream from highway U.S. 29). Tape is in feet and tenths, and exposed thicknesses are about 2.5–3.0 feet.

the water receded. However, they tend to occur in patches rather than covering the entire deposit, so they may or may not be present at any given spot. The ripple heights are rather low—about 0.01 or 0.02 foot. Wavelengths are only about 0.3 foot in most cases, the range being from about 0.25 to 0.5 foot. Crests are oriented approximately normal to the valley axis.

Mud cracks or shrinkage cracks also occur in patches on the surfaces of flood plains. The thickness of this mud layer can range from about 1 inch to a foot or more. The greater thicknesses tend to be found at longer distances downstream on the larger streams, for example, at the confluence of two major rivers or upstream from railroad embankments. Such deposits probably suggest considerable backwater during the flood.

The various sedimentary features listed above seem to be similar to those of floods in other geomorphic regions (Jahns, 1947; Williams 1970, 1971; McKee and others, 1967).

VERTICAL CHANGES IN GRAIN SIZE

Grain-size distributions conceivably could vary with depth at a single location, lateral distance toward the valley wall, and distance downstream. Size variations in depth at a single location could not be studied as far as large particles were concerned because one large rock fragment in many cases occupied the full thickness of the deposit. For finer grains two flood-plain localities were examined in regard to particle size-frequency characteristics with depth. This of course reflects the sediment deposition with time during the flood, at the sampling site. The localities, chosen mostly on the basis of accessibility and undisturbed vertical sections, were on Rucker Run, about 0.75 mile upstream from highway U. S. 29, and on Dillard Creek, about 1.3 miles downstream from highway U.S. 29. At the location of sampling, the deposit on Rucker Run is about 2 feet thick, that on Dillard Creek about 3 feet thick. Both locations are about the same distance downstream (4–4.5 mi) from the deposits at the heads of the valleys. A possibly important difference is that upstream from the sampling site Dillard Creek had a large highway delta, whereas Rucker Run had only a small or partial delta. The Dillard Creek highway delta trapped a very large amount of sediment, and this probably occurred before deposition at the sampling site farther downstream. Thus the grains at the Dillard locality may represent material transported later in the flood.

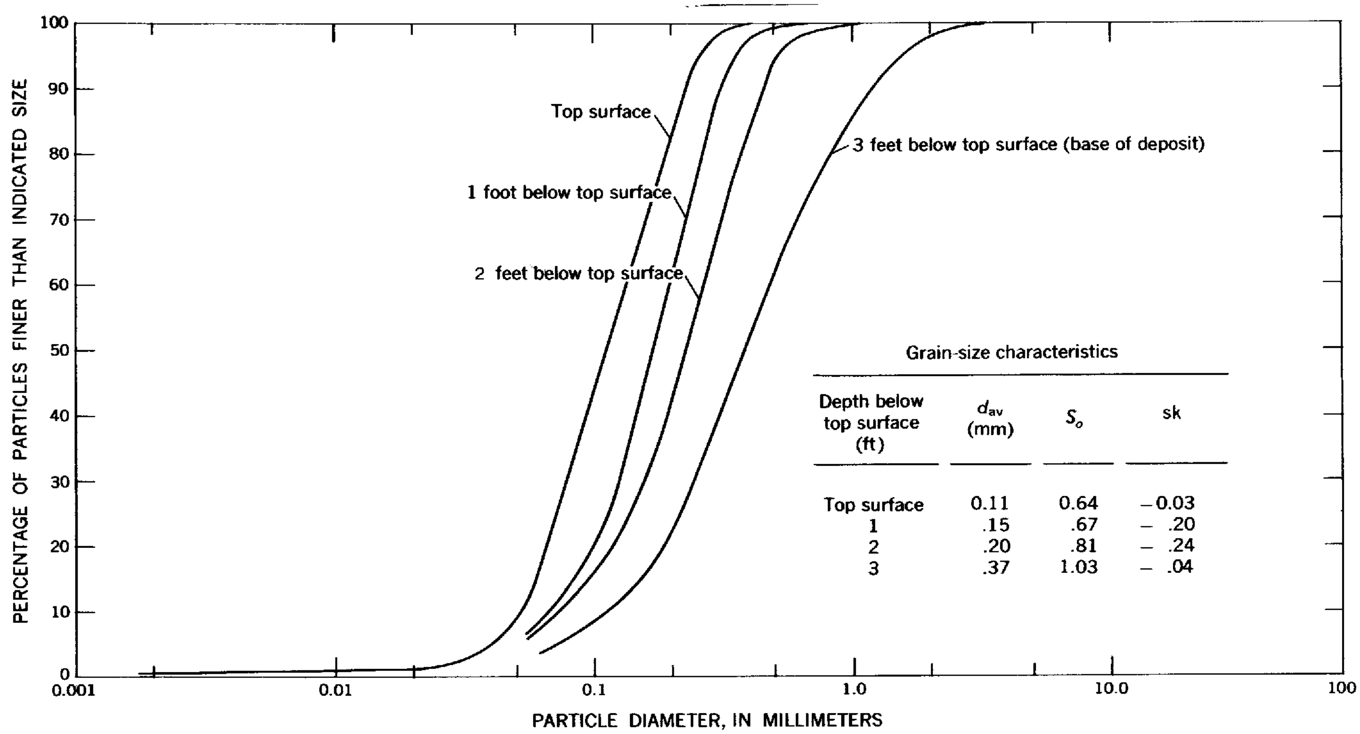
There is, unfortunately, no reliable way to determine the time of deposition relative to the peak water flow, nor is there any way of knowing the rates of sediment transport and deposition at various stages of the flood.

Nearly all the flood-plain sediments investigated for vertical changes in the grain size are sand-sized or finer. At both the Dillard and Rucker sites the top surface material was noticeably finer than that below the surface (figs. 46, 47). These fine-grained topmost sediments very likely were deposited in relatively slow-moving water near the end of the flood.

Below the top surface the deposits at the Dillard site are progressively coarser with depth. The average diameter at the base of the deposit was 0.37 mm, and this average size gradually decreases to 0.11 mm at the top surface. The sediments at this site also became better sorted upward (fig. 46). The d_{90}–d_{10} range encompasses one log cycle of grain diameter at the base of the deposit, and this range decreases to about two-thirds of a log cycle for the sediments deposited latest. The skewness shows no noticeable trend. The bottom and top deposits have approximately symmetrical distributions. The middle layers are also close to having symmetrical grain-size distribution but show a very slight tendency to include a greater weight-percent of coarse grains than fine.

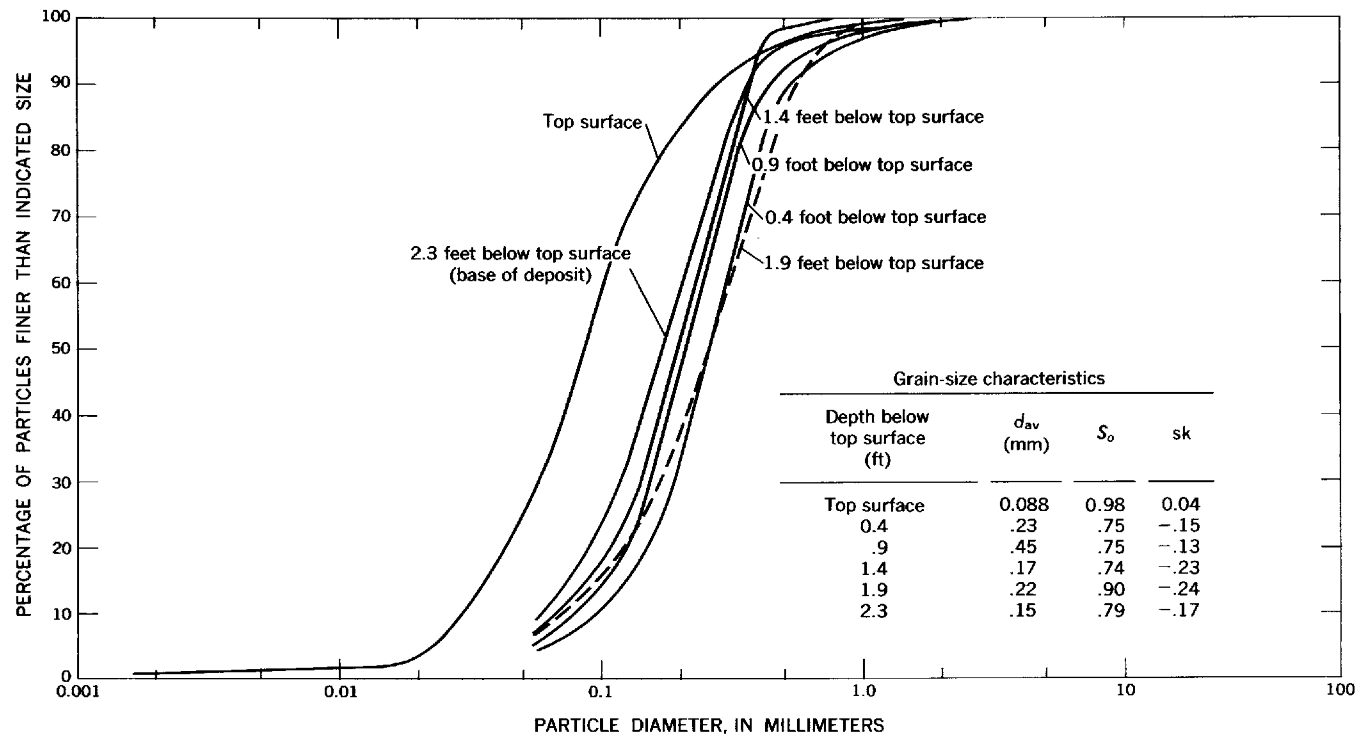
With the exception of the surface grains, the Rucker Run sediments, in contrast to those on Dillard, show no noticeable vertical trends. The average grain size of the Rucker Run sediments fluctuates between 0.15 and 0.45 mm throughout most of the deposit and becomes finer only in the upper part. Sorting stays about constant with depth. A possible exception is the top surface, which is slightly more poorly sorted than the sediments below. A more poorly sorted surface material would represent the opposite trend from the Dillard Creek site, where the topmost material was better sorted than the underlying sediments. Skewness values at Rucker Run, as with average size and sorting, did not vary widely for the deposits below the surface. The skewness range is about —0.13 to —0.24, indicating a very slight preponderance of coarser sizes. The surface material, on the other hand, has a skewness of +0.04, which means the size distribution is approximately symmetrical.

