Ephemeral Streams—Hydraulic Factors and Their Relation to the Drainage Net

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PHYSIOGRAPHIC AND HYDRAULIC STUDIES OF RIVERS

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ILLUSTRATIONS

SYMBOLS

- a coefficient in relation of width to discharge
- A_d drainage area
- b exponent in relation of width to discharge
- c coefficient in relation of depth to discharge
- d mean depth defined as ratio of cross-sectional area to width
- D_s size of sediment particle, expressed as median size
- f exponent in relation of depth to discharge
- j exponent in relation of suspended-sediment load to discharge
- k numerical coefficient having a specific but undetermined value
- l stream length
- L suspended-sediment load in units of weight per unit of time
- m exponent in relation of velocity to discharge
- n' a roughness parameter
- N number of streams in a basin
- O stream order
- p coefficient in relation of suspended load to discharge
- Q discharge in volume per unit of time
- $Q_{2.3}$ mean annual flood (equaled or exceeded on the average once every 2.3 years)
- r a numerical coefficient
- s stream slope
- t a numerical coefficient
- v mean velocity defined as quotient of discharge divided by cross-sectional area
- w mean width of water surface
- y exponent in relation of roughness parameter to discharge
- z exponent in relation of slope to discharge

PHYSIOGRAPHIC AND HYDRAULIC STUDIES OF RIVERS

EPHEMERAL STREAMS—HYDRAULIC FACTORS AND THEIR RELATION TO THE DRAINAGE NET

By Luna B. Leopold and John P. Miller

ABSTRACT

The hydraulic factors of width, depth, velocity, and suspended-sediment load of ephemeral streams near Santa Fe, N. Mex., were measured during flood flow. Later, channel slope was measured. These flood-flow data, in conjunction with an analysis of drainage-basin configuration by the methods proposed by Horton, are used to determine the generalized interrelation of stream order and hydraulic variables. The method developed for determining this interrelation allows an integration of the geographic and physiographic characteristics of a drainage basin with the channel characteristics; specifically, the interrelation of the length, number, and drainage area of streams of various sizes with their respective discharge, width, depth, velocity, slope, channel roughness, and suspended-sediment load.

These interrelations show that stream order is related to stream length, number of streams, drainage area, and discharge by simple exponential functions. The relation of discharge to width, depth, velocity, slope, and other hydraulic factors can be approximated by simple power functions. Thus, any pair of these factors is related by exponential or power functions.

The data indicate that suspended-load measurements made during various stages of a few individual floods provide a close approximation to the suspended-load rating curve obtained from periodic measurements taken at a sediment station over a period of years.

The analysis of the hydraulic data shows that in the ephemeral streams studied, velocity increases downstream at a faster rate than in perennial rivers. This appears to be associated with an increase in suspended-load concentration downstream in these ephemeral channels.

The tendency for stream channels to maintain a quasi-equilibrium with imposed discharge and load is shown to be characteristic of ephemeral channels in the headwaters of the drainage basin, even to the most headward rill.

The formation of a discontinuous gully is analyzed in terms of this tendency. This type of gully is characterized by a gradient of the channel bed flatter than the gradient of the valley floor in which the gully is cut. The low gradient is explained as a hydraulic adaptation of channel slope to quasi-equilibrium with the narrow gully width. Examples indicate that as a discontinuous gully widens, its gradient steepens and its length consequently increases. Thus a series of discontinuous gullies tends to coalesce into a continuous trench having a bed gradient nearly equal to the slope of the original valley floor.

INTRODUCTION AND ACKNOWLEDGMENTS

The flow of water in natural channels may be described as perennial, intermittent, or ephemeral. A perennial stream carries some flow at all times. An intermittent stream is one in which, at low flow, dry reaches alternate with flowing ones along the stream length. Those which carry water only during storms, and are therefore called ephemeral, are generally smaller but are much more numerous than perennial ones. From the divide to the mouth of a drainage basin, the increase in channel size is accompanied by a decrease in the number of channels.

Drainage channels are more apparent and more abundant in arid than in humid regions where vegetation hides small rills. Vegetation also tends to increase the length of overland flow, thereby decreasing channel density. For both kinds of areas, however, the land presents to the eye a contrast depending on the perspective one assumes. In arid areas there is the panorama of broad, apparently smooth surfaces sweeping up to the abrupt fronts of mountain ranges or mesa escarpments. A closer examination may show that the whole countryside is actually cut into a myriad of rills, gullies, and arroyos. Humid regions, on the other hand, are typically characterized by rounded, rolling topography which is covered by natural or cultivated vegetation. Although many details of drainage nets are commonly concealed beneath vegetation, careful search discloses the presence of many more channels than can be inferred from a distant view.

Despite the fact that the channels of ephemeral streams are generally recognized to have an important part in the erosion of the land and resultant production of fluvial landforms, they have not received careful or concentrated investigation.. The present report is devoted to a preliminary quantitative description of flow characteristics, channel properties, and configuration of the drainage net in arid regions only. However, it is believed that the principles established may be applicable to humid regions as well.

Practical considerations, along with personal bent, dictated that channels of arid rather than humid regions should receive our attention first. "Accelerated erosion" in the West is a problem of great social and economic importance. The arroyos that trenched alluvial valleys of the Southwest beginning in the latter part of the 19th century, some of which even now are eroding apace, are spectacular examples of ephemeral streams. Conservationists, range managers, geologists, and a host of others have debated the cause of gullying, have experimented with methods of gully control, and have tried to prognosticate the eventual status of the channel system. But in all such efforts practically no attempt has been made to study the process of gully erosion itself, to describe the hydraulic conditions in the eroding channels, or to understand the nature of the equilibrium which was upset by grazing and climatic change. It seems bootless to spend large sums on the control of gullies and arroyos without concomitant attempts to increase knowledge of their hydraulic characteristics.

The present study provides new data that are pertinent to a fuller understanding of ephemeral streams. Specifically, geometric and hydraulic properties of channels are related to the drainage-net configuration. The plan of presentation is as follows: First, the channels in the area studied are described and the methods of investigation outlined. Next, the measurements of hydraulic variables, including channel shape, discharge, and sediment load are presented. A method of integrating hydraulic characteristics with properties of the drainage net is shown to increase markedly the utility of available data. Finally, the role of equilibrium in ephemeral channels, as related to the accelerated erosion problem, is discussed.

Many colleagues and friends gave us continuing advice and assistance in the study, and particular thanks are due M. Gordon Wolman, Thomas Maddock, Jr., John T. Hack, and Walter B. Langbein. The field work was made considerably easier in many ways by the cooperation of Paul C. Benedict and Berkeley Johnson. Charles E. Stearns visited us in the field and later discussed the manuscript with us.

We are indebted to John T. Hack and Stanley Schumm for permission to include in this report some of their unpublished data.

GEOGRAPHIC SETTING AND BASIC MEASUREMENTS

Nearly all the data included in this report were collected in the basins of the Rio Galisteo and Rio Santa Fe, tributaries to the Rio Grande in semiarid central New Mexico (see fig. 1).

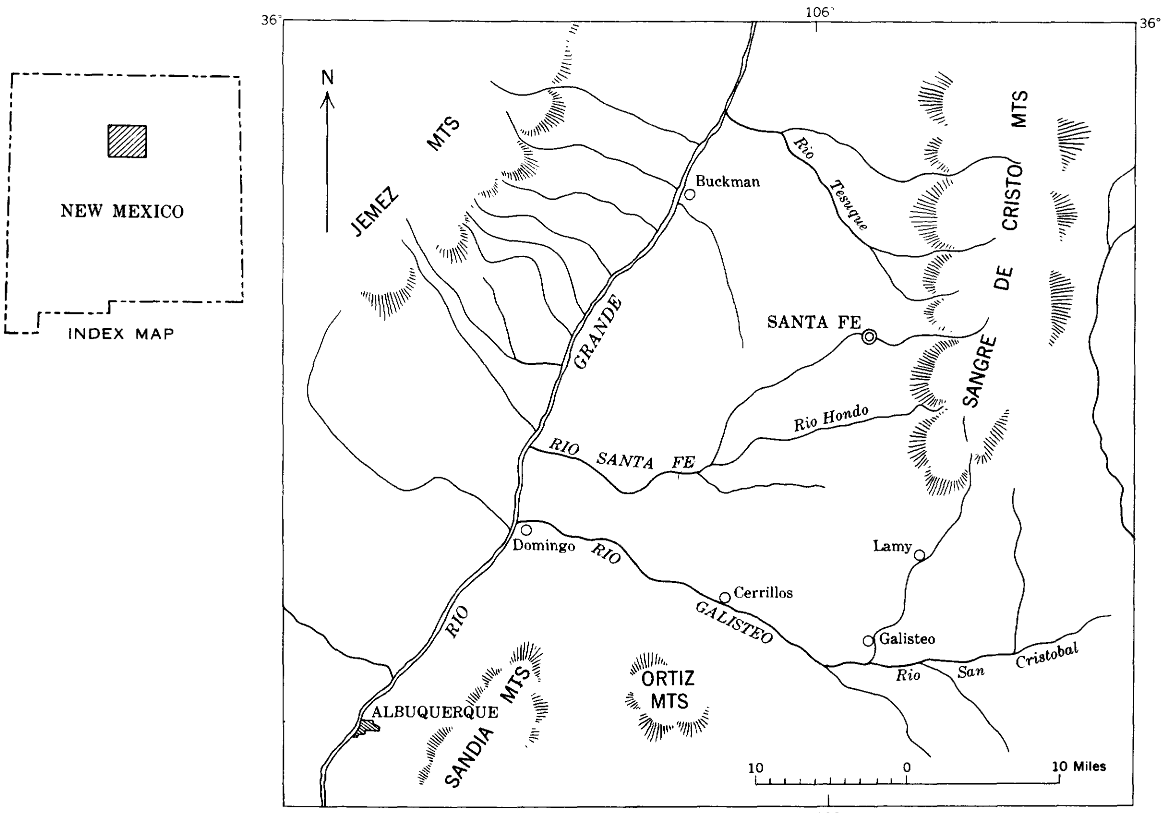


FIGURE 1.—Location map showing area in New Mexico where detailed studies were made.

As one approaches Santa Fe from the west, it appears that the Sangre de Cristo Mountains are abutted by a single relatively broad remnant of an erosion surface cut on poorly consolidated sand and gravel of the Santa Fe formation of Miocene and Pliocene age. From the top of any of the lava-capped mesas near the Rio Grande, one obtains a splendid view of this surface which slopes gently upward to the east toward the mountains. Actually, the plain, which from afar appears smooth, consists of several erosion surfaces differing but little in elevation. Furthermore, close inspection shows that the relief is in fact greater than it appeared from a more distant view. Rolling hills, dissected by gullies and rills and interlaced with sandy, flat-floored washes, are nearly everywhere at hand. The general appearance of streams in the area studied can be visualized by inspection of the photographs in figures 2 and 3. (For definition of stream order see p. 16.) The network of drainage channels ramifies upstream into increasing numbers of successively smaller gullies and rills extending almost to the divides.

As elsewhere in arid regions, these channels present a variety of forms. They range from tiny rills biting back into mesa escarpments to deep trenches or flumelike arroyos incised in otherwise flat alluvial valleys. Despite these striking differences in appearance, they all have certain common characteristics besides the ephemeral nature of their flow. First, vegetation in or along the channel is so sparse that it exerts almost no influence on the form of the channel. Second, because large amounts of fine debris are readily available for transportation both in interstream areas and in the channel itself, the sediment concentration during flows greatly exceeds that of streams in more humid areas. Third, downcutting appears to proceed slowly after the initial development of gullying in a given reach. In a developing arroyo system, channels quickly achieve maximum depth and thenceforth little change in depth occurs despite a considerable thickness of easily removed material.

FIGURE 3.—Channel of Arroyo San Cristobal just below Highway 41, 1 mile south of Galisteo, N. Mex. This stream is of 10th order. Mudballs in the channel bed attained a diameter of 2½ feet. High-water marks indicated recent flows flooded the willows in upper right and nearly reached the top of the vertical bank seen just to the left of the planetable.

FIGURE 2.—Channel of Arroyo Caliente, a small tributary to Arroyo de los Frijoles near Santa Fe, N. Mex. The channel shown is typical of a fifth-order stream in the area studied.

An arroyo in the Southwest discharges water only when a moderately heavy rain falls on the drainage basin. This is typically a summer phenomenon since flow-producing rain falls only from thunderstorms. Winter rains are of too low intensity to provide surface runoff and for this reason arroyos practically never flow in winter.

Summer thunderstorms in New Mexico typically produce rain over 5 to 50 square miles, but the larger coverage ordinarily results from movement of the storm. The storm itself covers, on the average, about 10 square miles. Depth-area curves of thunderstorm rainfall in New Mexico published by Leopold (1942) show that usually not more than 3 to 4 inches of rain falls over 3 square miles and the amount decreases to 1 inch over 50 to 100 square miles. It was our experience that flash flow in arroyos seldom occurs if the rainfall at the storm center is less than 1.5 inches. Another rule of thumb is that for arroyos to flow, the curtain of rain as viewed from a distance must be so dense that one cannot see through it.

During three field seasons we were constantly on the watch for thunderstorms and when one appeared close enough to be reached in less than half an hour, we hurried to get to it. By seriously chasing individual storms, and even though helped by a rather intimate knowledge of local roads and geography, we were able in this period to obtain hydraulic measurements less than a dozen times. Most of them were for locations near Santa Fe, with the others at scattered points in Colorado, Wyoming, and Nebraska. As will be described later, the hydraulic characteristics measured during storms include channel width, depth, velocity, and suspended-sediment load.

It was our plan to combine the measurements of hydraulic factors with data obtained in the field when the channels were dry, and with information taken from maps and aerial photographs. Slope was measured with a telescopic alidade, and channel width was determined by taping or pacing. At the locations where flow was measured, material comprising the channel bottoms was sampled. Drainage areas and stream orders were determined from the excellent planimetric maps compiled by the Soil Conservation Service from aerial photographs and available for parts of central New Mexico.

MEASUREMENTS OF HYDRAULIC VARIABLES IN EPHEMERAL STREAMS

GENERAL FEATURES OF FLOW

The ephemeral nature of flow in arroyos is their most impressive characteristic. Flash floods are the rule. A typical dry arroyo reaches peak discharge in less than 10 minutes, the high flood flow seldom lasts more than 10 minutes, and flow decreases to an insignificant amount in less than 2 hours.

A typical flood rise of small magnitude is shown by the series of photographs in figure 4. This series was taken by Herbert W. Yeo, whose diligence in obtaining sediment records added much to early knowledge of the suspended load carried by New Mexico streams. His work was the prelude to the establishment of regular sediment-sampling stations in that State. These photographs present a graphic account of what we observed several times during attempts to measure arroyo flow.

If a stream rises from nearly a dry condition to a depth of several feet within 10 minutes, one might well speak of a wall of water coming down the channel. Bores as large as 2 feet in height have been seen by the authors. However, observations made during the course of this study cause us to conclude that a "wall of water" is less common than a rapid increase in stage attained through a succession of small surges or bores each a few inches in height. The rise shown in the photographs of figure 4 is believed to be typical in that the initial front does not amount to more than half of the peak depth. Jahns (1949) stated that debris flows may have nearly vertical but slow-moving fronts. As he pointed out, desert floods may include all of the gradations from mudflows to nearly clear water. We believe, however, that the faster moving fronts are seldom vertical "walls of water," but a succession of small bores as described above; Jahns' photographs also support this contention.

A flood in Cañada Ancha Arroyo, July 26, 1952, provided an exceptional opportunity to observe the surges or bores. At maximum flow the width was about 100 feet, mean depth was estimated to be 1 foot, and mean velocity slightly exceeded 5 feet per second. During the 5 minutes immediately preceding peak stage, a series of bores each ½ to 1 foot high moved down the channel at a velocity estimated to be greater than that of the water itself.