To sum up the main grain-size changes of these flood-plain deposits with depth: at both the Dillard and Rucker sites the top-surface (top inch or two) grains are noticeably finer than the material below the surface. The Dillard sediments become progressively finer upward (that is, they become finer as time progressed during the flood). Grain sizes at the Rucker site, on the other hand, do not change



Depth below top surface (ft)	d_{av} (mm)	S_o	sk
Top surface	0.11	0.64	−0.03
1	.15	.67	− .20
2	.20	.81	− .24
3	.37	1.03	− .04

Figure 46.—Size-frequency characteristics of flood-plain sediments on Dillard Creek, about 4.5 miles downstream from headwaters.



Depth below top surface (ft)	d_{av} (mm)	S_o	sk
Top surface	0.088	0.98	0.04
0.4	.23	.75	−.15
.9	.45	.75	−.13
1.4	.17	.74	−.23
1.9	.22	.90	−.24
2.3	.15	.79	−.17

Figure 47.—Size-frequency characteristics of flood-plain sediments on Rucker Run, about 4 miles downstream from headwaters.

significantly below the surface. Sorting differences at the Dillard site indicate that as the flood progressed, the deposits became better sorted. At Rucker Run, however, the sediments deposited last were a bit more poorly sorted than those laid down earlier. At both sites the distributions show a limited tendency to be skewed (minor predominance of coarser grains), except for the surface material which is distributed approximately symmetrically, on a weight basis.

The measured trends of grain-size distribution with depth are probably a product of both the sizes of particles offered to the floodwaters upstream from the study sites and the transporting ability of the stream. Thus, the deposition of finer and finer sizes as the flood progressed (Dillard locality) could mean that the larger particles were less frequently eroded from the upstream hillsides and channels with time and (or) the transporting ability of the flowing water gradually diminished over the period of deposition. Similarly, various possible combinations of particle availability and stream power could account for an absence of a distinct size-distribution trend during deposition at the Rucker Run site.

LATERAL VARIATIONS IN GRAIN SIZE

For flood-plain deposits that include large ranges in particle size the pebble-count data generally show an apparently random distribution across the width of the deposit. Sand-sized and finer grains in these deposits tend either to form the matrix for the whole deposit or to collect in scattered patches. Some of the fringe areas along the outer edges of the deposits tend to have more sand than larger stones. On the other hand, stones up to 1 meter diameter occur in a few places where no finer materials are deposited.

The sand deposit described above on Rucker Run, about 0.75 mile upstream from highway U. S. 29, was sampled to determine any lateral changes in particle size of fine-grained deposits. The study area was the right bank flood-plain deposit, which was about 60 feet wide. Single laminae or strata were either too difficult to trace laterally or too thin to sample individually. Composite vertical samples representing the entire thickness of the deposit (about 2 feet) were therefore taken at each of five sites—15, 25, 35, 45, and 55 feet, respectively, away from the present stream bank. The cumulative size-frequency curves for these five samples very nearly coincide. For a composite curve derived from all five sieve analyses $d_{av} = 0.23$ mm, $S_0 = 0.942$ and $sk = -0.153$. Comparing the individual curves to the average curve, the maximum deviations were ± 13 percent for d_{av} (± 0.03 mm), ±8 percent for S_0 and ± 34 percent for sk. The relatively minor differences between curves could be due to actual differences in the grain-size frequencies at the sampling sites, sampling errors, or splitting of samples for sieve analysis. Thus for the flood-plain deposits sampled there was no significant lateral variation in grain-size characteristics.

DOWNSTREAM CHANGES IN FLOOD-PLAIN DEPOSITS—GENERAL

Flood-plain deposits of three streams—Polly Wright, Rucker Run and Dillard Creek—were chosen to study such features as changes in amounts of deposition and in grain size with distance downstream. From one viewpoint the ideal stream for studying such changes would have no tributaries at all, and all of the deposited sediment would come from a single upstream source. Unfortunately, Nelson County appears to have no such stream, particularly since some deposited material undoubtedly came from the bed and banks along the whole length of the creeks.

The stream in Polly Wright Cove received sediment from three major tributaries near the middle of the study reach. Rucker Run received sediment from at least 13 debris avalanches or minor tributaries within the upstream 8 percent of the total distance studied. Toward the downstream end, five creeks join Rucker Run between the last two sampling stations; however, of this group only Dillard Creek contributed much sediment. Some minor creeks also contributed sediment along other parts of the Rucker study reach. Three tributaries joined Dillard Creek, at about 0.2, 0.5 and 0.8 of the way, respectively, along the 6.6-mile study zone.

The streams were less than ideal for sediment-dispersion studies, also from the standpoint of time of introduction of sediment. The exact times when sediments emerged from the mountain ravine are unknown. Sediment eroded from the banks and beds of the mountain channels and higher order streams probably was transported more or less continuously during the flood, at a rate proportional to the stream power. However, the large quantities dislodged by the debris avalanches on the hillslopes entered the channel at various times during the storm. Therefore the resulting downstream deposits in small drainage basins would reflect several stages and rates of sediment transport and deposition. Transport and deposition would logically be more uniform with increase in upstream drainage area.

The various types of source rocks on the hillsides contain no distinctive and plentiful minerals which would identify points of origin for particles in the downstream deposits.

In spite of these possible disadvantages from the viewpoint of ideal dispersion studies, the deposits that remained after the flood subsided are probably typical of those left by other catastrophic floods in similar environments. Lag deposits and material in the stream channel prior to the flood were flushed out or buried during the flood; so the sediments sampled represent only material deposited by the August 19–20 flow.

The upstream sampling stations were usually chosen at approximately equal intervals of distance, although sometimes the location was shifted slightly to sample a deposit that seemed to be more typical or that was not buried under piles of trees and other debris. For at least the downstream half of the flood-plain deposits investigated, the sampling stations were selected on the basis of accessibility, freedom from tributaries entering near the station, and a subjective evaluation of how typical or representative the deposits appeared to be. Some downstream reaches had little or no deposition, often due to a local constriction in the channel.

The valley in Polly Wright Cove is only 0.8 mile long. After emerging from Polly Wright Cove, the stream merges with another sediment-bearing stream (Muddy Creek). In reporting changes in grain size with distance downstream, we will include the mountain-channel deposits in Polly Wright Cove. These deposits were continuous with the downstream flood-plain deposits.

Within a week or two after the flood, the streams had returned to their usual flow widths and depths, these commonly being 1–20 feet wide and 0.1–0.6 foot deep. These creek widths were negligible or minor compared with the widths of the flood deposits. Nearly all the sampling was done on dry ground.

DOWNSTREAM CHANGES IN AMOUNT OF DEPOSITION POLLY WRIGHT COVE

At 11 stations along Polly Wright Cove measurements were made of the volume of the flood-plain deposits, using the same techniques as on alluvial fans. These data, expressed in volume of sediment per unit (foot) of channel length, are shown in figure 48*A*. The downstream distance, x, was reckoned from the first major deposits, excluding debris piles and other scattered deposits in mountain ravines. The volume of sediment decreases downstreamward, although the scatter on the graph hinders determination of the precise relation. In an effort to reduce the scatter, we combined the data for the 11 individual

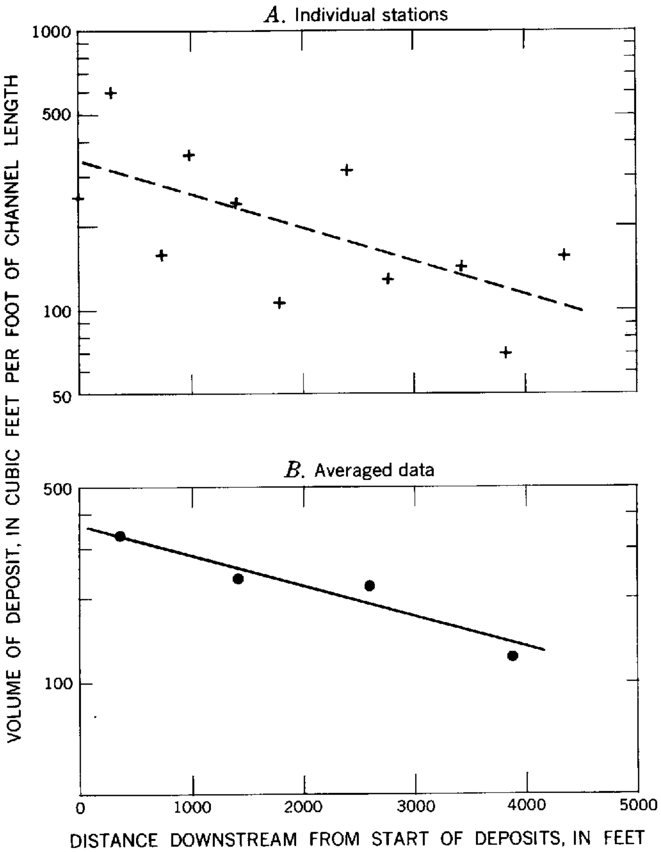


FIGURE 48.—Volume of flood-plain deposition per foot of channel length along Polly Wright Cove.

sections into four averages, consisting of the upstream three measuring stations, the second three, the third two and the downstream three. With the data combined into reaches about 1,000 feet long, the scatter diminished considerably. Figure 48*B*, a plot of these results, shows that the volume of deposition decreases exponentially with distance downstream. According to the eye-drawn line, the volume of deposition per foot of channel length at any downstream site is equal to $370e^{-0.00025x}$, where e is the base of natural logarithms. The deviations of data points from the regression line range from +15 percent to —12 percent.

The actual quantity deposited ranged from about 370 cu ft per ft at the upstream end to about 120 cu ft per ft at the downstream end of the 0.8 mile-long deposit.

The total volume of sediment dumped in Polly Wright Cove was about 0.97 million cu ft. An estimated 3.1 million cu ft was eroded from the headwaters (table 6). Sediment not accounted for in the flood-plain deposits must have been transported completely out of the cove. Extrapolating the exponential equation just derived to a distance beyond the cove is both risky and intriguing. If the equation were applicable for as far as appreciable deposition continued, the total amount deposited would be 370/0.00025, or about 1.48 million cu ft, somewhat less than the 3.1 million cu ft estimated to be eroded. Nearly all of the 1.48 million cu ft, according to the equation, would be deposited within three miles of the upper end of the deposits. If the measurements and the resulting equation for the 0.8-mile-long study reach are accurate, either the equation should not be extrapolated or the estimated 3.1 million cu ft of erosion is too high.

Table 9 gives the volume of deposition by particle-size categories and the size distribution by categories, for the four separate reaches along the 0.8-mile-long cove. If the entire flood-plain deposit along the length of the cove is considered, sand is by far the most abundant size class. The cove received more than twice as much sand (449,000 cu ft) as any other size group. The other outstanding feature is the scarcity of silt and clay, this group contributing only about 51,000 cu ft, or 5 percent of the deposition. The silt and clay, if eroded in substantial amounts, could have been carried away as suspended load. Most particles larger than sand, if eroded from the headwater regions, probably would have come to rest somewhere within the cove rather than moving through the cove to some downstream site; the volumes and percentages of these larger particles—gravel, cobbles and boulders—dwindle noticeably with distance along the cove, as table 9 shows. It therefore appears that a much greater volume of sand was eroded in the headwater regions, as compared to larger particles.

In the reach farthest upstream, sand (31 percent of the deposit at this site, or 90,000 cu ft) and boulders (29 percent or 85,000 cu ft) are the most abundant size classes. Boulders, if eroded from the headwater regions in appreciable quantities, should be abundant in this first reach because they would logically be deposited farther upstream than smaller particles. In the next reach (about one-third of the way along the cove's flood-plain deposit) sand and gravel are the most abundant groups, especially sand. The volume and percentage of cobbles and boulders are less than in the upstream reach, and the amount of silt-clay is about the same. Sand becomes even more plentiful in the third reach and amounts to 138,000 cu ft or 60 percent of the deposit. The quantity and percentage of silt-clay are about the same as in the two upstream reaches. Gravel is still more plentiful than cobbles and boulders, but these three categories, especially the latter two, are scarcer than in the previous reach. Finally, in the reach farthest downstream, sand comprises nearly three-fourths (111,000 cu ft) of the total deposit. Gravel is next in importance (18 percent), while silt-clay, cobbles and boulders total only 10 percent (16,000 cu ft) of the deposit.

In summary, the volumes and percentages of cobbles and boulders decrease steadily with distance downstream and are practically negligible at the end of the 0.8-mile study zone. Gravel everywhere makes up about 16–25 percent of the sediment and is more abundant about one-third of the way along the cove than elsewhere. Sand consistently becomes more important with distance, progressing from 31 percent to 72 percent of the deposit. Curiously, the percentage of silt-clay remains virtually constant at about 5 percent along the entire cove.

Later sections of this report examine other aspects of grain-size trends with distance.

RUCKER RUN

Figure 49 shows how the volume of flood-plain deposition varied with distance for the upstream 12.5 miles of Rucker Run. The somewhat erratic nature of the deposit volume for the upper two or three miles is due mainly to the addition of sediment from small tributaries. The point labeled "partial highway delta at U.S. 29" illustrates the amount of sediment trapped by the highway. More sediment probably would have accumulated here, that is, a complete highway delta would have formed, if a section of the highway had not washed out during the flood. The volume of deposition at any given site down to highway U.S. 29, about the first 4 miles, is

TABLE 9.—*Volume and percentage of deposition by size classes with distance downstream in Polly Wright Cove*

Proportional distance downstream to midpoint of reach	Length of reach (ft)	Volume of deposition in reach (1,000 ft³)	Volume by size classes (1,000 ft³)					Percentages in class				
			Silt-clay	Sand	Gravel	Cobble	Boulder	Silt-clay	Sand	Gravel	Cobble	Boulder
0.08	875	292	15	90	47	55	85	5	31	16	19	29
.32	1,233	288	14	110	72	40	52	5	38	25	14	18
.59	998	231	14	138	46	12	21	6	60	20	5	9
.89	1,254	155	8	111	28	5	3	5	72	18	3	2
Total	4,360	966	51	449	193	112	161					

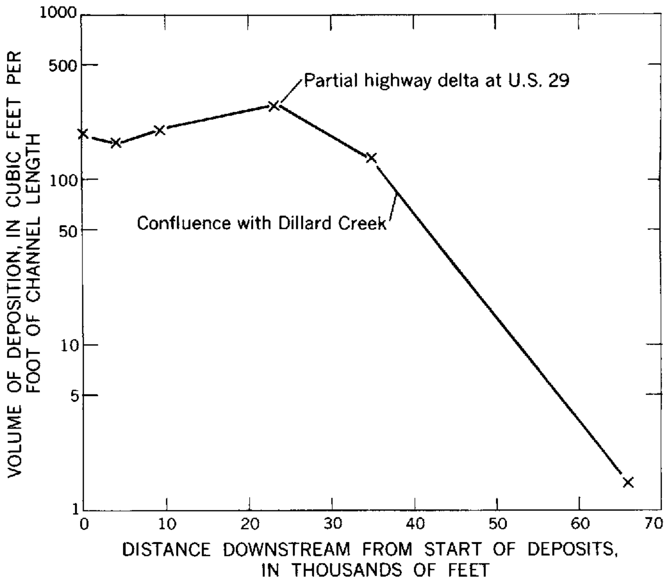


FIGURE 49.—Changes in volume of flood-plain deposition with distance downstream on Rucker Run.

in the approximate range of 150–300 cu ft per foot of channel length.

Downstream from the partial trap at the highway the amount of deposition dwindled rapidly. At the confluence with Dillard Creek (approximately the 7-mile station) the volume of deposition was about 90 cu ft, and at 12.5 miles downstream from the head of the cove the flood plain contained only 1 or 2 cu ft of sediment, a very thin layer.

The first 7 miles of the Rucker Run flood plain received a total of about 7.8 million cu ft of sediment, according to figure 49.

Some of the silt and clay, and possibly sand too, moved in suspension into the James River. Towns and flood plains along the James, both upstream and downstream from the junctions with tributaries draining Nelson County, were covered with fine-grained sediment to depths up to a foot. The source of this material could be any place from the upper reaches of avalanche scars in the headwaters to the banks of the James River itself. Downstream from Richmond the greatly increased concentration of suspended load in the James River estuary reached a maximum around Aug. 25, about 3–4 days after the peak water discharge at Richmond (Maynard Nichols, written commun., 1969).

One effect of the flood was to erode the mountain hillslopes and aggrade the upstream parts of the valleys. If rare catastrophic events produce most of the erosion and deposition over a long period, the results of this storm suggest a trend toward a peneplain, due to net hillside erosion and valley deposition. On the other hand, the storm may be a temporary deviation in a long-term trend which may in time be established by more moderate erosional and depositional processes.

Again, if this type of catastrophic event is dominant, the Wills, Polly Wright and Freshwater Cove flood plains, as well as certain others, are being constructed primarily by overbank deposition. They would thus represent an exception to the mechanisms which Wolman and Leopold (1957) proposed, whereby about 80–90 percent of a flood-plain thickness is build by point-bar (lateral) accretion within a migrating stream channel. In the present case the crucial factor in determining the amount of overbank deposition seems to be the quantity of sediment introduced at the head of the valley. Where the mountains underwent extensive debris avalanching, the heads of the valleys received considerable overbank deposition. This vertical accretion ranged from an average of about 3 feet at the upstream ends of the coves to 0.1 foot or less at a downstream distance of 1–8 miles. On the other hand, where few debris-avalanches occurred in the mountains, as in the Indian Creek basin, the downstream-valley flood plains received noticeably less sediment and may in fact have undergone a net sediment loss due to lateral erosion of the stream channel.

DOWNSTREAM CHANGES IN AVERAGE GRAIN SIZE

Most authors have found that average grain size generally decreases downstreamward (Pettijohn, 1957), although exceptions have been reported (McPherson and Rannie, 1969).

Figure 50 is a graph of the mean particle size d_{av} versus distance down-valley for Rucker Run, Polly Polly Wright Cove and Dillard Creek. as before, x is distance downstream from start of deposition. In general, d_{av} decreased with distance downstream. The distribution of points on at least two of the three plots is such that various lines can be fitted. For example, either the solid or dashed lines can represent the relation for the Polly Wright Cove and Dillard Creek sediments. The solid line for all three streams suggests that given enough travel distance, d_{av} may become asymptotic toward some limiting minimum value. The Rucker Run data are for a relatively long study reach (12.5 mi) and tend to support this type of relation. The dashed lines for Polly Wright Cove and Dillard Creek represent negative exponential functions, a type of relation rather common in geomorphology and sedimentology.

The average particle size at the upper end of the deposits was 1.6 mm for Rucker Run, 14.4 mm in

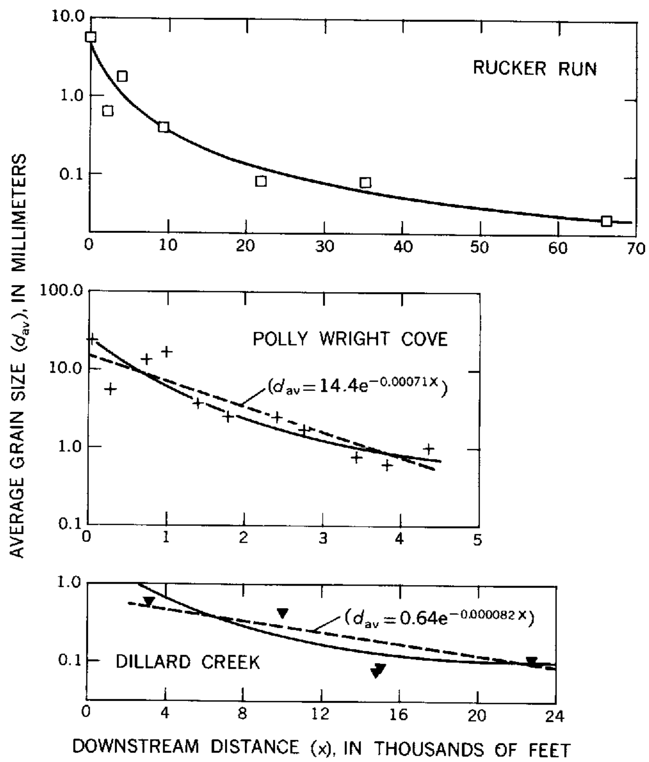


FIGURE 50.—Downstream changes in mean particle size (d_{av}) of flood-plain deposits.