The approach of the third bore made it apparent that they were spaced rather regularly in time. Thereafter, we measured with a stopwatch the intervals between successive bores which were 31, 35, 34, 48, and 60 seconds respectively. Between surges the water stage decreased somewhat, as judged by submergence and reemergence of a gravel bar in the channel. Furthermore, the peak stage was much less than the sum of the heights of the eight individual wave fronts.

The nearly constant period between five of the eight surges seems to rule out the possibility that they resulted from successive arrivals of flood peaks from different upstream tributaries. Rather the bores are a type of momentum wave associated with the hydraulics of the channel itself.

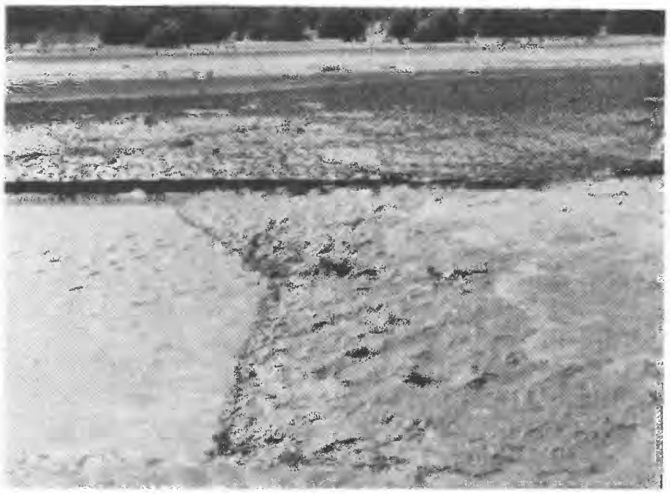
One of our first opportunities to measure the flow in an arroyo was in a narrow and steep-walled channel cut in fine alluvium. After the stage had fallen and the channel was nearly dry, we heard a succession of "plops" which attracted attention. The source of the noises turned out to be falling chunks of bank hitting the nearly dry stream bed. It became obvious that bank caving followed the flood recession.

On every subsequent opportunity, examples of undercutting and bank caving due to high velocity water were specifically sought, but never experienced. Yet, without exception, caving of wetted banks into the channel after the flood was observed. Moreover, on walking any dry arroyo one will find debris fallen

A. Time 1:14 p. m. Upstream view. Bore is covering the small sand island.

B. Time about 1:14¼ p. m. Bore advances faster in deep than in shallow part of channel.

C. Time about 1:14¾ p. m. Floating debris typical of rising flood stage can be seen on surface.



D. Time about 1:15 p. m. View across channel parallel to bore face. Note the slope of water surface as shown by shadow of vertical bank.

E. Time about 1:15⅒ p. m. View diagonally downstream at agitated and debris-strewn area just behind the bore.

FIGURE 4.—Passage of a small bore in the rising stage of an ephemeral flow in an arroyo channel, Rio Puerco, a tributary to the Rio Grande. Location is 8 miles north of Puerco Station, N. Mex., September 19, 1941. Soil Conservation Service photographs.

from the vertical walls and lying crumbled but uneroded in the channel. This debris is picked up and washed away by subsequent flows and is undoubtedly an important source of debris load.

Flash floods in arroyos, therefore, appear to do but insignificant amounts of bank cutting as a direct result of impingement of flow on the banks. Wetting of the banks, however, results in subsequent collapse of arcuate slabs of alluvium which tumble into the channel to become important additions to the load of later floods.

Collapse of gully walls is greatly facilitated by piping or tunnels which develop in the gully walls and lead waters from the adjoining surface to the channel by an underground route (see fig. 5). It was our observation that only a small proportion of total flow in a gully reaches the gully channel by direct overpour of the vertical banks. Piping tunnels and tributary gullies and rills deliver the bulk of the discharge.

FIGURE 5.—Water pouring into open pipe or tunnel leading downward to arroyo channel. The upper opening of this piping hole is at least 15 feet from wall of gully. Rio Pescado near Ramah, N. Mex., August 17, 1946.

The manner in which relatively large pebbles or cobbles move during flash flows is particularly worthy of comment. The bed material in the ephemeral streams studied in the Santa Fe area characteristically is composed of a matrix of moderately well sorted coarse sand, but it includes a certain number of cobbles, rocks, and even some small boulders. The cross section of flowing water during flash floods is wide and shallow, but the velocity of the water is high. Despite the small depth of flow, the large cobbles are effectively moved by rolling. Mudballs move in a similar fashion, rotating about the longest axis. Even cobbles which are irregular in shape and subangular roll along the stream bed for long distances without stopping. Cobbles were observed to roll spasmodically but rapidly even when the water was *no deeper than half the diameter of the rolling object*. At this depth the water seems to splash up on the upstream side of the cobble and plunge over its top, so providing a torque. It is indeed common to see particles, small and large, sticking well out of the general water surface and rolling rapidly downstream with only temporary interruptions.

PROBLEMS OF MEASUREMENT

Attempts to obtain precise measurements of arroyo floods are fraught with many difficulties and inherent dangers. It was necessary to adopt unorthodox methods that yield data which are admittedly crude. Nevertheless, the data themselves are unique, and they appear to be adequate for the kinds of analyses undertaken.

Three factors militate against good measurements of the rising stage of arroyo floods. First, the stage rises to peak so quickly that one can seldom be present during the few minutes of rise even when he is trying to. Second, peak stage is dangerous for a person wading in the flood because of high velocity and the occurrence of surges or bores. Finally, peak flow of consequence occurs generally near the storm center where lightning is a deterrent to wading operations. Hence, most of the hydraulic data presented here were obtained during the falling stage of the individual floods. All measurements were made by wading; velocities were measured with a Price current meter.

When rocks as much as half a foot in diameter batter one's feet and meter in a current flowing 6 feet per second, and when the sand is constantly undermined from under one's heels, short-cut methods inevitably are adopted. For reasons which follow, our discharge measurements of arroyo flow must be considered rough approximations. Instead of using 20 to 30 measuring points or "verticals" across the channel, only 10 to 15 were used. The duration of current-meter observation at each point was reduced from 40 to 20 seconds, and the meter was set at 0.6 depth in most cases. It is standard procedure to make adjustments for variation in water level during the measurement, but stage could be measured only crudely. We traversed back and forth along the tag-line without interruption during the falling flood. Each traverse of the 10 or 15 sections across the channel was considered a measurement of discharge, and the mean stage during the traverse was assumed to apply to that measurement.

Suspended-sediment samples were collected with the DH 48 hand-sampler or, in a few cases, by dipping a bottle without the aid of a hand-sampler. The samples were collected near the midpoint in time of discharge measurements. Usually two samples, each depth-integrated, were collected at two points in the channel cross section. The concentration of sediment was determined separately in the laboratory. The average of the two concentrations was considered to be representative of the flow during the discharge measurement.

CHANGES OF WIDTH, DEPTH, VELOCITY, AND LOAD AT INDIVIDUAL CHANNEL CROSS SECTIONS

Data for establishing the relation of width, depth, velocity, and load to discharge were obtained from two sources: our own wading measurements and gaging station records. Although there are several gaging stations in central New Mexico on streams that are essentially ephemeral, the stations command basins of considerable size. Thus, our wading measurements provide data not otherwise available and also extend considerably the ranges of the several hydraulic variables studied. The gaging station records used for comparison with the data from smaller ephemeral streams are for the Rio Galisteo at Domingo (drainage area about 670 square miles) and Rio Puerco near Cabezon (drainage area 360 square miles.)

Figure 6 shows the relation of width, depth, and velocity to discharge at the gaging station on the Rio Galisteo at Domingo. Figure 7 shows the similar graphs for the gaging station on the Rio Puerco near Cabezon. When graphs of this kind were first presented by Leopold and Maddock (1953) the scatter of points was poorly understood. Wolman (1954a) showed that in a stream where bed scour and fill are minor, the scatter of points is markedly reduced if the successive measurements are made at identically the same cross section. Clarification of this matter is important in that the scatter need not be attributed to lack of adjustment between the channel and the imposed load and discharge.

More scatter of points occurs on the at-a-station curves for streams having sandy or silty beds which are easily eroded than for streams flowing in gravelly channels. Except for the effect of varying position of measurement, the remaining scatter apparently is caused by relatively rapid adjustment of channel shape to changes in suspended-load concentration. These changes in concentration result from the varying location, intensity, or other characteristics of storms and the consequent varying amount of erosion on the watershed lands.

Data obtained from measurements of flash flow in small ephemeral streams are plotted in the graphs of figure 8, which show variation of width, depth, and velocity with changing discharge at each cross section. The same figure includes the lines representing the data from the two gaging stations discussed earlier as typical of the large ephemeral streams in central New Mexico.

When one sees the wide, flat-bottomed, sandy channels characteristic of the ephemeral streams in central New Mexico, he might obtain the impression that even at very low discharges the water flows over the full width of the channel. Actually, as shown in figure 8, width of the flowing water increases progressively with discharge in a quite uniform manner. The rate of increase is about the same as for midwestern perennial rivers. Values of b in the equation relating width, w, to discharge, Q,

$$w=aQ^b$$

at a given cross section range from .09 to .44, and have a median of .26. This is the same as the average value of b for river data studied by Leopold and Maddock (1953, p. 9). For a discharge of 100 cfs, the width of flowing water in the various channels measured ranged from about 30 to 90 feet.

The rate of increase of depth with discharge at a given cross section, shown by the slope of the lines, tends to be generally similar for the several channel cross sections measured. The slopes ranged from .61 to .24, with a median of .33; the slopes represent the values of the exponent, f, in the equation relating depth, d, to discharge,

$$d=cQ^f$$

For rivers, this exponent has an average value of .40. The values here are for mean depth, defined as the quotient of the cross-sectional area of flowing water divided by the width of flowing water. The available data are insufficient to conclude that depth increases with discharge less rapidly in the ephemeral streams studied than in rivers. It should be noted that, as in rivers, the depth in various arroyo channels for a particular discharge varies considerably. For instance, at a discharge 100 cfs, for which six examples are available, the depth ranges from 0.25 to 0.8 feet.

Mean velocity (defined as discharge divided by cross-sectional area) increases with discharge at about the same rate in both the ephemeral streams and perennial rivers. The median value of m in the equation relating velocity, v, to discharge,

$$v=kQ^m$$

is .32 for the arroyos as compared with .34 for rivers previously studied. At a discharge of 100 cfs, mean velocity in the various channels measured ranged from 2.9 to 5.0 feet per second.

As shown previously (Leopold and Maddock, 1953, p. 8),

$$\text{because} \quad Q=wdv$$
$$\text{then} \quad Q=aQ^b\times cQ^f\times kQ^m$$
$$\text{and} \quad b+f+m=1$$

Even without weighing or adjustment, the sum of the respective median values of these three exponents for arroyo data equals .91. With so few data available there is no rational way to adjust the median reported for each exponent. Refinement of these values must await collection of additional data.

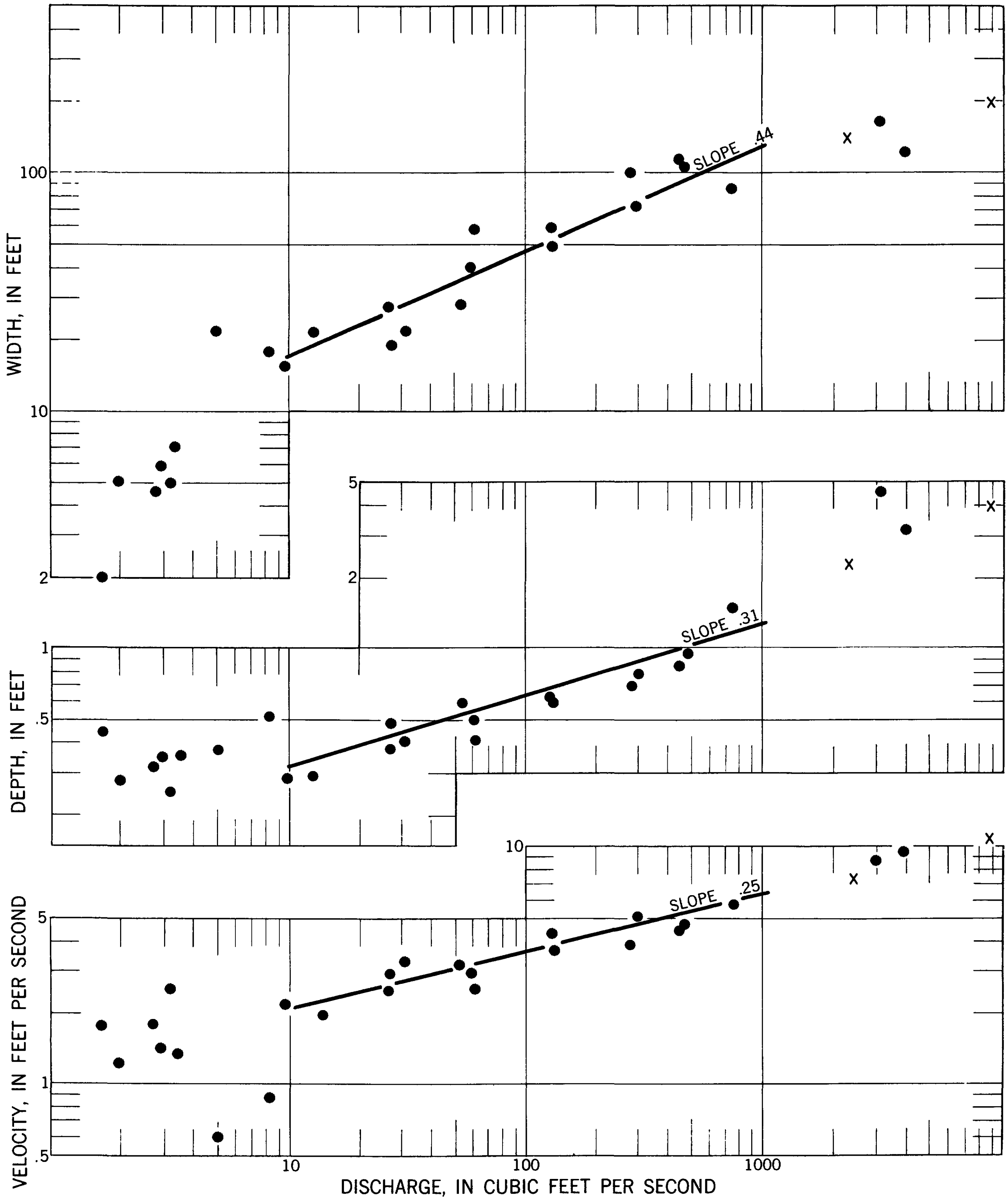


FIGURE 6.—Relation of width, depth, and velocity to discharge at the stream gaging station on Rio Galisteo at Domingo, N. Mex. Crosses indicate slope-area measurements.