Polly Wright Cove and 0.64 mm (extrapolated) for Dillard Creek. This value depends on at least two factors: the sizes of particles dislodged from the hillsides and the sizes and numbers of particles transported to the flood plains. Lithology and erosional history undoubtedly influence the sizes of the rocks dislodged from hillsides. Also, many (but not all) large rocks were retained in mountain ravines or alluvial fans and therefore were not included in the flood plain analysis. This, however, was not the case in Polly Wright Cove, where the deposits were virtually continuous over the entire study reach and where no alluvial fan formed. The relatively large value of d_{av} (14.4 mm) at $x=0$ ft in Polly Wright Cove reflects this condition.

Why does d_{av} on Polly Wright decrease nearly 10 times as fast as on Rucker and Dillard? The answer seems to lie in the sizes of particles introduced at the head of the study reach and in the effect of several side tributaries which supplied more large rocks downstream. The upstream Polly Wright deposits were closer to the steep mountain hillslopes which provided the debris. This fact, plus the absence of an alluvial fan, means that the upstream deposits in Polly Wright Cove contained a greater proportion of large boulders, thus contributing to a larger average particle size. As mentioned elsewhere, large rocks were not moved very far from the mountainous areas during the flood (usually considerably less than a mile). Thus the average particle size in Polly Wright Cove decreased rather rapidly with distance. On Rucker Run many slides and tributaries furnished large rocks to the flood-plain deposits along the study reach, so d_{av} did not decrease as rapidly as if large rocks had been introduced at a single point source at the head of the reach. In the Dillard Creek basin some deposition in mountain ravines and in a fan trapped most of the large rocks before deposition of the flood-plain sediments. Hence the possible range of variation in d_{av} along the Dillard reach was much more restricted.

A decrease in grain size can result from selective deposition, breakage of particles, and abrasion or wearing down of the grain surface. Any of these can be influenced by such factors as flow properties (discharge, mean velocity, depth, width), channel properties (gradient, roughness, sinuosity), particle characteristics (size, shape, specific gravity, mineralogy), and erosional history or degree of weathering. For two streams in the Black Hills area of South Dakota, Plumley (1948) concluded that selective deposition accounted for 75 percent and 84 percent of observed grain-size reductions, respectively, with abrasion accounting for the remaining small percentage. Scott and Gravlee (1968) reported that for a flood surge on the Rubicon River in California selective deposition caused more than 90 percent of the measured size decline and that breakage was not significant in the overall size reduction. Bradley, Fahnestock, and Rowekamp (1972) concluded that sorting processes caused 87 percent of an observed reduction in the size of flood-transported gravel along a 16-mile reach of the Knik River, Alaska. The remaining small percentage in size reduction was attributed to abrasion.

Preflood weathering which could promote breaking during transport could be a minor factor with the present sediments; however, on the basis of field observations, most particles, at least those visible on the surface and in cuts eroded through the fresh deposits, seemed to be relatively fresh. Breakage, however, could have occurred during transport, especially with the larger rocks. Aside from the possibility that some of the measured size decline could be due to breakage, there is no firm evidence to evaluate the extent to which breakage during transport was important. Intuitively it would seem to be a minor factor. Abrasion or wearing down of the particle surface probably could not have been significant, because the rock (igneous and metamorphic) generally was hard and fresh, and the distance of travel was quite short, ranging from a few feet to about 12.5 miles for the data obtained here. Thus the most likely reason for the progressive decrease in size is selective deposition, with breakage possibly affecting the trend to a minor extent.

AVERAGE PARTICLE SIZE, LOCAL SLOPE, AND DRAINAGE AREA

What factors might be most influential in determining the particle sizes that are left at a given site? The two which appear most reasonable are the local gradient and water discharge. For Rucker Run and Dillard Creek local gradients were measured from topographic maps (1:24,000). These slopes equal the vertical distance on the maps between the contour line upstream from the sampling site and that downstream from the site, divided by the horizontal distance along the channel between the two contour lines. All slopes along Polly Wright Creek were measured in the field with a Zeiss level and stadia rod.

Average particle size shows some correlation with local flood-plain slope S (fig. 51). The line fitted by least-squares in figure 51 has the relation $d_{av}=2{,}270\ S^{1.98}$. Thus the average particle size decreased as the slope became flatter. The deviations of data points from the regression line range from +600 percent to —50 percent.

Hack (1957, p. 57) studied Virginia and Maryland streams near this area and included a much greater variety of geologic environments and drainage-basin sizes. He found that with decrease in slope the size of the bed material could increase, decrease, or remain constant, depending on the surface lithology of the drainage basin. Also, the sediments Hack studied had been extensively reworked by normal streamflows. (In fact, he felt that the sizes of bed-material particles partly determined the local slope, whereas the reverse probably was true for the Hurricane Camille deposition.) Plumley (1948) found power relations between median grain size and local slope for ancient stream sediments along three streams in the Black Hills of South Dakota. Plumley's three exponents, with slope the independent variable, were 0.53, 0.53 and 1.25, and d_{av} decreased at a lesser rate as local slope decreased, as compared with relations in Nelson County.

Water discharge, as mentioned, was not measured. However, the drainage area usually shows a power relation to water discharge. Consequently, drainage area, as measured from topographic maps (1:24,000)

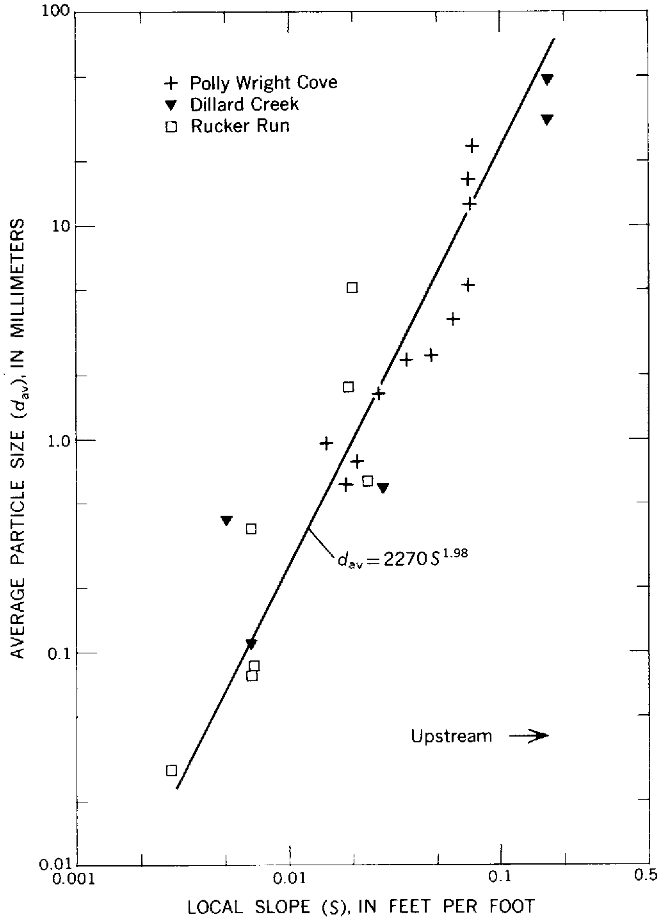


FIGURE 51.—Change in average particle size with local flood-plain slope. The two points at steepest slopes for Dillard Creek are alluvial fan sediments.

with a planimeter, was plotted against average grain size. The drainage area for any sampling site includes the drainage area of all upstream tributaries which enter the major stream.

Figure 52 indicates that with increase in drainage area A_d, the average particle size decreased according to the least squares relation $d_{av}=1.9A_d^{-1.34}$. This equation pertains to all of the plotted points as a group. Deviations of measured data from the regression line range from +300 percent to —26 percent. Individual drainage basins may not follow the general equation. In Polly Wright Cove, for example, the average grain size decreases with the —2.8 power (approximately) of drainage area, as suggested by the dashed line in the diagram.

Since average particle size increases with the 1.98 power of local slope S and decreases with the 1.34 power of drainage area A_d, these relations can be rearranged and combined into the form $S \propto (d_{av}^{0.25})/A_d^{0.34})$. Writing the relation this way implies that slope is the dependent variable, probably not the case

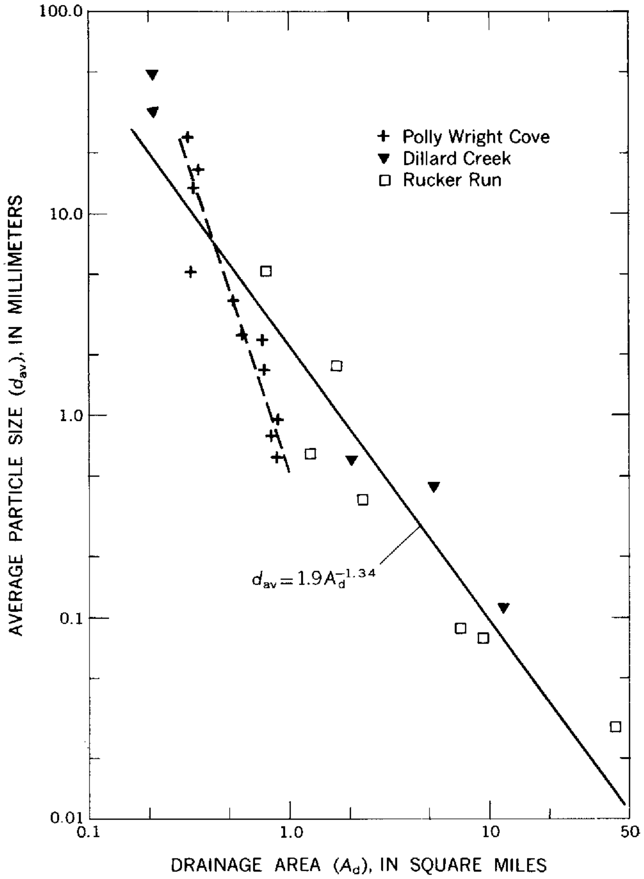


FIGURE 52.—Variation of average particle size with drainage area. Dillard Creek points include two alluvial fan stations (coarsest values of d_{av}) and do not include highway delta deposits.

in the present study, but the equation in this form can be compared to Hack's (1957, p. 58) relation $S \propto (d_{av}^{0.6})/(A_d^{0.6})$. The differences in the exponents are not very big, considering the amount of scatter on the plots in both investigations. Hack's equation applies to streams in various rock types.