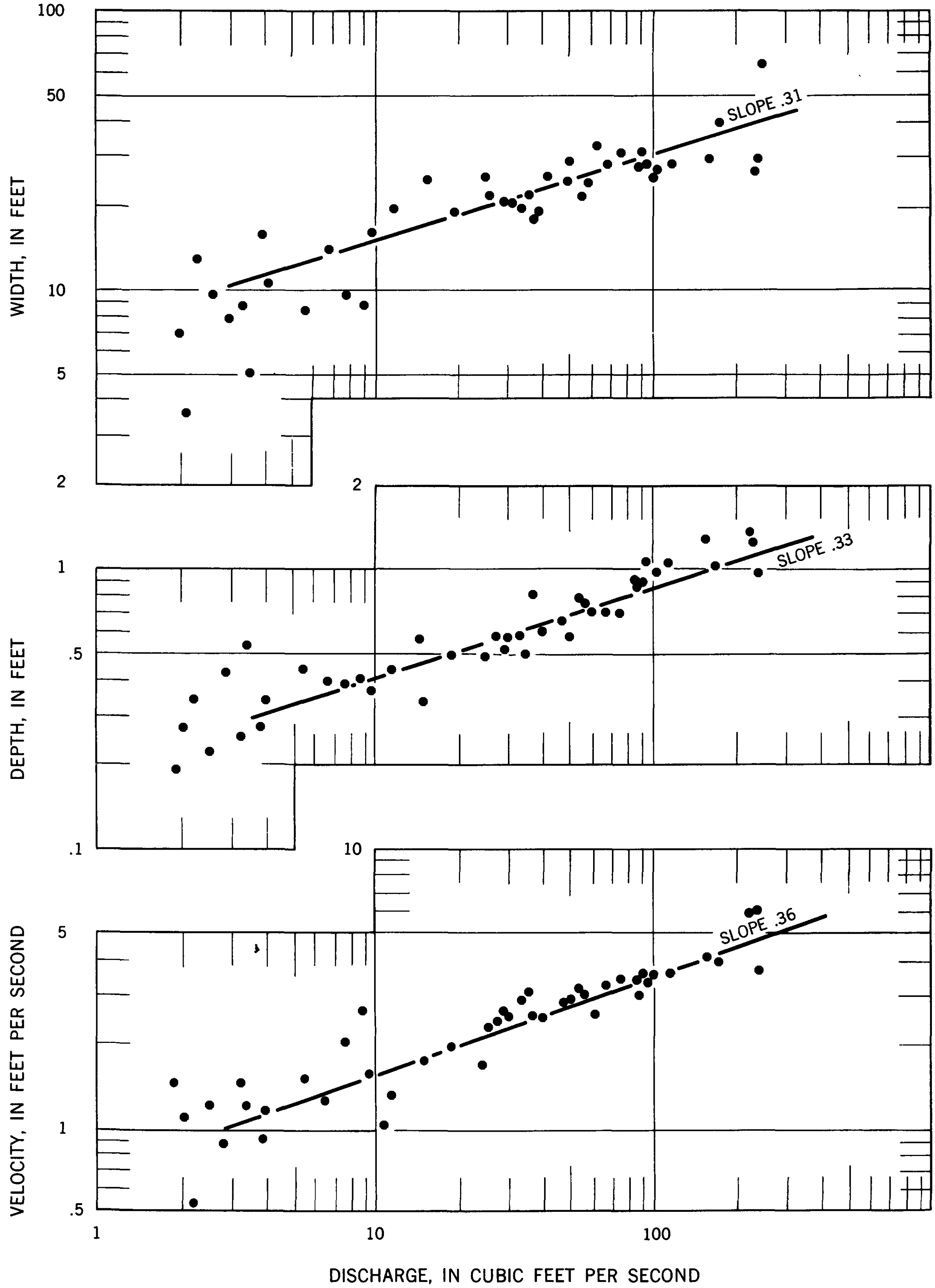


FIGURE 7.—Relation of width, depth, and velocity to discharge at the stream gaging station on the Rio Puerco near Cabezon, N. Mex.

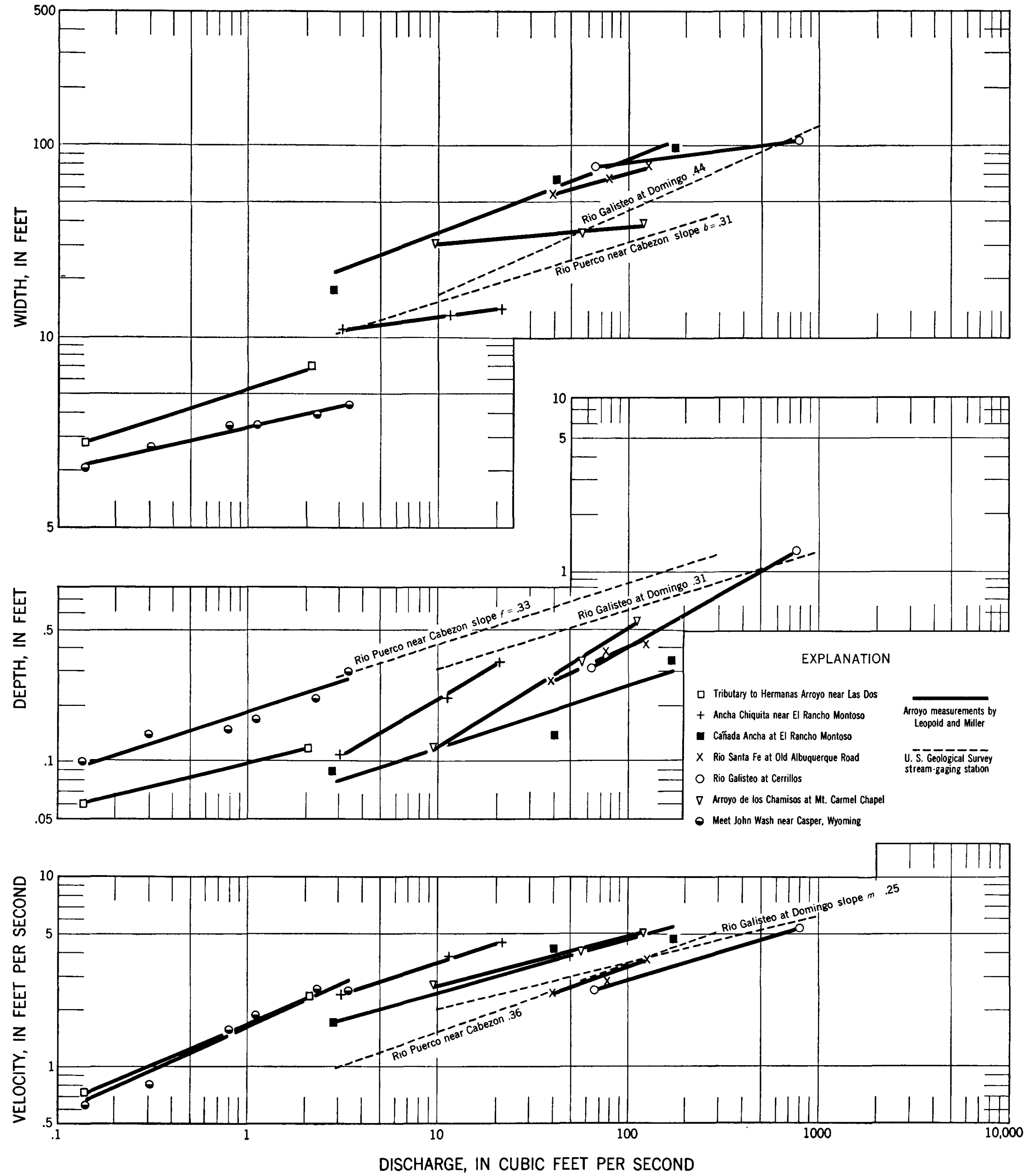


FIGURE 8.—Relation of width, depth, and velocity to discharge at several cross sections in western ephemeral stream channels. These represent at-a-station relations.

Figure 9 presents the data for suspended load at various discharges (the suspended-sediment rating curve) measured at two Geological Survey sediment stations. The same section is used for both discharge and suspended-load measurements at the Domingo station on the Rio Galisteo. For the other example, the sediment station is located 11 miles downstream from the gage on the Rio Puerco at Cabezon. Because no major tributaries enter in this reach of the stream, the sediment and discharge records may be considered essentially comparable.

The data from sediment measurements in small ephemeral streams are shown on figure 10, along with curves for the two sediment stations. In each arroyo cross section we measured only a single flood, whereas in regular sediment operations each measurement represents generally a different hydrograph rise or different day. Nevertheless, lines drawn through our meager data agree very well with the sediment-rating curves of the regular sediment stations. Furthermore, the scatter of our data obtained by wading measurements (fig. 10) is no greater than that for the stations (fig. 9). This suggests that the suspended-load rating curve at a given cross section in a western ephemeral stream can be approximated by a series of measurements during one or two floods.

The relation of suspended load to discharge has been expressed as

$$L = pQ^j$$

where L is suspended load in tons per day (Leopold and Maddock, 1953). Values of j (slope of the suspended-sediment rating curve) for arroyos range from 1.09 to

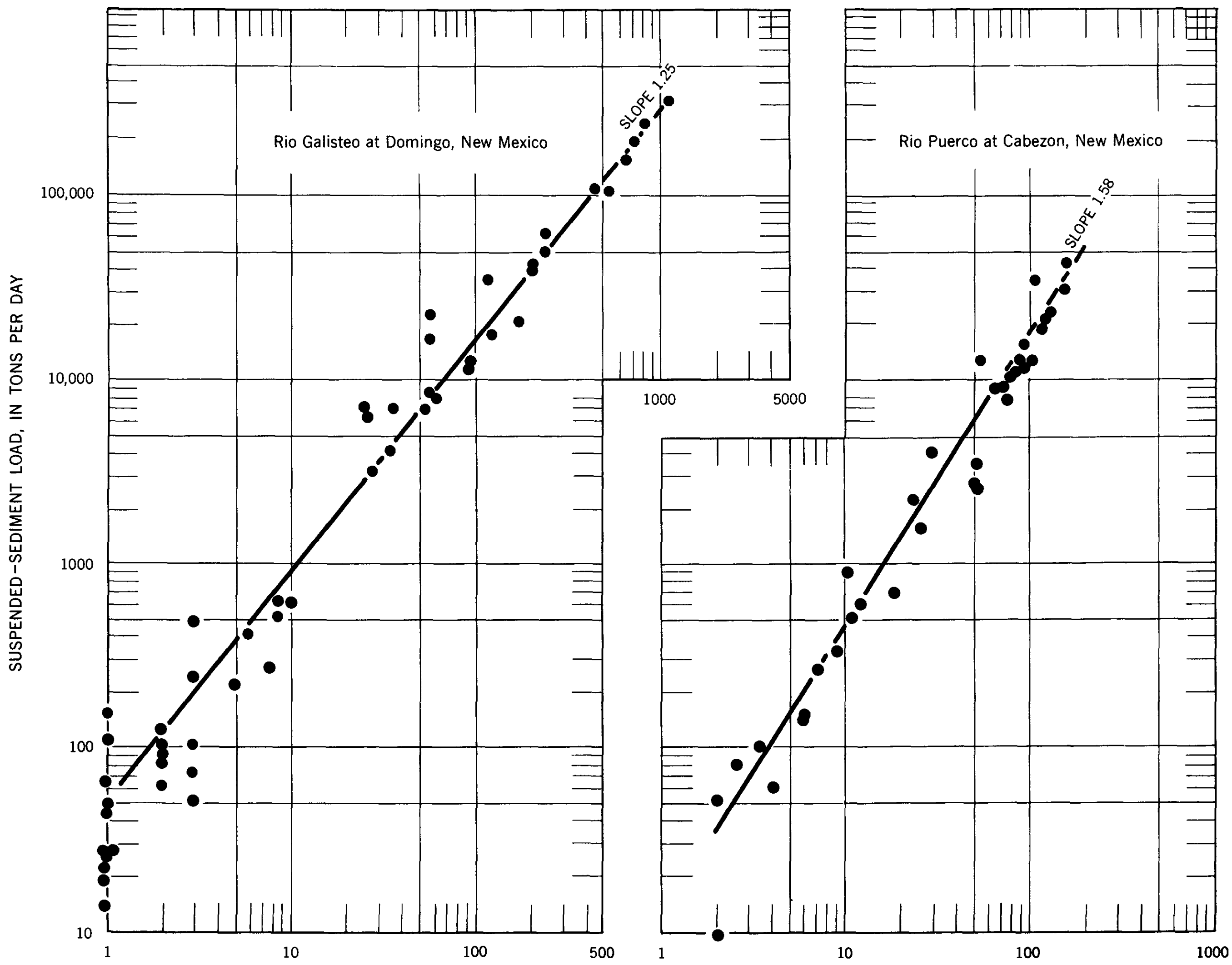


FIGURE 9.—Suspended-sediment rating curves for Rio Galisteo at Domingo, N. Mex., and for Rio Puerco near Cabezon, N. Mex., measured one-fourth mile above the mouth of Chico Arroyo.

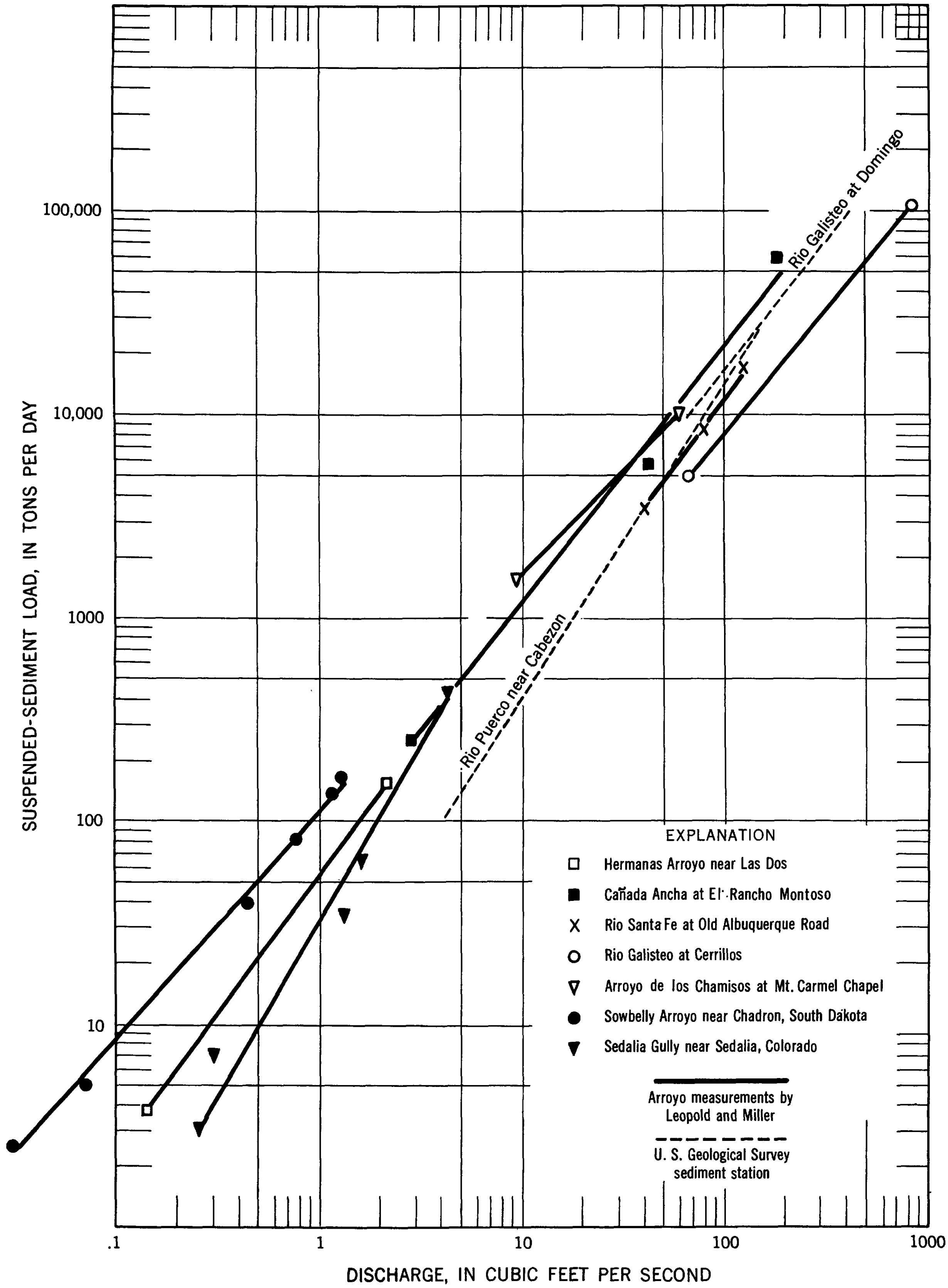


FIGURE 10.—Change of suspended-sediment load with discharge at various cross sections during individual storms. Comparison is made with similar curves for Geological Survey sediment stations.