It will be interesting to see if future research finds empirical relations of the sort discussed here to be applicable to other physiographic regions. Denny (1965) studied desert washes in California and Nevada and could not find any correlation at all between local slope, median particle size, and drainage area.

DOWNSTREAM VARIATION IN SIZE OF LARGEST STONES

The largest stones at a site, if all sizes were available for transport, provide some indication of the particle sizes the flow was able to move. Figure 53 is a plot of the average size (intermediate diameter)

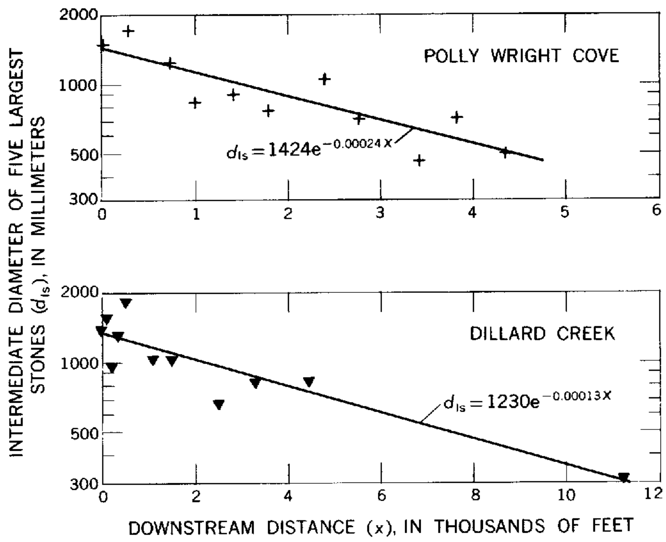


FIGURE 53.—Downstream changes in average intermediate diameter of the five largest stones at each station. Lines fitted by least squares.

of the five largest stones (d_{ls}) as a function of distance downstream for Polly Wright and Dillard Creeks. The Dillard Creek data include the Wills partial fan at the head of the cove (five data points), and the downstream distances begin at the fan apex for this particular diagram. Only three pebble counts—a number insufficient for plotting—were taken on Rucker Run.

A negative exponential function, fitted here by least squares, describes the data reasonably well. For Polly Wright d_{ls}=1,420$e^{-0.00024x}$, with maximum deviations ranging from +135 percent to —75 percent of the regression value. For Dillard Creek d_{ls}= 1,330$e^{-0.00013x}$, and maximum deviations range from +145 percent to —70 percent of the regression value. The Polly Wright exponent of —0.00024 means that for every 1,000 feet of distance the geometric mean of the largest stones decreases by 24 percent. Similarly, along Dillard Creek the size at each 1,000-foot station was about 13 percent less than that at the previous 1,000-foot station. The exponents show that the size of the largest stones decreases nearly twice as rapidly along Polly Wright Cove as on Dillard Creek. (Among other differences, the flood-plain slope decreases more rapidly in Polly Wright Cove, too.)

As a general trend the largest rock fragments along Polly Wright Cove decreased from 142 cm in the deposits farthest upstream to about 50 cm after 0.8 mile of travel. In Dillard Creek the average diameter of the five largest stones was 133 cm at the upstream end of the creek and about 30 cm a little more than 2 miles downstream.

The best-fit lines should not be interpreted as suggesting that at any given downstream location the average intermediate diameter of the five largest stones would necessarily be the diameter indicated by the line. At some intermediate locations no particles larger than pebbles or coarse sand appeared on the surface. Considerations governing the selection of sampling sites were mentioned earlier in the report.

DOWNSTREAM CHANGES IN SORTING

The flood-plain deposits generally became better sorted with distance down-valley (fig. 54). Again an asymptotic relation seems to be the best fit to the Rucker Run points, as is true for Dillard Creek. Most of the boulders came to rest within a mile or so of the upstream deposit so that the sorting has the largest values and improves most rapidly in this zone. Sorting on these creeks continues to improve with distance downstream but at a progressively lesser rate. The sorting can only become zero when all grains between d_{90} and d_{10} are the same size, a situation extremely unlikely to occur. Thus the sorting probably should not continue to improve at a constant rate with distance down valley; instead, a more reasonable relation is a change by lesser amounts as the possible range of improvement decreases, that is, S_o should become asymptotic toward some limiting minimum value.

A straight line fits the data points for Polly Wright Cove, indicating a negative exponential relation for the study reach. As explained above, this relation probably could not continue indefinitely with distance. For example, extrapolating the straight line suggests that about two miles down from the upper end of the study reach, S_o would have the unlikely value of zero, that is, all grains from d_{90} to d_{10} on the cumulative size frequency curve would be the same size. In fact, the data for the 4,500-foot reach of Polly Wright Cove fit reasonably well onto the points for Rucker Run and Dillard Creek, as shown in the bottom part of figure 54. This suggests that the S_o relation may be steep and linear (exponential) within the first mile of emergence from the mountainous area and then takes on a curve approaching a limiting value some distance downstream.

In addition to a downstream limiting minimum value, the sorting probably had a limiting maximum value in the upstream deposits, due to the finite sizes of boulders available for movement and movable by the floodwaters. This value seems to be about 4, that is, the range log d_{90}–log d_{10} for the poorest sorted deposits did not encompass more than about 4 log cycles of grain size. This range usually extended from about 0.1 or 0.2 mm to about 1,000 or 2,000 mm. Sorting values higher than 4 could have been restricted by a relative scarcity of larger boulders available for transport or an inability of the flow to move more than a few such boulders from the upstream areas to the valley flood-plain deposits. Inspection of the stream channel upstream from depositional areas revealed only a few really huge boulders; so not many boulders larger than about two meters in intermediate diameter were available. A few boulders of such dimensions were moved by the flood, for example in Ginseng Hollow, but probably only for short distances along the mountain channel.

These probable limitations on sorting show up more clearly in a plot of S_o versus d_{av} (fig. 55). In the upstream reaches d_{av} is large, in spite of the ubiquitous sand and silt, because of the presence of the large rocks. The associated large value of sorting tends to remain high, as d_{av} decreases from about 55 mm to 4 mm. With further decreases in d_{av}, that is, proceeding farther downstreamward, sorting improves rapidly as the larger rocks become scarcer. But in the downstream reaches, while d_{av} continues to decrease, S_o begins to level off, that is, the range of sizes present in the deposits tends to become constant. This suggests that in the downstream regions the flow could easily transport all the available grain sizes.

DOWNSTREAM CHANGES IN SKEWNESS

Along Dillard Creek and much of Rucker Run the skewnesses are close to zero (fig. 56), indicating approximately symmetrical distributions. Rucker Run actually may show a very slight trend over the full 12.5-mile study reach whereby the size distributions go from slightly positive (abundance of small grains) at the upstream end to slightly negative (preponderance of the larger grains within the distribution) at the downstream end; however, it is difficult to say how much of this possible trend is significant and how much could be due to chance sampling. In Polly Wright Cove the deposits in the upstream 1,000 feet had a preponderance of large rocks (negative skewnesses), whereas the remaining section of the study reach had slightly positive skewnesses. Thus in these downstream areas of Polly Wright Cove the smaller grains were more predominant than the large rocks in terms of weight-percent.

The skewness varies with the average grain size, for the flood-plain and alluvial fan deposits, in a

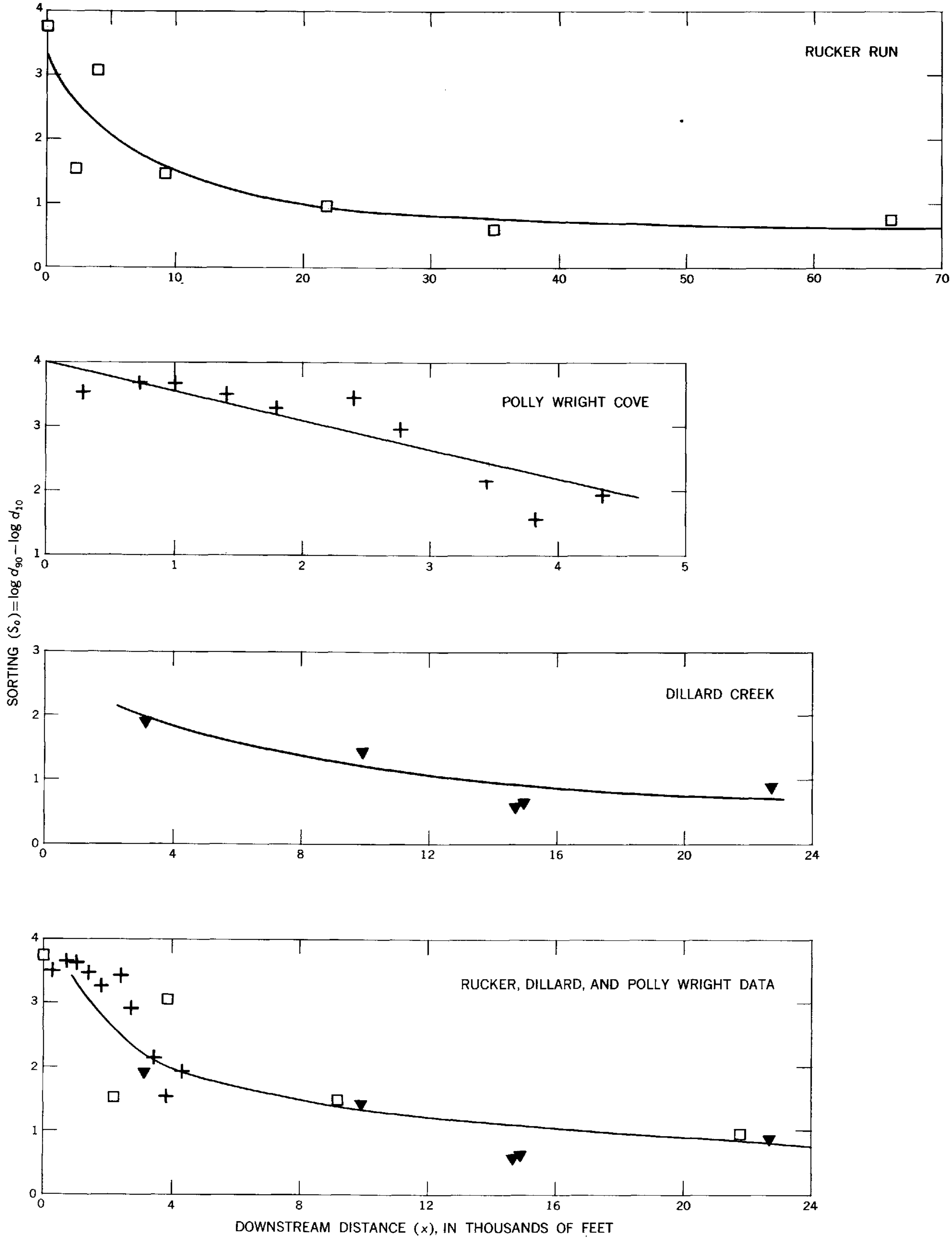


FIGURE 54.—Downstream changes in sorting of flood-plain deposits.