1.58, with a median of 1.29. For perennial rivers, the average value of j lies between 1.5 and 2.0. As this exponent is greater than unity, load must increase faster than discharge, or in other words, suspended-load concentration increases with increasing discharge.

CHANGES OF HYDRAULIC CHARACTERISTICS IN THE DOWNSTREAM DIRECTION

The graphs just presented show how width, depth, velocity, and suspended load vary with changing discharge at individual channel cross sections. Each discharge on these at-a-station curves represents an occurrence of different frequency. The hydraulic variables (width, depth, and others) at several cross sections along the length of a stream may appropriately be compared only for some constant frequency of discharge. Thus, the change of a variable in the downstream direction can be determined by drawing a line through those points on the at-a-station graphs which represent a discharge of equal frequency at all cross sections. This construction is part of the *hydraulic geometry.*

As the arroyos were measured whenever and wherever an opportunity was afforded, it is not possible to estimate the frequency or recurrence interval of the discharge recorded in the individual floods. We feel certain that none of the arroyo flows gaged were from unusually severe storms.

Lacking other alternatives, we assume that the lines defined by plotting *all* measurements of each variable against discharge represent the downstream changes at some constant frequency of discharge. The plot of width against discharge (fig. 11) lends credence to this assumption. Previous work has shown that the most nearly constant relation among the hydraulic variables in natural stream channels is the rate of increase of width with discharge in the downstream direction. This is the value of the exponent b, in

$$w=aQ^b$$

The average value of b for rivers and regime canals is .5. In figure 11 the slope of the width-discharge relations for all our arroyo measurements is nearly .5 regardless of exactly how one chooses to draw the line through the points. Because of the general agreement in the value of this exponent, it is believed that the plots in figure 11 approximate the downstream relations at a constant, though unknown, frequency of discharge.

The values of f and m, respectively the downstream rate of increase of depth and of velocity, are .3 and .2 for arroyos as compared with .4 and .1 for perennial streams. No great reliance should be placed on the exact value of these exponents for arroyos, considering the nature of the data and the limited number of observations. Indeed, the average values of the same exponents in river data need verification with additional data. Tentatively, however, we conclude that for arroyos the downstream increase of depth is less and velocity greater than had been found for river data. An explanation of these differences appears to be found in a consideration of the sediment load.

Parallelism of the individual at-a-station graphs simplifies development of downstream relations for the sediment data. Regardless of the frequency chosen, any line representing the increase of suspended load with increasing discharge in the downstream direction would nearly coincide with the at-a-station lines already plotted on figure 10. Such a line has a slope of about 1.3, which is the median of the slopes of the individual at-a-station graphs. Although the scatter allows some latitude of choice, it is apparent that any line through the data plotted on figure 10 must have a slope greater than unity. This leads to a conclusion of considerable interest.

From analyzing data for rivers, Leopold and Maddock concluded that in the downstream direction suspended-sediment load increased less fast than discharge. By a rather roundabout analysis, they found the mean value of j to be 0.8. Sediment stations are not arranged along the length of any single river in sufficient number to analyze directly the change of sediment load in the downstream direction. Whereas the j value of 0.8 also could be justified by deduction, the fact remains that those authors did not have so good a check on the downstream change of sediment load as is available in the present data.

The arroyo data indicate that suspended-sediment load increases downstream more rapidly than discharge and, therefore, suspended-sediment concentration increases downstream. From analysis of data for perennial rivers the opposite conclusion was reached; namely, that suspended-sediment concentration decreases in the downstream direction.

A downstream increase in sediment concentration requires (1) that runoff entering the river at downstream points carry larger sediment concentrations than in the headwaters, or (2) that the water pick up progressively more and more sediment off the stream bed as it flows down the channel, or (3) that water be progressively lost by percolation into the channel.

For an area of uniform rock and climate, slopes tend to be steeper near the headwaters than downstream. Drainage density tends to decrease downstream, and this means that the length of overland flow contributing to headwater rills is less than to minor channels which enter the main stream in down-

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stream areas. The flatter slopes and greater lengths of overland flow in the lower parts of drainage basins seem to militate against a downstream increase in sediment concentration contributed to the channel system from the watershed.

Our measurements in ephemeral streams were confined to relatively small drainage basins, all of which were characterized by fairly uniform lithology. It appears necessary then, to assume that the observed increase in concentration results either from a gradual

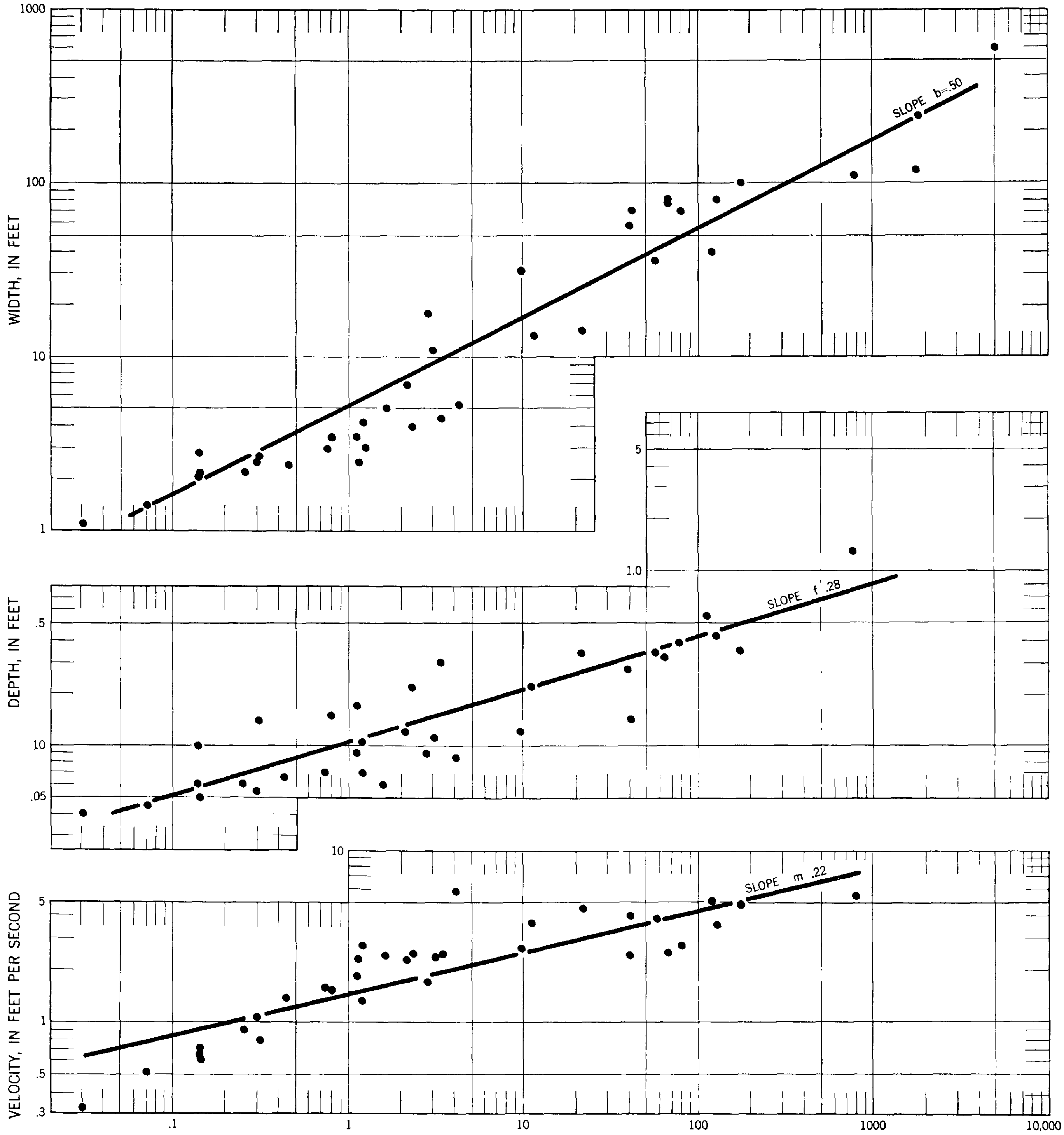

FIGURE 11.—Change in width, depth, and velocity with discharge as discharge increases downstream with drainage area. Points plotted include all individual measurements made on ephemeral channels in New Mexico.

and progressive pickup of sediment off the stream bed or from loss of water. The relative amounts of these two is not known. We presume that the downstream increase in sediment concentration measured by us can be attributed primarily to percolation into the channel bed, but this needs further verification.

Many ephemeral streams in the study area have degraded somewhat since 1860, when a combination of climate factors and overgrazing set off an epicycle of erosion. Most of the gullying occurred relatively soon after this erosion epicycle began. The beginning was by no means simultaneous in all valleys, for some had begun to erode in 1860 and others as late as 1915 (Leopold, 1921). Many are still actively cutting, at least in the headwaters. So, there is some reason to suppose that the channels measured might still be undergoing some deepening which would imply a progressive pickup of debris along the length of the channel. To the extent that this is true, it would account for a downstream increase of suspended-load concentration.

It is well known that arroyo flows gradually lose water by percolation into the stream bed, but the amounts of such loss have been studied only in a few instances (Babcock and Cushing, 1941). These investigators found that the average rate of water seepage into the channel of a dry wash in Arizona during flash floods was about 1 foot of water depth per day. R. F. Hadley (personal communication) observed a flow in Twentymile Creek near Lusk, Wyo., which he estimated as 40 cfs in an upper reach but which was entirely absorbed by the channel in a distance of 6 miles. This channel loss is particularly important in the ephemeral channels in semiarid areas where stream beds are often underlain by considerable thickness of sandy or gravelly alluvium.

Though it is not possible from our data to allocate to these two processes proportions representing their relative importance, the fact remains that it is logical that suspended-sediment concentration should increase downstream in these channels. It should be expected, therefore, that other channel characteristics would show some tendency to adjust to this condition. The manner of adjustment should be shown by the average values of the exponents representing the change of channel factors with discharge, and those values will be summarized and discussed in the following paragraphs.

Before summarizing the salient features of the hydraulic geometry of ephemeral streams described in this report, it should be stressed again that the median values of the exponents relating width, depth, velocity, and load to discharge are derived from a small number of measurements. These median values should, therefore, be considered as merely indicative of magnitude. With this qualification, the data presented here show that the general principles laid out in Professional Paper 252 find verification in western ephemeral streams as well as in a single eastern stream, Brandywine Creek, Pa. (Wolman, 1954a). At cross sections in ephemeral stream channels the relations of width, depth, and velocity each to discharge are generally similar to those found in the perennial rivers previously studied, as can be seen by the tabulation below. The basic equations are listed in the left column, and the main body of the table presents values of exponents in the equations for ephemeral channels and for the average river.

	Average downstream relations		Average at-a-station relations	
	Ephemeral channels	Average river channels[1]	Ephemeral channels	Average river channels[1]
$w=aQ^b$	$b=0.5$	$b=0.5$	$b=0.26$[2]	$b=0.26$
$d=cQ^f$	$f=.3$	$f=.4$	$f=.33$[2]	$b=.40$
$v=kQ^m$	$m=.2$	$m=.1$	$m=.32$[2]	$b=.34$
$L=pQ^j$	$j=1.3$	$j=.8$	$j=1.3$[2]	$j=1.5$–2.0
	$\frac{m}{f}=.7$	$\frac{m}{f}=.25$		

[1] Leopold and Maddock (1953). [2] Unadjusted median values.

To summarize, at discharge of constant frequency in both perennial rivers and ephemeral streams, width increases downstream as the square root of discharge. Velocity, however, increases more rapidly downstream in ephemeral streams, and this tendency is accompanied by a less rapid increase in depth. This more rapid downstream increase in velocity can be interpreted as a response to the downstream increase in suspended-sediment concentration. Indeed, the observed relations are in agreement with the principle enunciated by Leopold and Maddock (1953, p. 26) that "the rate of increase of suspended-sediment load with discharge is a function of the ratio:

$$\frac{\text{rate of increase of velocity with discharge}}{\text{rate of increase of depth with discharge}}\text{''}$$

Those authors state (p. 26) the generalization in this way: "for any given value of b * * * the value of j increases with an increase in the m to f ratio." The agreement with this principle can be visualized by comparing the downstream values of these exponents in the ephemeral channels with those in the average river data.

So, it can be seen that the channel factors in the ephemeral channels appear to differ from those of the average river in a manner which is in accord with the observed change of suspended-load concentration.

INTERRELATION OF DRAINAGE NET AND HYDRAULIC FACTORS

RELATION OF STREAM ORDER TO STREAM NUMBER, STREAM LENGTH, AND DRAINAGE AREA

The quantitative description of drainage nets developed by Horton (1945) related stream order to the number, average length, and average slope of streams in a drainage basin. Our purpose in this section is to show how this useful tool may be extended to include the hydraulic as well as drainage-net characteristics.

All the data required for the Horton type of drainage-net description can be obtained from maps. As maps of several different scales were required for our own analysis, some explanation of the procedure is in order.

Figure 12 presents planimetric maps of a sample area near the city of Santa Fe. The map on the right, which shows the drainage net in a typical watershed about 9 miles long by 2 miles wide, was compiled from planimetric maps made by the Soil Conservation Service from aerial photographs at an original scale of 2 inches to the mile. The left map shows in more detail the drainage net in one small tributary which for purposes of this report we will refer to as Arroyo Caliente. This map was made by pace and compass after a planetable traverse had been run for control. Each tributary rill was paced out to its farthest upstream extension in order that the map would include all recognizable channels.

The orders of various channels in the basin of Arroyo Caliente are indicated by numbers appearing in the upper part of the left map of figure 12. A small unbranched tributary is labeled "order 1," and the stream receiving that tributary is labeled "order 2." All streams of orders 3, 4, and 5 are labeled with appropriate numbers near their respective mouths.

On the right map of figure 12, the little basin called Arroyo Caliente is one of the minor tributaries which even on this small-scale map appears to be unbranched like the tributary just west of it. If only this map were available, one would conclude that Arroyo Caliente is a first-order stream. This points up an important qualification to the Horton scheme of stream-order classification; namely, that the definition of a first-order stream depends on the scale of the map used. The first-order stream, by definition, should be the smallest unbranched channel on the ground. The designation of which stream is master and which is tributary is somewhat arbitrary, but we have followed the guide suggested by Horton (1945, p. 281).

The largest drainage basin which is included in the present analysis is that of the Rio Galisteo (fig. 1). At its mouth this basin drains about 670 square miles. Such an area contains a very large number of small tributaries. It was desired to estimate the number of tributaries of various sizes, their lengths, and other characteristics. The task of counting and measuring each individually would be inordinately great but approximate answers could be obtained by a sampling process. Arroyo Caliente is one of the samples used.

The detailed map of Arroyo Caliente was used to determine the number, the average length, and average drainage area of each order of stream in its basin. At its mouth, Arroyo Caliente is of fifth order.

The Arroyo de los Frijoles basin shown in small scale at the right in figure 12 was used as another sample and similar measurements were made. The small, unbranched tributaries on this 1 mile to 2 inch map, of which Arroyo Caliente is one, would be designated order 1, in accordance with the definition of stream order. The detailed study of Arroyo Caliente showed, however, that on the ground this tributary which had appeared unbranched was, in reality, composed of a drainage network of still smaller tributaries. Arroyo Caliente and other channels which appeared as order 1 on the right-hand map are, on the average, actually of 5th order. Thus the true order of any stream determined from the right-hand map of figure 12 is increased by adding 4, so that an order 1 stream on that map becomes order $1+4$, or 5.

This provides a way of combining maps of different scales to carry the numbering of stream order from one map to the other. It can be seen in the plot of stream length against stream order (left graph of figure 13) that this relation in stream orders 1 to 5 (average values of Arroyo Caliente) fits well with data from the small-scale maps after the values of stream order were adjusted as described above.