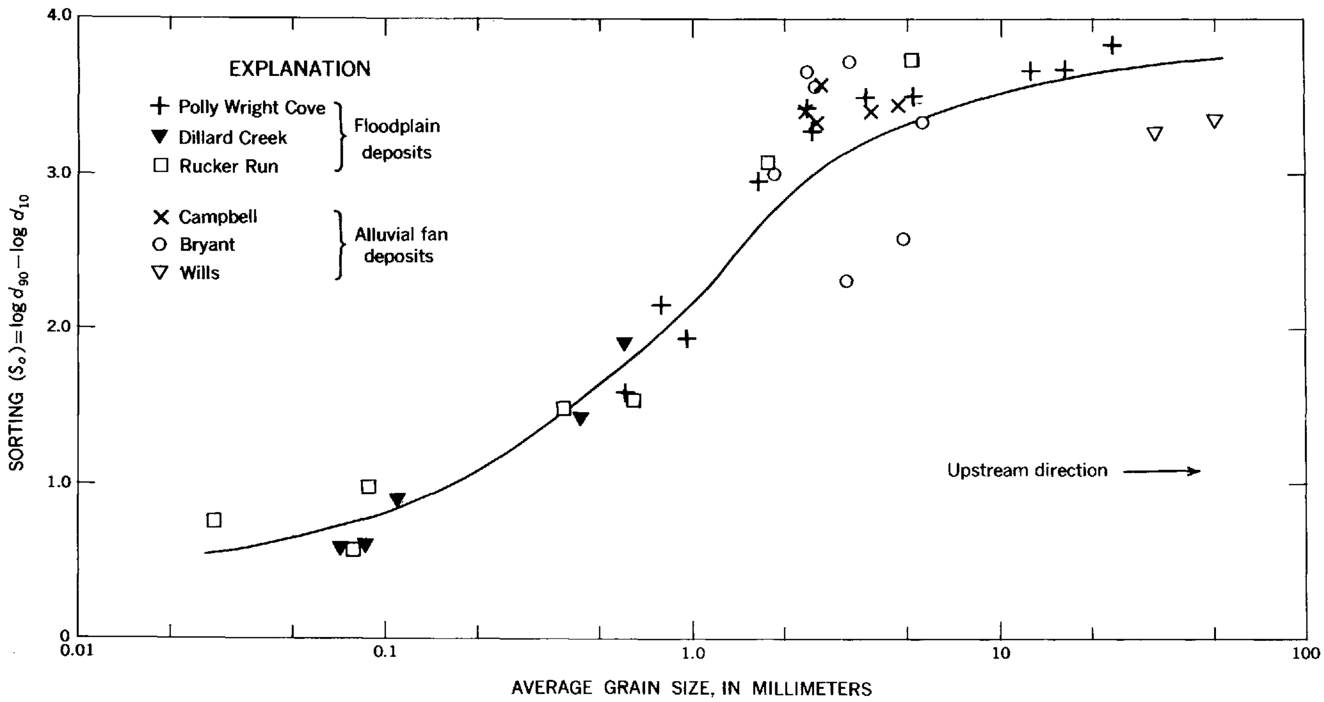


FIGURE 55.—Change in sorting with average grain size.

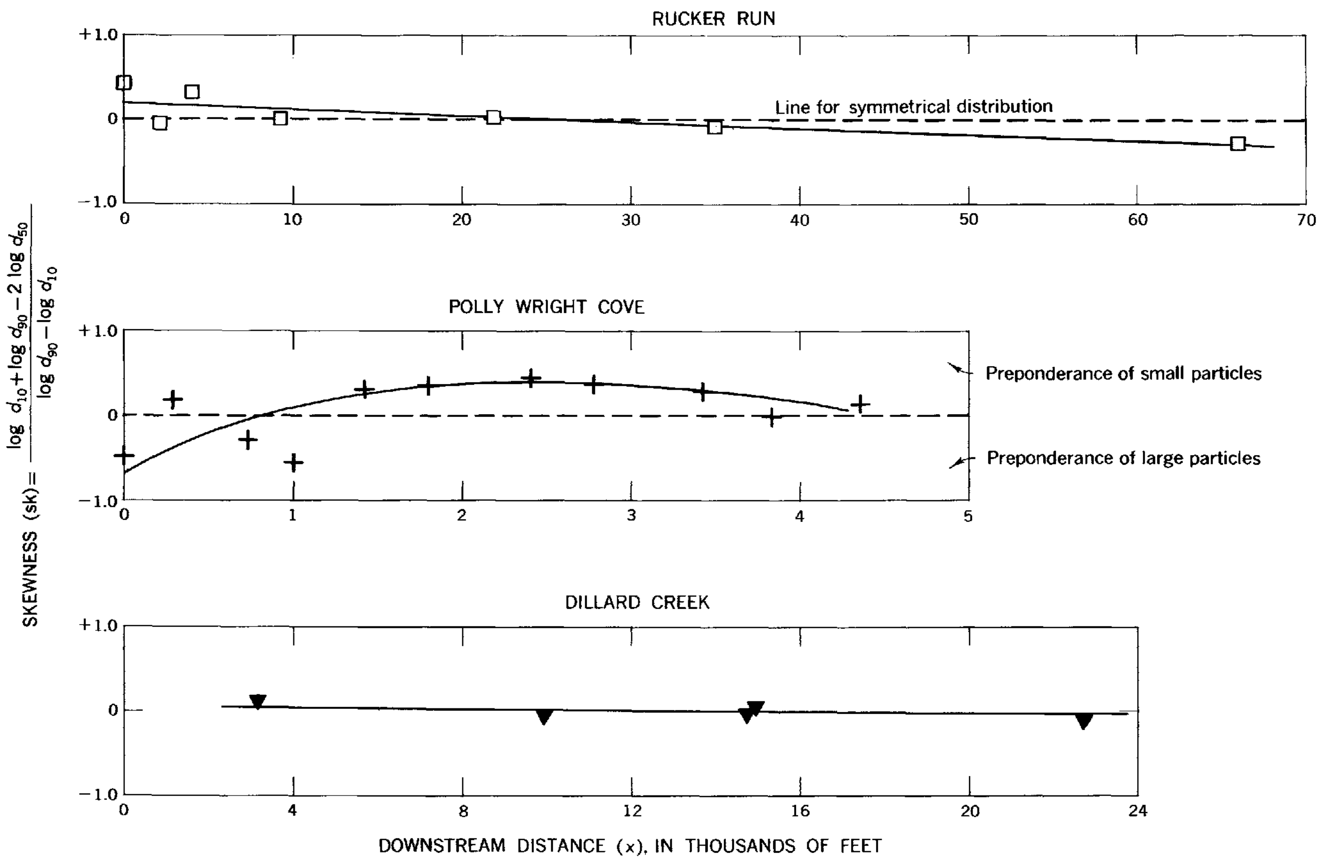


FIGURE 56.—Downstream changes in skewness of flood-plain deposits.

rather peculiar way (fig. 57). This graph should be interpreted as showing a relation between d_{av} and skewness, rather than as an implication that a change in d_{av} causes a change in skewness. Distributions that have a large average grain size are negatively skewed, meaning the large boulders predominate as far as weight frequency is concerned. As d_{av} decreases to about 8 mm the distributions gradually become more and more symmetrical. With a further decrease in d_{av}—down to about 2 or 3 mm—the distributions become more and more skewed in the opposite direction, that is, gain a preponderance of smaller grains. The reason for this skewness trend is the progressively smaller percentages of large stones contained in the deposits. Average size gets smaller, in other words, and the larger particles within each sample become progressively less common. As d_{av} decreases from about 2 mm, the sorting improves rapidly (fig. 55), as the large stones begin to disappear from the deposits. This means a trend back toward more symmetrical distributions, as shown in figure 57. Approximately symmetrical distributions are reached when d_{av} has decreased to about 0.5 mm. For smaller average grain sizes, the distributions tend to be approximately symmetrical or slightly negatively skewed.

SCARCITY OF COARSE SAND AND GRAVEL

Geologists have long speculated on the curious fact that many grain-size analyses show a dearth of particles in the coarse sand or fine gravel ranges. This supposed scarcity of certain sizes causes the histograms or size-frequency curves to show more than one peak or mode, as long as the range of sizes represented goes at least from medium sand to coarse gravel. Pettijohn (1957, p. 44) gives a good review of cases where such polymodality has been reported in the literature. Russell (1968) goes into the subject in some detail and concludes that grains of about 1–6 mm diameters are deficient in fluvial deposits because such grains are more readily entrained and more rapidly transported than larger or smaller particles. The "missing" grain sizes, in his view, are moved downstream beyond river mouths and tend to be concentrated on beaches. The following paragraphs summarize the Nelson County flood deposits in regard to polymodality of size-frequency distributions.