Similarly, this principle was used to obtain the estimate of the number of streams of each order in an 11th order basin in the area studied, as shown in the right diagram in figure 13. An actual count of the number of streams of highest order was made. From the small-scale maps the order of the Rio Galisteo at its mouth (fig. 1) was determined to be 7, which when adjusted by 4, indicates the true order of 11. Because this is the only stream of order 11 in the area studied, the graph must go through the value of 1 on the ordinate scale at an abscissa value of 11. The mean relation was drawn for the numbers of streams of highest orders and extrapolated to determine the number of streams of order 5, the smallest tributaries shown on the small-scale map. The graph of number of streams of orders 1 to 5 in the Arroyo Caliente basin which included only one stream of order 5 was superimposed on the graph determined from the 2 inches to 1 mile map and placed so that the points representing order 5 coincided. By extending the graph to order 1, the number of 1st order tributaries in the 11th order basin could be estimated.

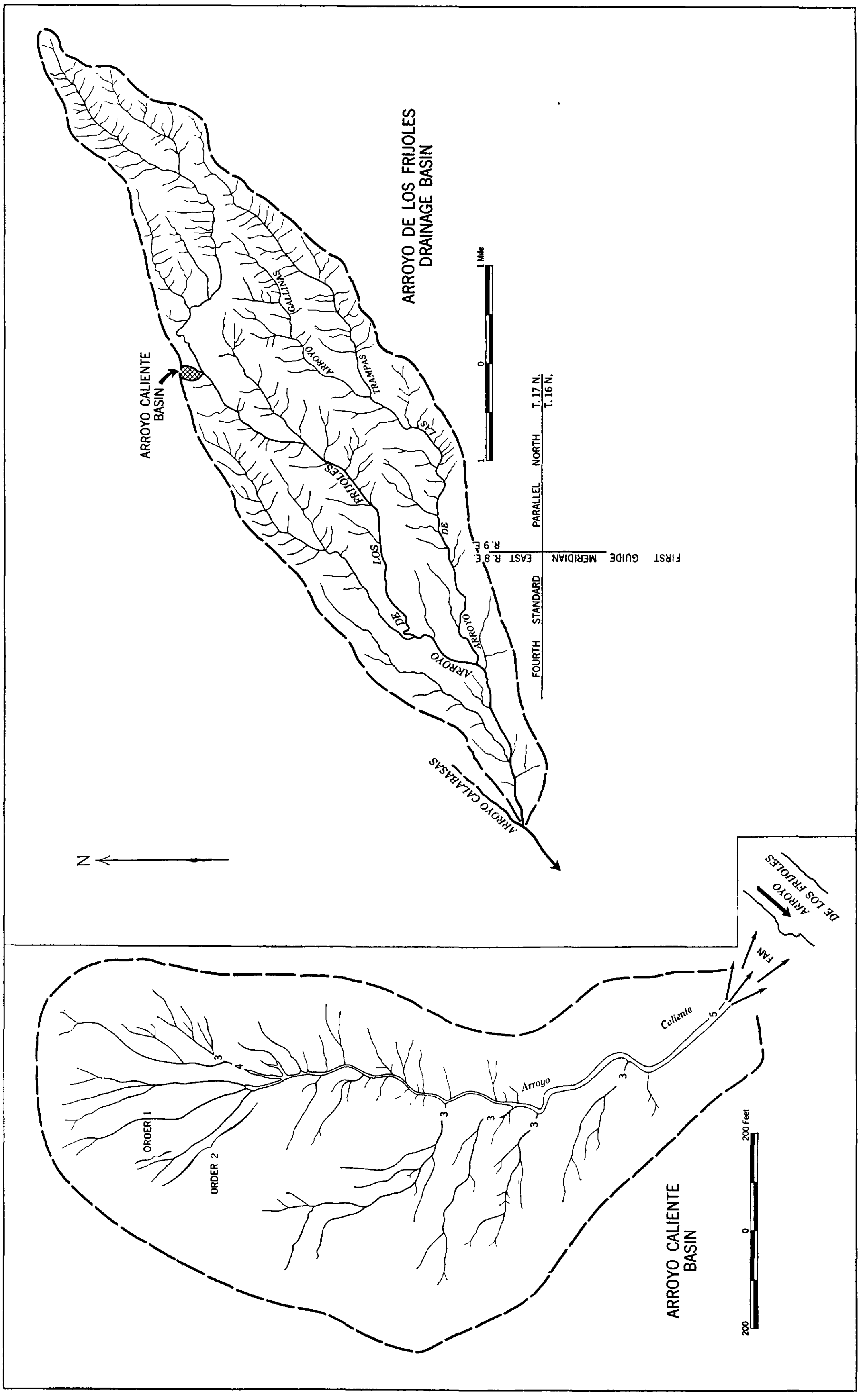

FIGURE 12.—Drainage basin of typical ephemeral arroyo near Santa Fe, N. Mex. Left, basin of Arroyo Caliente, a tributary to Arroyo de los Frijoles; right, basin of Arroyo de los Frijoles showing location of the tributary Arroyo Caliente.

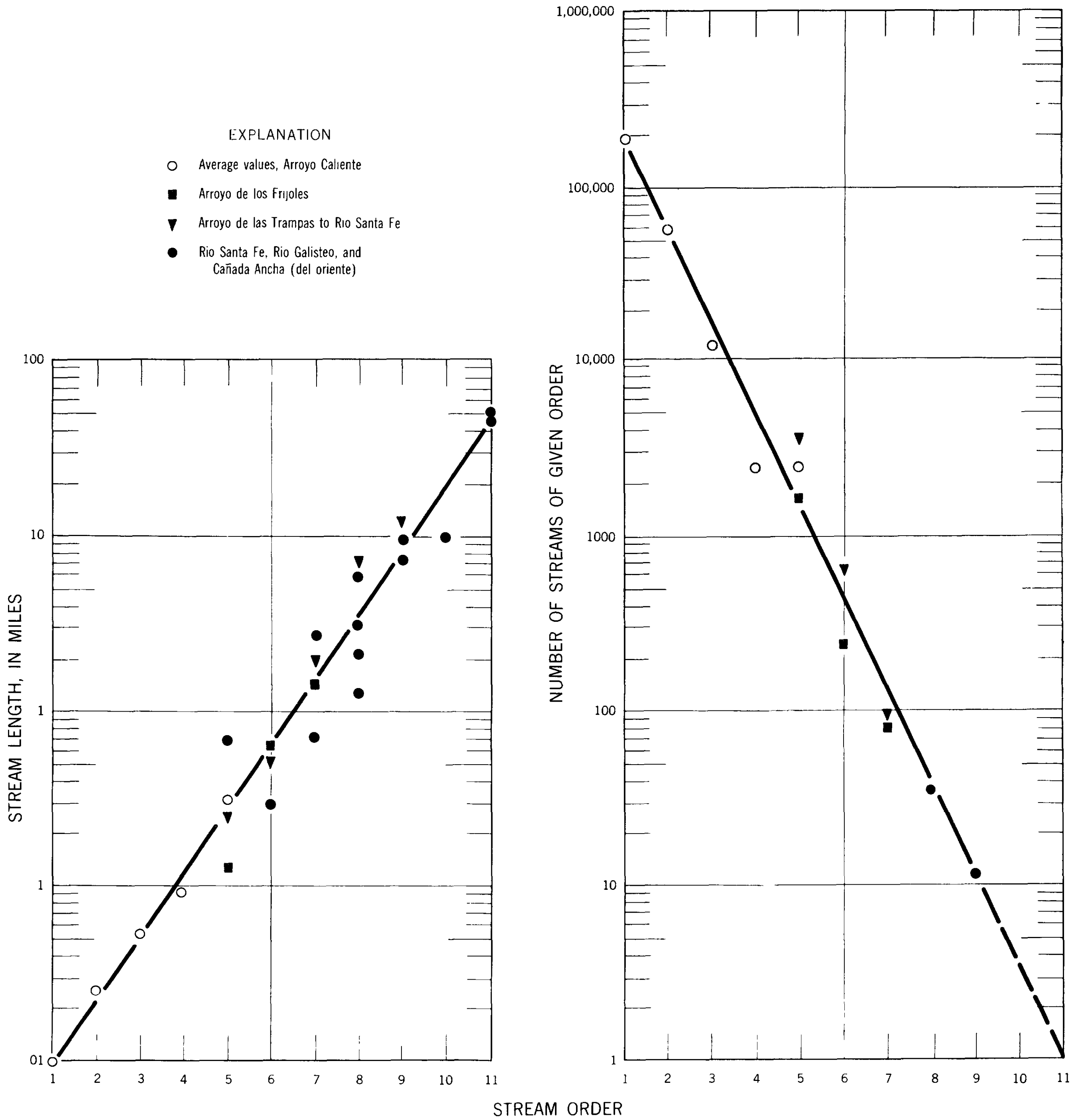

FIGURE 13.—Relation of stream length and number of streams to stream order in basins of 11th order in central New Mexico. Left, stream length plotted against stream order; right, number of individual streams of a given order plotted against order.

This same general procedure was followed for several other sample drainage basins. Order 11 was the highest found, this being for the main stem of Rio Galisteo. The number and average length of streams in each order were determined and the results are included in figure 13.

It will be noted that the plots of stream order against stream length and stream order against number of streams are straight lines on semilogarithmic paper, as Horton discovered. Lengths range from about 50 feet for the 1st order tributaries to 54 miles for the Rio Galisteo.

In order to visualize better the types of channels studied and their relation to stream order, the reader is referred to the photographs in figure 14, which show first-order tributaries in the basin of Arroyo Caliente,

FIGURE 14.—First-order tributaries in basin of Arroyo Caliente. These show the most headward extensions of the smallest tributary rills in the area.

and figure 15 which pictures Arroyo de los Frijoles at a place where its size is typical of an 8th order stream. Channels of order 5 and order 10 can be seen in figures 2 and 3, respectively.

Because maximum stream length is a function of drainage-basin area, it is not unexpected that the relation of drainage area to stream order is also a straight line on semilogarithmic paper, as can be seen in figure 16. The smallest unbranched tributaries, which are rills about 8 inches wide and 1 to 4 inches deep, drain on the average about .00006 square miles or .04 acre. In the 670-square-mile basin of Rio Galisteo there are roughly 190,000 such first-order tributaries, as estimated from figure 13.

FIGURE 15.—Arroyo de los Frijoles at place where it typifies a stream of eighth order.

EQUATIONS RELATING TO HYDRAULIC AND PHYSIOGRAPHIC FACTORS

From the previous work of Horton or from our data plotted in figures 13 and 16, it is apparent that stream order, O, bears a relation to number of streams, N, in the form

$$O=k \log N \text{ or } O \propto \log N \tag{1}$$

and a similar relation to stream length, l, slope, s, and drainage area, A_d,

$$O \propto \log l \tag{2}$$

$$O \propto \log s \tag{3}$$

$$O \propto \log A_d \tag{4}$$

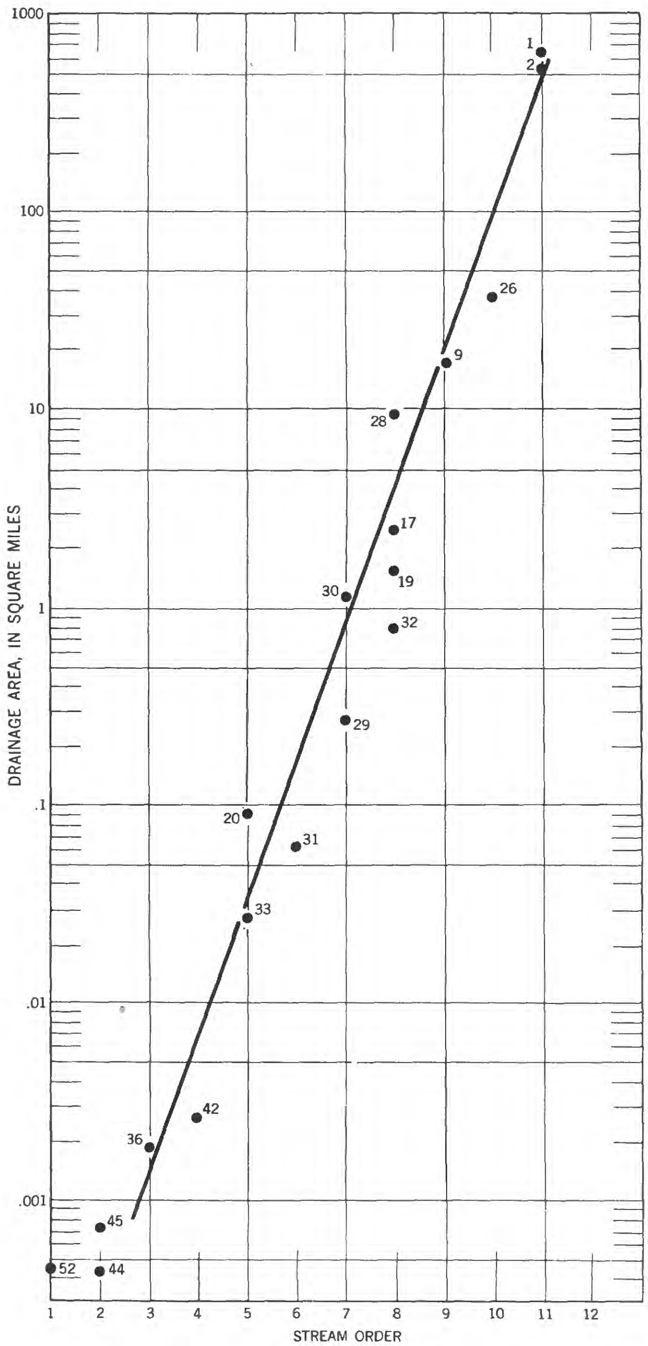

FIGURE 16.—Relation of drainage area to stream order for ephemeral channels in central New Mexico. Number beside point is serial identification keyed to data in appendix.

It follows, then, that there must exist a power-function relation between each two of these variables other than order, in the form

$$l \propto A_d^k,\ s \propto A_d^k,\ s \propto l^k$$

where k is some exponent having a particular value in each equation. Moreover, because it is known that discharge, Q, bears a relation to drainage area of the form

$$Q \propto A_d^k \qquad (5)$$

then discharge must be related to stream order according to

$$O \propto \log Q \qquad (6)$$

Reiterating some of the hydraulic relations discussed earlier:

$$w \propto Q^b \qquad (7)$$

$$d \propto Q^f \qquad (8)$$

$$v \propto Q^m \qquad (9)$$

$$L \propto Q^j \qquad (10)$$

$$s \propto Q^z \qquad (11)$$

$$n' \propto Q^y \qquad (12)$$

Where Q=discharge
w=width
d=depth
v=velocity
L=suspended load
s=slope
n'=a roughness parameter
b, f, m, j, z, and y are numerical exponents

It follows from equation (6) and equations (7) to (12) that there definitely is a relation between stream order and width, depth, velocity, suspended load, slope, and roughness of the form

$$O \propto \log\ (\text{width, depth, etc.})$$

Also, there are several interrelations among variables in equations (1) to (4), and among those in equations (7) to (12). An example of such a relation is that which would exist between sediment load, L, and drainage area, A_d.

Because $L \propto Q^j$
and $Q \propto A_d^k$
then $L \propto A_d^{jk}$

Thus, it can be seen that a whole series of hydraulic and drainage-network factors are interrelated in the form of power or exponential functions. These equations can be added to the several others derived by

Horton (1945, p. 291) which, in his words, "supplement Playfair's law and make it more definite and more quantitative. They also show that the nice adjustment goes far beyond the matter of declivities."

The equations listed above merely state the condition of proportionality; for them to be definitive the constants involved must be determined. In addition to the values of certain exponents already discussed, data collected during this investigation established the relations among order, slope, and width, as will now be described.

Although channel depth, velocity, and discharge at a particular cross section can be measured only when the stream is flowing, the important hydraulic variables, channel width and channel slope, can be estimated even in a dry stream bed. During the many days when no thunderstorms were occurring in the area studied, our field work included measurement of width and slope in the dry channels. The procedure used will be described briefly.

The width was defined as width near bankfull stage. In Eastern United States where the development of the river flood plain is the rule rather than the exception, the bankfull width is relatively easy to define and measure. It is the width which the water surface would reach when at a stage equal to the level of the flood plain. In dry arroyos where flood plains are the exception and alluvial terraces exceedingly common, it is difficult to point specifically at an elevation which might be called bankfull. Nevertheless, field inspection allows fairly consistent estimates of width corresponding to an effective or dominant discharge, even by different observers. The positions on the stream banks representing the two ends of the cross section were chosen independently by each of us, and the recorded widths represent a compromise between our individual judgments.

Channel slope is somewhat easier to determine. Although there are local dry pools or deeps resulting from both definite patterns of flow and random channel irregularities, a smooth profile of channel bed drawn through several points in a reach provides an estimate of channel slope which is reproducible by successive measurements. Our procedure was to place the planetable in the center of the dry stream bed and take a series of sights both upstream and down as far as the stadia rod could be read with a leveled instrument. In most instances this gave a measurement of slope through a reach of about 600 feet.

Width and slope were measured at more than 100 channel cross sections. Because of the widespread geographic distribution of measured cross sections and because of the lack of adequate maps, we have not determined the order of the streams on which many of these are located. The graphs of figures 17 and 19

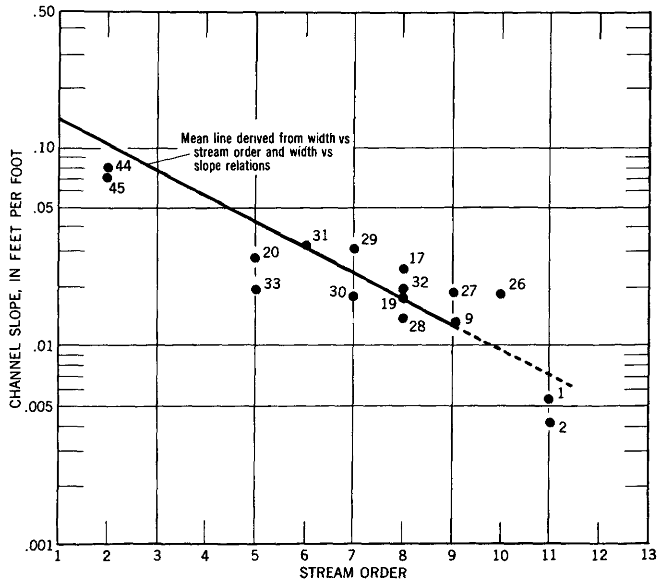

FIGURE 17.—Relation of channel slope to stream order, showing the individual measurements and the mean relation (solid line) derived from other parameters. Number beside point is serial identification keyed to data in appendix.

were plotted from data on cross sections of streams of known order. As expected, order is related to width and slope by equations of the type.

$$O=k \log w \text{ and } O=k \log s$$

The relation between width and slope can be determined by equating the relations above or by direct plotting of the data as in figure 18. To represent the relation in larger streams, a straight line on logarithmic paper has been drawn through the numbered points (fig. 18). The equation of this line is

$$s=0.12w^{-0.5}$$

This means that for the area studied, channel slope decreases downstream approximately as the reciprocal of the square root of channel width. This expresses an interrelation and does not imply a direct dependence between these two parameters.

Establishing the relations among width, slope, and order merely requires collecting and plotting the appropriate data. It should be possible to combine those factors which are most easily and definitively measured, and thereby arrive at approximations of other factors which are more difficult to measure. Such a scheme would expand the usefulness of the techniques now available. An example to illustrate this possibility will now be cited.

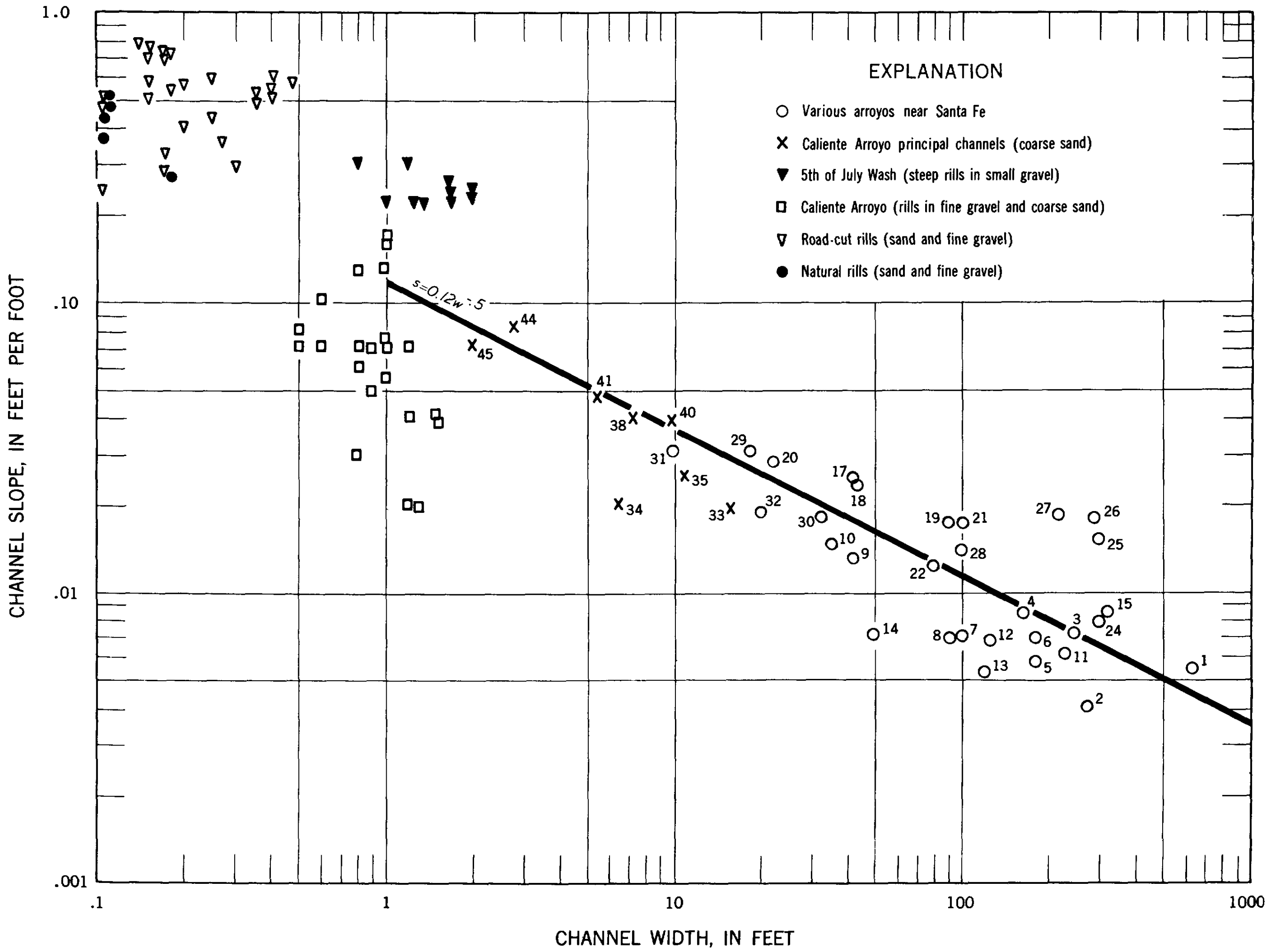

FIGURE 18.—Relation of channel slope to channel width. Data from moderate and large arroyos and differentiated from rills by the different symbols. Number beside point is serial identification keyed to data in appendix.

The manner in which discharge increases with stream order is of particular concern, as it provides the link between the Horton analysis and the hydraulic geometry. For perennial streams, gaging-station data are available in quantity for the larger stream orders. The determination of stream order at a gaging station is laborious but can be made, and order can be plotted directly against discharge of a given frequency derived from gaging-station data. In arid regions there are relatively few gaging stations, and for the most part they are located on the larger streams. To obtain a relation between order and discharge for ephemeral streams requires discharge measurements of the smaller streams. But even these measurements do not provide at a given cross section a specific value of discharge which can be plotted against stream order, for it must be remembered that discharge at any given location may fluctuate through a wide range and that sufficient record is required to allow some kind of frequency analysis.

If the discharge chosen has some specific relation to the position of the cross section along the length of the stream, an approximation to constant frequency might result. It is proposed to use that discharge which corresponds to the average width representative of the stream order at the point in question. To obtain this value, the relations already demonstrated can be utilized. Specifically, the relation between width and order indicated on figure 19 will be combined with the relation of downstream increase of width with increasing discharge. The latter is, fortunately, one of the most consistent of the graphs representing the hydraulic variables. From figure 19 the stream width for each order can be obtained and, by use of those values of width, the corresponding discharges can be read from the width-discharge graph of figure 11. In such a manner the relation between stream order and discharge may be derived, and it is plotted as the unbroken line on figure 20.

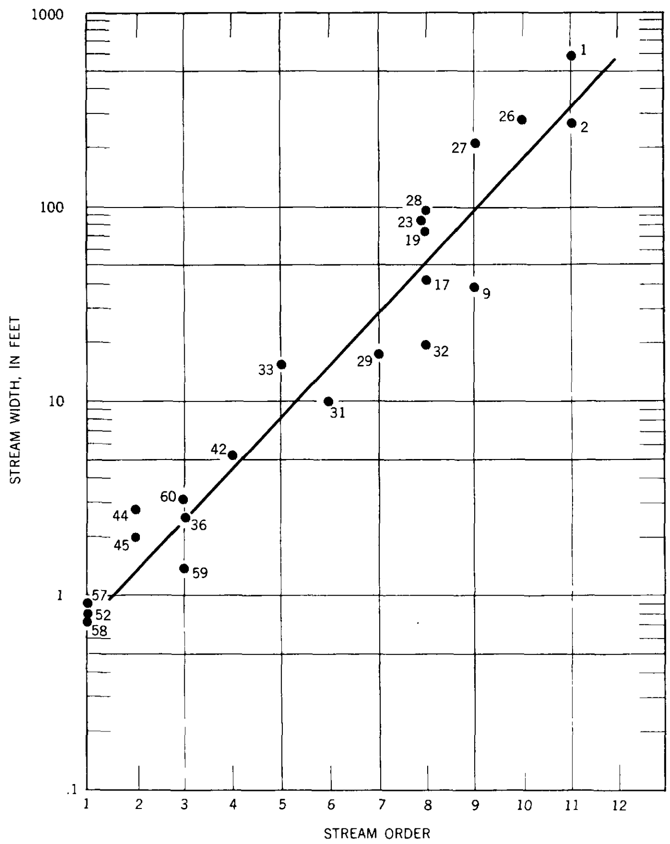

FIGURE 19.—Relation of stream width to stream order in arroyos. Number beside point is serial number in appendix.

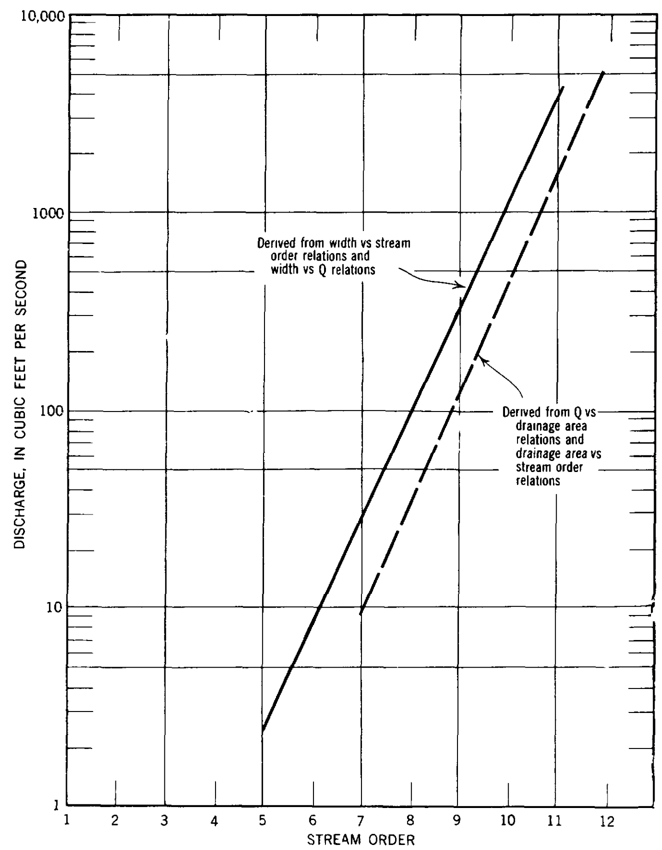

FIGURE 20.—Relation of discharge to stream order derived by two separate types of analyses.

A second approach to the discharge-order relation makes use of flood-frequency analysis. On streams of small and moderate size in central New Mexico there are only a few gaging stations that have a relatively long period of record. Despite the paucity of data, flood-frequency curves for all gaging stations considered applicable to the area being studied were obtained from an unpublished study by H. H. Hudson. Also, the records of some additional stations were analyzed by the authors. In eastern streams the bankfull stage is attained about once a year. A conservative quantity little affected by the length of record is the average of the highest flood each year of record. This is called the mean annual flood. It has a recurrence interval of 2.3 years, and thus we assume roughly approximates the discharge at bankfull stage. The relation between 2.3-year flood discharge and drainage area for gaging stations in central New Mexico is presented on figure 21.

It is known by hydrologists that the discharge of a flood of a given frequency increases somewhat less rapidly than drainage basin size. It is typical for the relation of flood discharge to drainage area to plot as a straight line on logarithmic paper, expressed by the equation

$$Q \propto A_d^k$$

where k is a constant which for the mean annual flood has a value of between 0.7 and 0.8. The slope of line on figure 21 is consistent with values known from other areas. In drainage basins of the same size, those studied in the West produce floods of about one-fifth the magnitude of the typical one in the East.

By combining the discharge-area graph with the order-area graph, discharge can be related to order. The drainage area corresponding to each stream order is read from figure 16, and the value of discharge for an equal drainage area is determined from figure 21. The resulting plot is shown as the dashed line in figure 20.