The first problem was how to define polymodality. For example, if a class had a decrease of only one or two percent below the percentages in the adjoining size classes, it seemed unrealistic to call such a slight decrease a deficiency, because the minor differences could easily be due to sampling or splitting (for sieve analysis) errors. Then the question was how much of a difference in weight percentage to require in order to label the distribution polymodal.

To resolve this problem we measured the approximate error that could be attributed to laboratory splitting of the field sample. This was done by sieving the entire contents (1–4 lb) of 10 of the samples brought from the field (called herein the field sam-

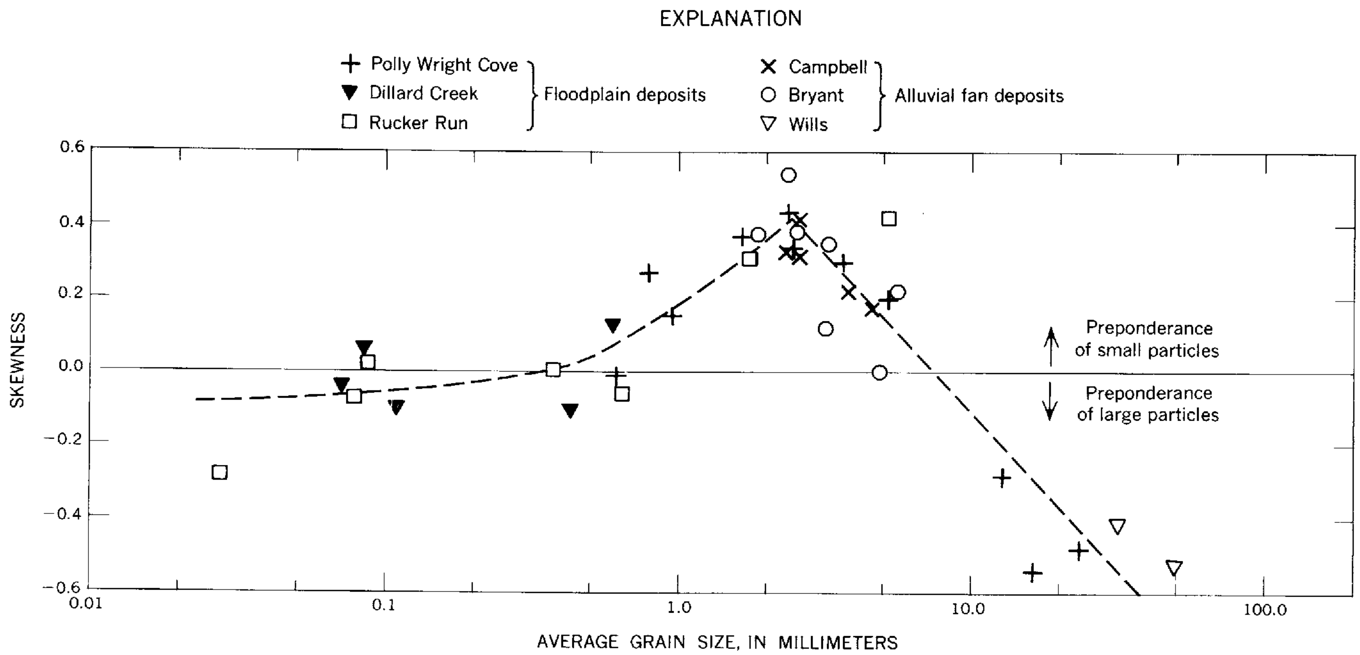


FIGURE 57.—Change in skewness with average grain size.

ple) and comparing the results to the corresponding split samples. The field samples were split 1–3 times in the laboratory, to obtain the split sample. The particular size classes of interest in the test were those which were usually present in the field and which were often involved in the apparent scarcity: 32–16, 16–8, 8–4 and 4–2 mm. For these four size classes both the full field sample and the split sample were sieved, and the percent by weight in the total field sample was compared to that in the split sample, for a given size class. The largest discrepancy or difference was 5.3 percent, that is, one of the field samples had 5.3 percent by weight in one of the size classes, whereas for the same class the split sample had zero percent by weight. General agreement was quite good, however, with the arithmetic average discrepancy between full field sample and split sample amounting to 1.5 percent. Based on the data from these sieve tests, we estimated the standard deviation of the discrepancies or differences between any two size classes of a given sample, using certain simplifying assumptions. Two times this standard deviation was 3.05 percent; that is, the chances were about 95 out of 100 that within a size-frequency distribution a difference of 3 percent or less between any two size classes could be attributed to splitting error. The criterion for polymodality therefore was that at least one size class have at least 3 percent by weight less than a coarser and finer class.

There were 63 grain-size analyses, involving two flood-plain sites and four highway-delta sites, which covered the complete range of sizes at a sampling station. Of these analyses 27 dealt with a study of grain size changes with depth. All such samples were medium sand or finer and were analyzed by sieving, occasionally including also a pipet treatment. None of these 27 analyses was polymodal.

The remaining 36 analyses were surface sediments which usually involved a wide range of particle sizes, thus requiring in most cases a pebble count in addition to the sieve analysis. Of these 36 cases, 24 showed a definite polymodality.

Table 10 shows the frequency with which each size class was deficient, according to the 3 percent rule described earlier. Particles which were especially scarce were those in the size range from 4 to 64 mm.

The polymodal samples have certain common characteristics which are generally absent in the unimodal distributions. As mentioned above, polymodality showed up markedly in surface samples but did not appear in sediments below the surface.

TABLE 10.—*Number of cases of deficiency in weight-percent, for the indicated particle size class*

[Classes not listed had no instances of deficiency. Deficiency is at least 3 percent by weight less than a finer and a coarser size class, within same sediment sample]

Stream or study area	Size class, in millimeters										
	1,024–512	512–256	256–128	128–64	64–32	32–16	16–8	8–4	4–2	2–1	1–0.5
Polly Wright (11 stations, 8 polymodal).	-	1	-	2	7	5	6	6	3	2	2
Dillard Creek (7 stations, 2 polymodal).	1	-	-	1	2	1	2	2	-	-	-
Rucker Run (6 stations, 2 polymodal).	-	-	1	1	1	2	1	1	-	-	-
Campbell Fan (5 stations, all polymodal).	-	-	-	1	5	5	5	4	4	4	1
Bryant Fan (7 stations, all polymodal).	-	-	2	3	2	7	5	5	2	2	1
Sums: 36 stations 24, having polymodal distributions	1	1	3	8	17	20	19	18	9	8	4

Deposits such as alluvial fans which had significant proportions of boulders and cobblestones consistently showed polymodal size distributions. Also, the presence of such large rocks meant that the final distribution represented a combined pebble count and sieve analysis. (All 24 cases of polymodality were combined pebble-count-sieve analysis distributions; an additional five such combined analyses did not show polymodality.)

The scarcity of grains in the 4–64 mm range may be either real or apparent. Apparent in this sense means the scarce grains actually existed on the surface of the deposit but were somehow missed in the analysis. The possible ways they could have been missed are: (1) failure to be tallied in the pebble count; (2) failure to be included in the scooping up of the "sand" sample; (3) erroneous loss in percent by weight due to the mathematical analysis of the data, that is, resulting from the method of combining the pebble count with the sieve analysis.

The pertinent sizes might be missed during a pebble count in at least two ways. Due to the presence of larger rocks, they may well have been sheltered or overlapped by these rocks. Also, unless the investigator is particularly careful, he tends to select large rocks in preference to smaller grains during the pebble-counting.

In a few places, especially on flood plains downstream, a surface deposit contained only a few rocks larger than gravel, so that the pebble count necessarily had a small number of observations. However, this possible source of sampling error would apply in only a few of the 24 cases of polymodality. Furthermore, such errors should apply equally to all size classes rather than only to certain classes.

Failure to be scooped up with the "sand" sample is a possible but not very probable source of error, because most of the scarce sizes would not normally be included in a sample of fine material. The field sample almost always contained grains 8 mm in diameter, often included pebbles up to 16 mm diameter, and not uncommonly contained particles up to 32 mm in diameter.

The method of combining the pebble count with the sieve analysis certainly depends on several questionable assumptions, as explained earlier. However, four size classes showed about an equally large frequency of scarcity (table 10). A fault in the mathematical treatment would more likely vary in some way with the grain size.

If the scarcity of the particles in question is real, the possible causes are that: (1) such grains were never produced to any significant degree of abundance; (2) the particles were either retained upstream (mountain channel deposits) or were carried downstreamward beyond the deposits we inspected; (3) the particles were laid down in the sediments we inspected but were not abundant on the surface and so were not sampled.

The theory that such grains were not produced in significant quantities cannot be disproved but does seem highly unlikely, inasmuch as particles ranging from clay to 10-foot-thick boulders were eroded by the floodwaters. The appearance of the polymodality in deposits upstream from which no mountain-channel deposits existed shows that the scarce grain sizes probably were not retained upstream in the headwater region. If particles ranging from 4 to 64 mm are particularly susceptible to entrainment, they should nevertheless have been trapped in highway deltas like the one where Dillard Creek meets highway U.S. 29. In fact, being scarce relative to adjacent size classes upstream, they might even be expected to be more abundant in highway deltas. This, however, was not the case, nor could they be located at greater distances downstream. The distances studied amounted to as much as 12 miles, by which distance the deposits were strictly sand and finer, with average grain sizes less than 1 mm (fig. 50) and excellent sorting (fig. 54). These considerations, especially the absence of excessive quantities of the pertinent grains in highway deltas, cast doubt on the theory that the missing grains were moved all the way to the James River or farther.

It is quite possible that the pertinent grain sizes were present in the flood deposits but were not abundant on the surface. This situation could result, for example, if the floodwaters laid down rocks of all sizes at one or more stages but deposited medium sands and finer grains in scattered patches toward the end of the flood. The scattered sand patches would tend to cover gravel and pebbles while leaving the cobbles and especially the boulders exposed. A surface sampling would then record a relatively large number of big rocks and sand grains but would miss many pebbles and gravel-sized particles. If the deposits had not been destroyed, this possible cause of polymodality could be tested in the field.