Thus, the relation between stream order and discharge, which is the link between the Horton analysis and hydraulic geometry, has been derived for a central New Mexico area in two ways which are at least somewhat independent. Comparison between the results is indicated by the two lines in figure 20. Only the slopes

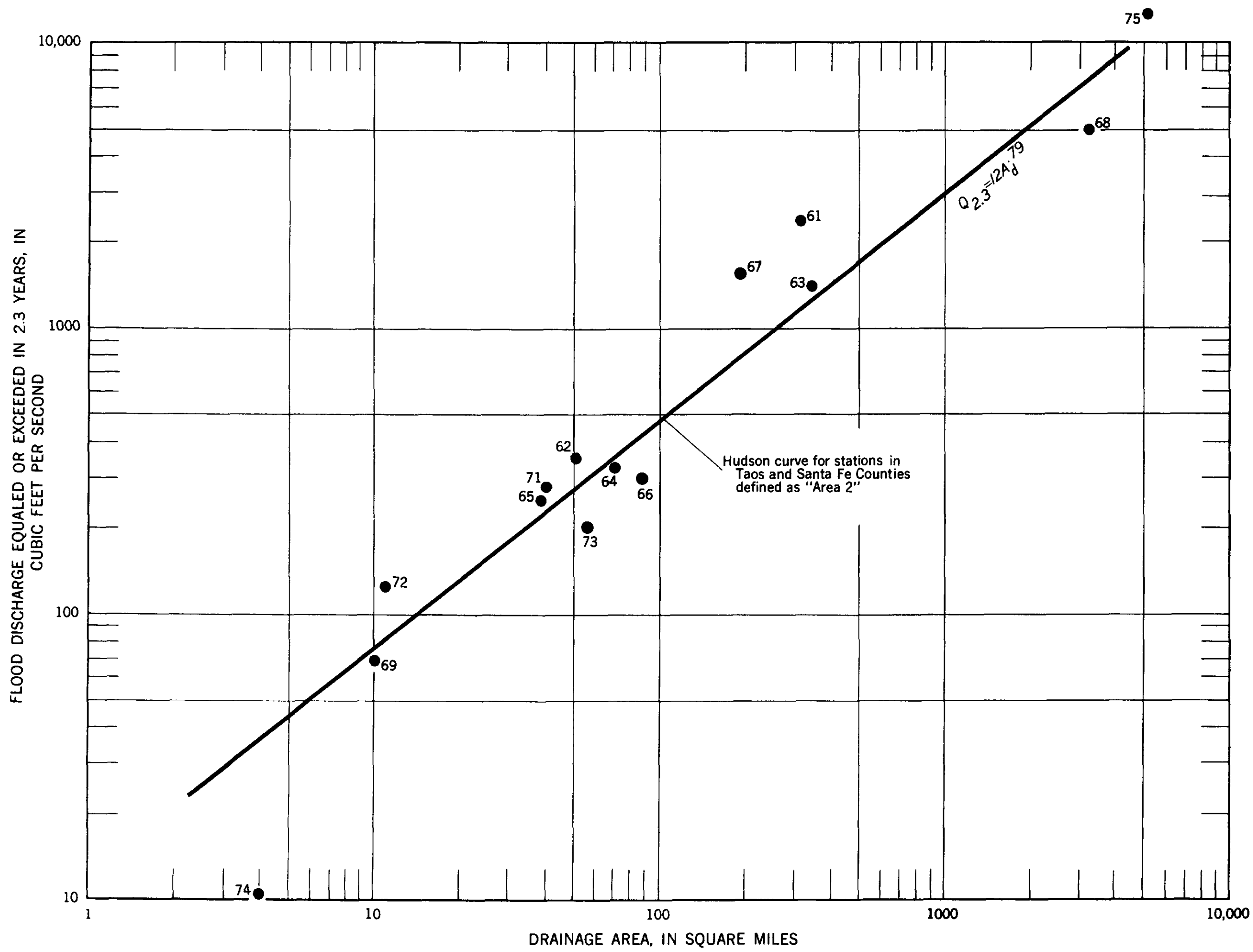

FIGURE 21.—Relation of mean annual flood discharge (equaled or exceeded in 2.3 years) to drainage area. Number beside point is serial number in appendix.

of the two lines should be alike, and indeed, they are similar. The intercepts should not be expected to be exactly the same because the discharge determined from flood-frequency data was chosen to represent a frequency of 2.3 years. The frequency of that discharge which corresponds to full channel width is unknown and need not be identical to the discharge having a 2.3-year recurrence.

It should be recognized that each of the graphical relations presented has considerable scatter owing to the nature of the measurements and the inherent variability of these factors in the field. It should not be inferred that a relation such as that presented between discharge and stream order is considered precise. Rather we are concerned with explaining a methodology by which generalized relations can be obtained, and with demonstrating the nature of the interrelations between a variety of hydraulic and physiographic factors.

In summary, it has been shown that the ephemeral streams in New Mexico are characterized by a uniform downstream increase of width, depth, and velocity with stream order, and also with drainage-basin size. Increasing size of drainage basins is accompanied by a downstream increase in discharge. Furthermore, the interrelations among all of these factors, both hydraulic and physiographic, may be expressed in simple terms, either as exponential or power functions. The method of combining the hydraulic variables with factors measured on a map or obtained in the field from a dry stream bed allows a simple means of obtaining interrelations which cannot be measured directly.

SOME RELATIONS OF HYDRAULIC AND PHYSIOGRAPHIC FACTORS TO THE LONGITUDINAL PROFILE

CHANNEL ROUGHNESS AND PARTICLE SIZE

The longitudinal profile of a river is ordinarily drawn as a graph showing elevation as a function of horizontal distance from the headwaters. Hack (1955) showed that if slope were substituted for elevation as the ordinate, the gradual downstream decrease in stream gradient makes the graph a straight line on logarithmic paper. Furthermore, the slope of this line is a measure of the concavity of the profile.

Hack demonstrated for several streams in the Shenandoah Valley that the rate of decrease of bed-material size in the downstream direction is related to the degree of profile concavity. The greater rates of particle-size decrease downstream are associated with the more concave profiles.

The relation of channel characteristics to the longitudinal profiles of streams was discussed in an initial way by Leopold and Maddock (1953). They stated that under most conditions in a river there is a tendency for channel roughness to be conservative and to change downstream less rapidly than some of the other hydraulic variables. They say (1953, p. 51) that "the suspended load and its change downstream characteristic of natural rivers require a particular rate of increase of velocity and depth downstream. Under the conditions of a nearly constant roughness, to provide the required velocity-depth relations, slope must generally decrease downstream; it is for this reason that the longitudinal profile of nearly all natural streams carrying sediment is concave to the sky."

The data presented in the present paper provide some indication of how the channel factors are related to the changing size of bed material, or more specifically, how the idea expressed by Leopold and Maddock can be integrated with the later work of Hack to provide a more nearly complete picture of the interaction of geologic and hydraulic factors.

The hydraulic variables slope and roughness can be related to the other factors in an approximate way by considering a Manning-type equation,

$$v=1.5\,\frac{d^{\frac{2}{3}}s^{\frac{1}{2}}}{n'}$$

in which d is mean depth (approximately equal to hydraulic radius), s is slope, and n' is a roughness factor. It is necessary in our data to use the slope of the stream bed as an approximation to the slope of the energy grade line. Let n' be the corresponding roughness parameter. If these factors are expressed as functions of discharge in the form used by Wolman (1954a)

$$s=tQ^{z}$$

$$n'=rQ^{y}$$

and substituting the function of discharge (equations 7–12) for each variable, then

$$kQ^{m}=\frac{1.5(cQ)^{\frac{2}{3}f}(tQ)^{\frac{1}{2}z}}{rQ^{y}}$$

it follows that

$$m=\tfrac{2}{3}f+\tfrac{1}{2}z-y$$

If the downstream change of width is related to discharge as $w\propto Q^{.5}$ (shown to be a characteristic of all streams studied), because $b+f+m=1$, then $m+f=0.5$. Thus, when m increases, f decreases.

The equation

$$m=\tfrac{2}{3}f+\tfrac{1}{2}z-y$$

is represented graphically by the lines in figure 22, assuming that $m+f=0.5$. The average relations of the hydraulic variables as functions of Q in the downstream direction give values of $f=.4$ and $m=.1$. For the

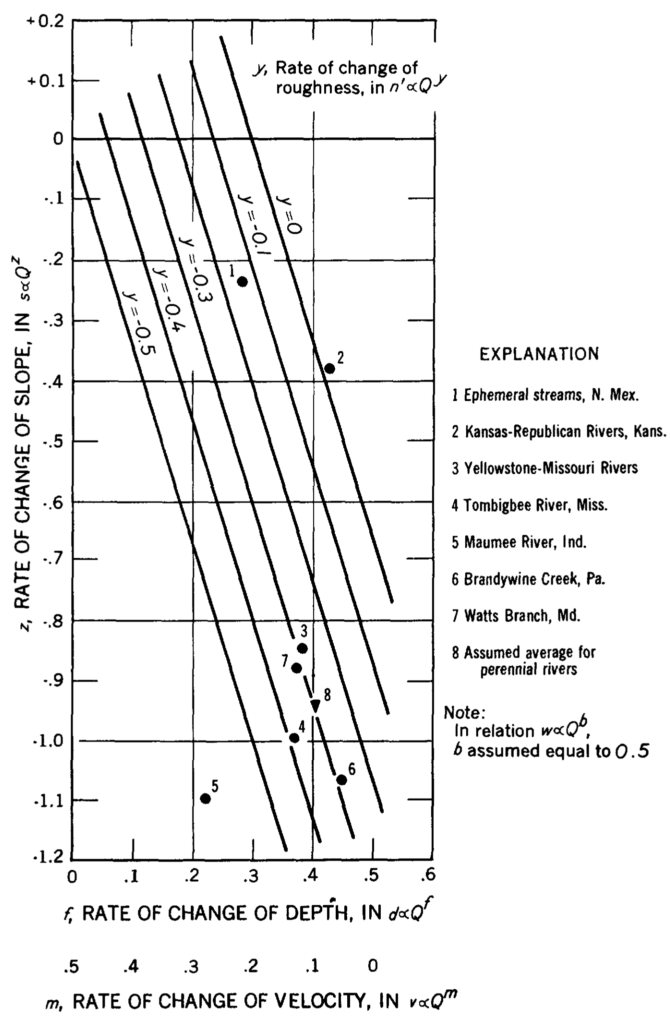

FIGURE 22.—Downstream relations of depth, velocity, slope, and roughness expressed as values of exponents in equations relating these factors to discharge.

average value of z, Leopold (1953, p. 619) suggested $-.49$, but analysis of additional data leads us to revise this to $-.95$.[1] The average downstream relations, therefore, can be represented by point 8 marked with a triangle in figure 22. This indicates that the average value of y is -0.3, which means that channel roughness decreases slightly downstream for discharge of constant frequency. This apparently is related to some progressive downstream decrease in bed-particle size resulting from sorting and abrasion.

From the graph it can be seen that if width, depth, and velocity changed downstream at rates respectively equal to those in the average river ($b=.5, f=.4, m=.1$) and if roughness (measured by Manning-type n') remained constant downstream ($y=0$), then slope would decrease downstream with increasing discharge in the form

$$s \propto Q^{-.33}$$

In few natural rivers, however, does roughness remain constant downstream.

Values of f and z for several rivers are plotted in figure 22, and the corresponding values of the exponent y, expressing rate of change of roughness, can be read from the scale given by the sloping lines. Point 2, representing the Kansas-Republican River system, indicates a near-average value for rate of change of depth ($f=0.43$). The nearly constant value of roughness downstream ($y=0$) in that river system is presumably related to the fact that bed-particle size decreases a relatively small amount, ranging from sand near the headwaters to silt farther downstream. Consequently, the river gradient decreases downstream less rapidly than in the other rivers; that is, the value of z is about $-.4$, whereas most of the other rivers have values between $-.8$ and -1.1.

The headwaters of the Yellowstone and Bighorn Rivers rise in coarse gravel which decreases to fine sand in the main Missouri, and this river has a correspondingly rapid change of gradient as indicated by the large negative value of z in figure 22. Thus it appears that an important effect of decreasing particle size downstream is to decrease the roughness downstream.

EFFECT OF CHANGE OF PARTICLE SIZE AND VELOCITY ON STREAM GRADIENT

Channel roughness is not determined entirely by particle size. That our roughness factor is not necessarily proportional to the sixth root of bed particle size—as would be expected from the Strickler equation $n=ck^{1/6}$ (see O'Brien and Hickox, 1937, p. 314)—follows from the fact that n' involves not only the grain roughness but also resistance attributable to channel configuration and other factors. However, for load composed of coarse sand or gravel, channel roughness is materially influenced by the size of grains. The manner in which particles of finer materials are piled into dunes and riffles becomes an important determinant of channel resistance.

For a given rate of change of depth and velocity, the rate of decrease of stream slope (concavity of the profile) is adjusted to the rate of decrease of particle size as expressed by roughness. In figure 22 this may be seen by comparing points 2 and 6. Each has a value of f approximately equal to 0.45; that is, in these two streams the downstream rates of increase of depth with discharge are nearly identical. River 2 is characterized by nearly constant roughness downstream ($y=0$), whereas river 6 has a rapid downstream decrease of roughness ($y \approx -0.3$). Likewise, the slope of river 6 decreases much more rapidly downstream ($z \approx -1.1$) than river 2, ($z \approx -0.4$).

During the initial field work in New Mexico particle-size measurements of bed material were made by sieving scoop samples. The results were quite inconsistent owing to the large range of size over the bed even within short reaches. In 1954 the senior author returned to the field area with M. Gordon Wolman to obtain new measurements. Using a grid-system sampling method (described by Wolman, 1954b), bed-material size was measured at those reaches where discharge measurements had been obtained earlier, and these data are summarized in appendix D.

These data were plotted against discharge corresponding to the stream order at the places of measurement, and the graph is presented as figure 23. The line drawn through the points shows the downstream relation of discharge to grain size, and can be expressed as

$$D_{50} \propto Q^{-.08}$$

where D_{50} is the median grain size; that is, 50 percent of the grains are finer.

The streams in central New Mexico are characterized by the following set of values: $f=0.3$, $z=-0.24$; and the coordinates of these values in figure 22 are indicated

[1] Other ways of deriving an approximate value for this exponent can be obtained from the equation presented by Hack (1955) relating stream length, l, and slope, s,

$$s \propto l^{-1}$$

and his equation relating length to drainage area

$$l \propto A_d^{0.6}$$

Combining these with the generalization typified by figure 21 that bankfull flood discharge

$$Q_{2.3} \propto A_d^{.70}$$

then

$$s \propto Q_{2.3}^{-0.8}$$

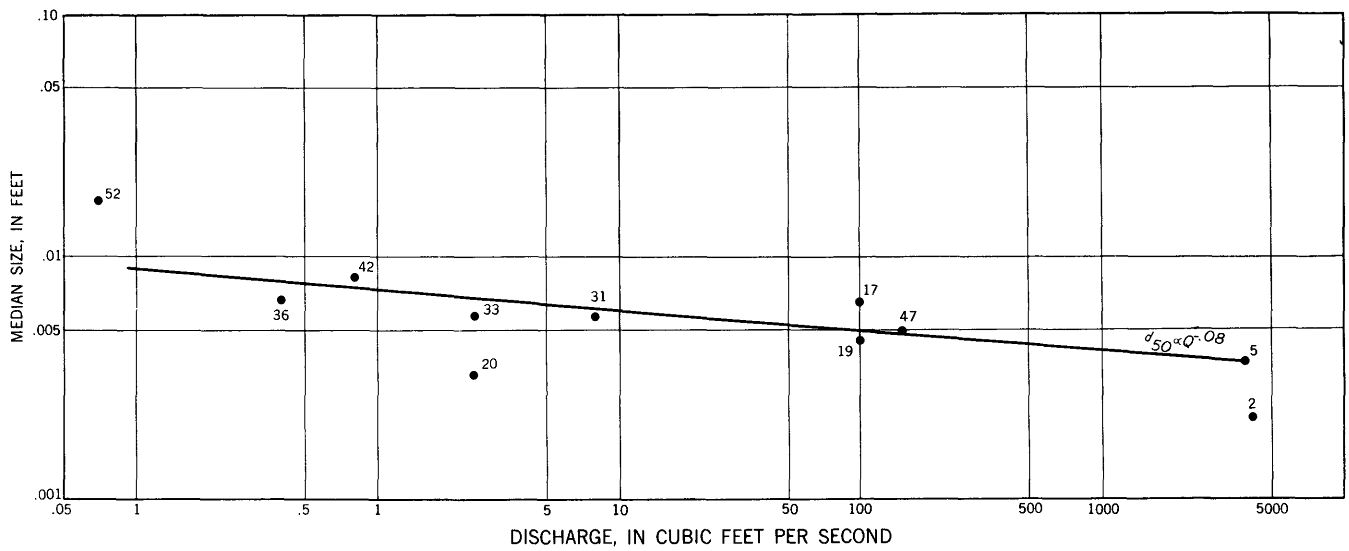

FIGURE 23.—Downstream change of median size of bed particles with discharge, ephemeral streams near Santa Fe, N. Mex. Discharge determined from discharge-order relation, solid line, figure 20. Number beside point is location serial number in appendixes A to D.

by point 1. All these streams had only a small decrease of particle size downstream, and if particle size alone controlled roughness, point 1 would be expected to fall near the line $y=0$ (constant roughness). That the data indicate some downstream decrease of roughness ($y \approx -0.12$) presumably indicates that in addition to the slight downstream increase in particle size, there is also some effect of bed configuration. Possibly also width-depth ratio itself tends to influence channel resistance.