In conclusion, flood deposits consisting mainly of grains smaller than coarse sand were all unimodal. Deposits which included a wide range of particle sizes, however, were usually polymodal with a relatively scarcity of particles 4–64 mm in diameter. The most likely reason for this polymodality is that the scarce grains were covered up to a significant extent by sands deposited during the recession stage of the flood. Other possibilities are that many of the deficient sizes were hiding under the edges of the larger rocks and that sampling bias favored the selection of larger rocks during the pebble counting.

DOWNSTREAM CHANGES IN GRAIN SHAPE AND ROUNDNESS

The roundness of a grain is the degree of smoothing or rounding, ranging from very sharp to perfectly rounded, of the edges or corners of the grain. Particles become more rounded due to weathering and to abrasion and wear, that is, with longer time and greater distance of transport. Observations on grain roundness therefore might be helpful in estimating distance of transport.

Grain shape, defined more specifically below, is a three-dimensional measurement of a particle's general form and is a separate concept from roundness. Grain shape depends on amount of abrasion, initial shape, cleavage and natural hardness. The mineralogy generally influences these last three items. Because it affects settling velocity, the grain shape conceivably could influence various aspects of particle behavior in water, such as ease of entrainment and mode and rate of transport. It could also provide information on the hydrodynamic conditions of deposition.

Changes in grain shape and roundness with distance downstream are still poorly understood due to scarcity of data, especially for river flood conditions. Krumbein (1940) studied California flood gravels over a distance of about 7 miles. He found that roundness increased from 0.28 at the 1-mile station to 0.44 at the 7-mile location (0.1 being very sharp particle edges and 1.0 being perfectly rounded edges); grain shape, however, showed no significant change over this distance. Scott and Gravlee (1968) found that a flood surge on a stream in the Sierra Nevada mountains caused boulders to increase in average roundness from about 0.25 to 0.40 over the first two miles of travel. Studies of shape and roundness changes for other environments, such as beaches, rivers at normal flow, and in laboratory experiments, show various and even contradictory relations (Pettijohn, 1957, p. 549).

We examined the deposits of Polly Wright Cove for possible downstream changes in particle shape and roundness. Most of these sediments orginated in a reasonably localized area (about 0.6 sq mi). The total distance of deposition along the stream unfortunately was only about 0.8 mile, the upstream half of which was subject to "contamination" from two minor tributaries, as mentioned earlier. This 0.8-mile distance might be considered too short to show any shape and roundness changes, but laboratory experiments (Krumbein, 1941a) have suggested that roundness values during the first mile of travel can increase from about 0.15 to 0.45, that is, roundness increases considerably during the first mile or two of transport.

The Polly Wright grains examined were in the 0.250–0.500 mm sieve class, from each of the 11 field stations. A colleague coded the 11 samples so that the operator could not know the field location of the grains being examined. Using a microscope, the operator then measured the shape and estimated the roundness of 50 randomly-chosen grains, for each sample. Shape equals the short particle axis divided by the square root of the intermediate times the long axis, where all three axes are mutually perpendicular (Schulz and others, 1954). Krumbein's (1941b) visual comparison chart served as the criterion for roundness. Possible values of both these definitions range from 0 to 1.0, where 1.0 is maximum possible shape or roundness. Owing to the complex and varied mineralogy of the parent rocks it was not possible to restrict the measurements to only one mineral. The grains measured were two unidentified translucent minerals that occurred the most abundantly and consistently throughout all 11 samples.

Table 11 lists the average shape and roundness values for each sampling station. All shape values fall within the range 0.63–0.76. Grain corners and edges were quite sharp, and the low roundness values range from 0.17 to 0.24. The data show no significant change in either shape or roundness with distance downstream.

TABLE 11.—*Average grain shape and roundness values with distance downstream in Polly Wright Cove*

Distance downstream from first deposits (ft)	Average grain-shape value	Average roundness value
0	0.74	0.19
290	.63	.21
740	.74	.23
1,007	.76	.21
1,413	.71	.24
1,803	.73	.20
2,413	.72	.22
2,773	.74	.20
3,439	.71	.21
3,834	.68	.18
4,358	.71	.17

Krumbein's (1940) and Scott and Gravlee's (1968) roundness studies showed a definite increase in particle roundness over the first few miles of travel. The use of gravel-to-boulder-sized particles in those studies may be one reason why the present roundness results differ from the previous work. Also, the infusion of sediment from side tributaries near the middle of the study reach may have confused the present results slightly. Other possible causes of discrepancy might be the range of grain sizes of the transported and deposited materials, the mineralogy and the physical conditions (hydrology and geomorphology) of the floods. All of these discrepancies warrant future study in both field and laboratory.

The absence of an increase in grain shape agrees with Krumbein's field study (1940). His tumbling-barrel experiments (1941a), which may not have simulated catastrophic flood conditions, produced a very slight increase in grain shape over the first mile of travel.

CONCLUSIONS

The storm and flood of August 1969 in Virginia was an extreme event (probably occurring no more than once every several hundred years on the average) which caused enormous amounts of sediment erosion and deposition. In the region most affected, Nelson County, rainfall amounted to as much as 28 inches over the 8-hour storm. Peak streamflow was estimated to be about 10,000–12,000 cfs per sq mi of drainage area.

The longitudinal profiles of the five mountain streams studied, including alluvial fans where present, are very closely described by a hyperbolic equation of the form $F = L/(a + bL)$, where F and L are vertical fall and horizontal distance respectively, from the drainage divide, and a and b are coefficients.

Nearly half of the erosion took place in the form of debris avalanches on upland hillslopes. Field observations and measurements on avalanche scars reveal the following characteristics: (1) the scar on the hillside usually extended up the slope to the region where the local hillside gradient was steepest: (2) this upslope tip of the avalanche tended to be located at the point on the hillside where the convex upper zone merged with the concave or straight section immediately below, a point which could be from 0.3 to 0.95 of the horizontal distance from base to top of hill: (3) debris avalanches usually occurred down previously-existing grooves, minor channels or depressions on the hillside, a finding in agreement with other studies: (4) slopes facing north, northeast and east had more debris avalanching than hillslopes facing other directions; and (5) on a unit drainage area basis, longer hillslopes had a greater total length of avalanche scars but fewer scars than short hillslopes.

Slightly more than half of the erosion that occurred was estimated to be from the enlargement of stream channels.

The total volume of sediment eroded during the storm was estimated to be about 6.1 million cu ft (3.9 million cu ft per sq mi) from the headwaters of Wills Cove, 3.1 million cu ft (4.6 million cu ft per sq mi) from Ginseng Hollow and 3.1 million cu ft (3.2 million cu ft per sq mi) from Polly Wright Cove. These amounts correspond to average denudations of 1.4, 2.0 and 1.4 inches from Wills Cove (upper part), Ginseng Hollow and Polly Wright Cove, respectively, and are probably the equivalent of several thousand years of "normal" denudation.

The types of deposits associated with the storm and flood were: (1) debris-avalanche deposits, quite rare, left at or near the base of hillslopes; (2) mountain-channel deposits, in some cases in the form of huge debris piles behind log jams; (3) alluvial fans where the narrow mountain channels entered the open intermontane valleys; (4) deltas where a swollen stream's path was temporarily blocked by a highway embankment; and (5) flood-plain deposits. The size-frequency distribution of the particles at any locality on such deposits is described by combining a sieve analysis with a pebble count in a method which apparently has not heretofore been used.

Except for highway deltas and some downstream flood-plain sediments, the particle sizes in a deposit generally ranged from clay or silt to boulders. However, nearly all particles larger than about fine sand were deposited after 5 or 10 miles of travel from their source. Some of the silt and clay probably reached the James River and its estuary.

Two prominent alluvial fans each contained about 300,000 cubic feet of newly deposited sediment. The average grain size on alluvial fans showed no distinct trend with distance down the fan. Several other authors have reported that particles tend to become smaller with distance on a fan. Possible reasons for this difference are the short lengths of the present fans (up to about 2,000 ft) and the probably fast velocity of the flow at the fan apex, for this catastrophic event.

Highway deltas vary widely in size. One of the largest contains an estimated 1,900,000 cubic feet of sediment. Such deltaic sediments consist mainly of sand that has distinct laminations. Cross-stratification near the surface and ripple marks and mud on the top surface are also characteristic. As the flood progressed, the particles trapped in highway deltas generally became finer in size, better sorted and more symmetrical in weight-frequency distribution. The storm-average sediment transport rate at the apex of one major delta was estimated to be about 33 lb per sec per foot of channel width.

Flood-plain deposits in the upland region have a wide range of particle sizes and show no distinct sedimentary features. The sandy flood-plain sediments farther downstream commonly have laminations, ripple marks and mud cracks. Of two sites chosen to study textural changes with depth (reflecting relative time of deposition during the flood), the particles at one site become finer and better sorted toward the top surface, whereas at the other site the grains show no noticeable trend except for a layer of finer material on the surface.

The amount of deposition on flood-plains decreased with distance downstream, though at varying and irregular rates. Side tributaries, highway deltas and other factors influenced the measured rates at which volume of flood-plain deposition diminished with distance. Upstream deposition on Rucker Run was about 150–300 cu ft per foot of channel length, but at 12.5 miles downstream the amount of deposition was only minor. Beginning at the head of a valley just beyond the mountain front, flood-plain sediments show the following textural changes with distance downstream: (1) average grain size decreases rather systematically, reaching about 0.1 mm in about 5–7 miles of travel; (2) average grain size therefore decreases with both a decrease in local gradient and an increase in drainage area; (3) the average diameter of the five largest stones, reflecting the stream competence, decreases exponentially; (4) sorting improves to the extent that the d_{90}–d_{10} range decreases from about 2–4 log cycles of grain size to less than one log cycle of grain size, over 5–7 miles of travel; and (5) most size distributions are approximately symmetrical, a notable exception being the deposits in Polly Wright Cove.

Some of the deposits apparently were polymodal, being deficient in fine gravel, a situation common to many other sediments. The gravel particles in all probability were as available as other sizes in the headwaters, were not selectively retained upstream from the floodplain deposits, and were not selectively transported beyond the several miles of study reach. The most probable reason for the supposed scarcity of gravel-sized particles is the possibility that the "missing" grains were mostly covered by sands deposited during the recession stage of the flood.

Particles from Polly Wright Cove in the 0.250–0.500 mm sieve class showed no significant changes in either shape or roundness, over about 0.8 mile of travel.

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