The manner in which the analysis given above verifies and integrates previous work of Leopold and Maddock (1953) and Hack (1955) can now be seen. In figure 24 the relation between stream gradient and drainage area is presented for a selected set of streams in the United States, including both ephemeral and perennial streams representing both eastern and western conditions. The two examples from Hack are for two eastern streams having different rates of change of bed-material size. In Calfpasture River the size of bed material is constant downstream; in Gillis Falls Branch the size of bed material increases downstream. Comparison of the graph for ephemeral streams with those for Calfpasture River and Gillis Falls Branch would lead one to surmise that the New Mexico streams have an increase in particle size in the downstream direction. However, our measurements showed that particle size tended to decrease slightly downstream.

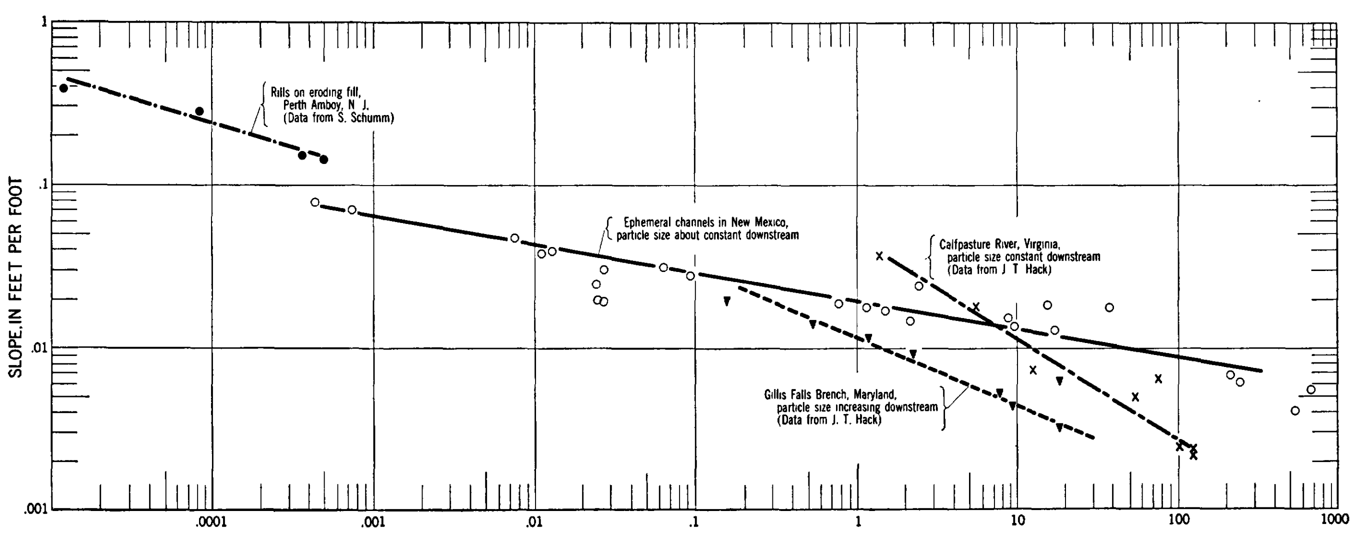

FIGURE 24.—Downstream change of stream gradient in selected drainage basins.

Other very small ephemeral streams are similar to the New Mexico example in that the slope of the graph is relatively small compared with that of Calfpasture River. Note the line representing the rills developed on an artificial fill at Perth Amboy, N. J., which was studied by Schumm.[2] Although analyses of particle size on Perth Amboy streams are not available, Schumm's report implied that particle size tended to remain more or less constant or decrease slightly in the downstream direction.

The difference in slope of the lines representing Calfpasture River and the New Mexico streams in figure 24 is associated with difference in rates of downstream change of channel characteristics. It is presumed that in the ephemeral streams, the small value of z (slope of the line in the gradient-discharge relation) is related to the downstream increase in sediment concentration reported in a previous section of this paper. It is reasoned that to maintain a quasi-equilibrium with this increase in sediment concentration downstream, there must be an accompanying downstream increase of velocity somewhat larger than that which characterizes normal rivers. As can be seen in figure 22, the more rapid downstream increase in velocity (large value of m) is provided by a relatively small value of z for any given value of the rate of change of roughness (value of y).

In summary, the longitudinal profile of a river is related to a combination of geologic and hydraulic factors. These hydraulic factors are, first, the rate of change of particle size of bed sediment downstream, which tends to govern the change of channel roughness; and, second, the relative rate of increase of velocity downstream, which appears to be controlled by the rate of change of suspended-sediment concentration downstream. Other things being equal, the less concave profiles are characterized by greater rates of increase of suspended-sediment concentration downstream or by less rapid decrease of bed-particle size in the downstream direction.

EQUILIBRIUM IN EPHEMERAL STREAMS

MUTUAL ADJUSTMENT OF HYDRAULIC FACTORS

As has been discussed previously (Mackin, 1948; Leopold and Maddock, 1953), the factors which may be considered independent of the channel itself are discharge and quantity of sediment load, and to some degree, caliber of load. Slope, width, velocity, and depth are determined by or altered by forces operating within the channel. In the words of Rubey (1952, p. 131), the dependent variables "are determined not separately but jointly"; that is to say, when circumstances require or impose a change in one or more of the factors, the others will become adjusted to maintain a quasi-equilibrium between forces of erosion and forces of deposition.

That some of the forces operating within the channel may be determined by previous history is generally recognized. There may be a difference in the "degrees of freedom" characteristic of the different hydraulic situations. As an example, one might consider a flume with a movable bed. If the sides of the flume are made of wood or metal so that the width is not adjustable, there must be an adjustment among all the remaining variables to a condition of mutual accommodation. The fixed width may be thought of as the restriction of one degree of freedom in the process of mutual accommodation.

If a particular valley had been carved during the Pleistocene by a relatively large river, and through change in climate or other physiographic factors the valley at present carries only a minor streamlet, it is likely that the small stream would not be able to recarve the valley bed to such an extent that the valley slope is materially altered. In such a circumstance the minor stream, initially at least, would have to accommodate itself to the slope of the valley in which it flows.

The organization of a drainage system and the remarkable adjustment of its various hydraulic factors have been amply demonstrated. These have been discussed by several authors, Rubey (1952), Leopold and Maddock (1953), and in detail by Wolman (1954a). The present paper presents further evidence that nearly all rivers approach a condition of quasi-equilibrium which was formerly attributed only to a graded stream. This postulate implies further that if the degrees of freedom of the hydraulic parameters are reduced, the remaining factors tend toward a mutual accommodation or quasi-equilibrium. It is our purpose here to demonstrate that this reasoning can be extended even to a headwater stream in youthful topography as far as the uppermost tributary rill.

In figure 18 a line has been drawn to represent the relation between channel slope and channel width for arroyos of large and moderate size. In the upper left part of this figure the open squares represent values of slope and width of the smallest unbranched rills in the headwater tributary basin of Arroyo Caliente. These rills averaged somewhat less than 1 foot in width. Gradients were measured by laying a board in the bed of each rill and determining its slope with a Brunton compass. The measurements may be considered crude and the scatter of the points on the graph may be partly attributed to the method of obtaining the data; nevertheless, the position of those points indicates that the

[2] S. Schumm, 1954, Evolution of drainage systems on an eroding fill at Perth Amboy, N. J., Ph. D. thesis, Columbia Univ.

relation of channel slope to channel width is generally in line with that of the larger ephemeral channels.

The solid triangles apply to similar unbranched rills occurring in steeper tributaries in the same area which drained the unconsolidated gravels of a pediment remnant somewhat nearer the mountains than the Arroyo Caliente. The difference in stream slope between Arroyo Caliente and Fifth of July Wash can probably be attributed to the difference in size of the gravel characterizing the two small basins.

To determine whether this relation between slope and width would be maintained even by the very smallest rills, measurements were made of the smallest natural rills which could be found in the area. The values of slope and width of these miniature features are represented by the black circles in the far upper left part of figure 18. They fall in a position so nearly representing an extension of the line in the diagram that it may be inferred that the slope-width relation in the area studied applies as far upstream as the smallest observable rill.

In the same area are many places where modern highways have required deep road cuts through the same material as that constituting the drainage basins under study. On the steep road cuts little or no vegetation has become established, and numerous steep parallel rills of an average depth of 0.2 feet have developed. The width and slope of these road-cut rills were measured. The slope had changed but little from that determined by the blade of the highway grader. The rills on road cuts are shown by open triangles on figure 18. These points are no more scattered from the mean line than any other data. It appears then, that if the degrees of freedom are reduced, in this instance by a prescribed slope, the width of the rill will be formed in accordance with the slope-width relations for natural streams having a larger number of degrees of freedom. This is interpreted as further evidence that channels cut by water carrying sediment tend to maintain a quasi-equilibrium even as far upstream as the most remote ephemeral rills.

THE DISCONTINUOUS GULLY

Beginning late in the 19th century the alluvial valleys of the West experienced an epicycle of erosion characterized by the development of valley trenches or arroyos (see fig. 25). These arroyos range in size from insignificant rills to canyons 600 feet wide, 50 feet deep, and 150 miles long. The notorious Rio Puerco (del Oriente) in New Mexico has the latter dimensions. Some gullies are narrow enough to step across but are deep enough to lose a giraffe in. As Gregg (1844, p. 184) said more than a century ago,

FIGURE 25.—A typical large continuous arroyo trenching an alluvial valley in New Mexico: Rio Puerco (del oeste) near Manuelito, looking downstream.

> The sides are usually perpendicular—indeed, often shelving at the base, and therefore utterly impassable . . . Though, to a stranger, the appearance would indicate the very head of a ravine, I would sometimes be compelled to follow its meandering course for miles without being able to double its "breaks."

It is characteristic of the large arroyos that depth remains quite uniform through very long reaches. This uniformity of depth means that the gradient of the channel bed had attained a slope almost parallel to the original valley floor. There is also another distinctive type of gully that is characterized by a vertical head-cut, a rapidly decreasing depth downstream, and a fan at the lower end. Such channels generally occur in groups arranged irregularly along the length of the drainageway, and because of this characteristic are called discontinuous gullies.

Bryan (1928) has presented evidence that when the Rio Puerco was first reached by the reconnaissance teams of the Army of the West in 1846, the valley floor was already being dissected by discontinuous gullies. Some of them were evidently large, for in August 1846, Lt. Simpson had to cut down the gully wall of the Rio Puerco in order to get his brass cannon across. Yet, at the same time, there were long reaches in the Puerco valley so smooth that the native grass was cut for hay, and water was diverted from the channel by felling a cottonwood tree to form a dam.

Discontinuous gullies have long presented a problem to the erosion-control engineer. In the first place, the mechanics of gully formation is very poorly understood. Although it is generally presumed that discontinuous gullies, at least in places, can coalesce and form a continuous channel, it is not known whether the nature of formation of discontinuous and continuous arroyos is different. If the mechanics of the two are appreciably different, the choice of control measures presumably would be affected.

The salient characteristic of a discontinuous gully is the relatively small gradient, or slope, of its bed. It is this flat slope—less steep than the floor of the original ungullied alluvial valley—that makes the gully discontinuous, for the bed profile must at some downstream point intersect the profile of the original valley floor. At that point the gully depth has diminished to zero, and the flow of the gully spreads out over the valley floor and at least part of the load is deposited in a low fan.

Therefore, it is important to find an explanation for the low gradient. Any discussion of such gullies which does not provide an explanation of the mechanics by which this flat slope is developed would miss the dominant feature. Our approach to this aspect of the problem is essentially inductive, for there is a paucity of adequate data on essential elements in the hydraulic relations. However, as will be shown later, field characteristics appear to support the conclusion reached inductively.

Consider the wide, grassy floor of an alluvial valley in the semiarid West before the recent epicycle of erosion. Though the summer rains are of short duration and intensity, the flash flows spread widely over the valley, and channel storage consequently keeps the peak discharge at any point to a small value per unit of drainage area. The grassy floor offers resistance to the flow of water and sediment contributed to the valley floor from adjoining slopes. This resistance keeps the velocity low and thus the sediment concentration is small. The profile and cross section of such an ungullied valley floor are shown as diagram 1 of figure 26.

Assume that local weakening of vegetation allows an initial furrow, scarplet, or small basin to form by erosion. The cause of this lowered resistance to erosion could be grazing or trampling by stock, fire, or an exceptional local storm. The regional erosion in the West, which began about 1880, is considered to be a result of a combination of overgrazing and meteorologic shift.

Subsequent storms cause the head of this initial erosion feature to progress up-valley, and the debris excavated splays out at the downstream toe in the form of a low fan. As soon as a short channel is formed, terminating in a vertical head-cut, the concentration of water in the flumelike trench reduces channel storage. Consequently, from a storm of given size, the peak discharge passing through the channel is greater than would have been experienced on the ungullied valley floor. This increased peak discharge is accompanied by greater velocity and cutting power, and the initial gully advances so rapidly during storms that growth of vegetation in the intervals between storms cannot heal it.

Water pours over the lip and develops a plunge pool at the toe. The original sod, even when weakened and incomplete as a protective cover, keeps the lip of the head-cut relatively stronger than the underlying alluvium and the latter is cut away by the turbulence in the plunge pool, leaving the root-bound lip overhanging the plunge pool. Undercutting by plunge-pool action during storm flow is greatly aided by—and perhaps is even less important than—slumping of the moistened headwall after the storm flow ends. This slumping is promoted by piping holes (see fig. 5) which develop on a diagonal line between the lower part of the vertical headwall and the valley floor many feet upstream from the head-cut. We have often observed the upper entrance to these piping holes on the swale floor 50 feet from the head-cut. Piping permits water to penetrate deeply into the material into which the head-cut is eating, and this moisture aids the process of sapping and slumping at the base of the headwall.

The plunge pool is always seen to be dug deeper than the level of the floor of the discontinuous gully just downstream from the plunge pool. This clearly means that the floor of the discontinuous gully is composed of a layer of newly deposited material which overlies the undisturbed alluvium. This deposit is laid in immediately below the plunge pool, and deposition proceeds upstream as fast as the plunge pool does.

These details of gully growth are important to an understanding of the mechanics, for they show that when a plunge pool is present, and it usually is, the flat slope of the discontinuous gully is a *grade of deposition*. Furthermore, these details point up a significant difference between the action of flood flow in the channel of a normal river and that of a developing gully. The plunge-pool action tends to deepen the channel faster than to widen it. In the early stages of gully development, then, the channel is relatively narrow and deep.

At any time, the gully floor is built at such a slope that the material coming into the reach will be transported through the reach under the particular condition of roughness. At the early stage of gully development now under consideration, the channel has considerable depth but a restricted width. At the same time it is forming a deposit just downstream from the plunge pool and thus is currently forming its bed slope under conditions in which slope can be adjusted with relative rapidity, as compared to width. Under conditions which prevail in ordinary rivers the reverse is true; width adjusts rapidly during floods, but because of the large amounts of material involved in appreciably changing slope through long reaches of river, slope adjusts only slowly.

Analysis of river data showed that at a given discharge, the same suspended-sediment load will be carried by a narrow, deep channel at a particular velocity as in a wider, shallower channel at somewhat higher velocity (see fig. 15, p. 23, in Leopold and Maddock, 1953). The early stage of a discontinuous gully is considered to represent the narrow, deep condition, and at a later stage, lateral cutting and slumping of the arroyo walls lead to widening, with a consequent shallowing of the cross section of flowing water at a particular discharge. Now, under conditions of a given and constant roughness, a relatively large depth and small velocity requires a small value of slope; that is, a flat gradient, as can be seen by considering the Manning-type equation

$$v=1.5\,\frac{d^{\frac{2}{3}}s^{\frac{1}{2}}}{n'}$$

As widening progresses during a later stage in gully development, to carry a particular sediment load at a given discharge requires a slightly increased velocity as the width-depth ratio increases. To achieve the larger velocity with decreasing depth at a given roughness means an increase in the value of slope.

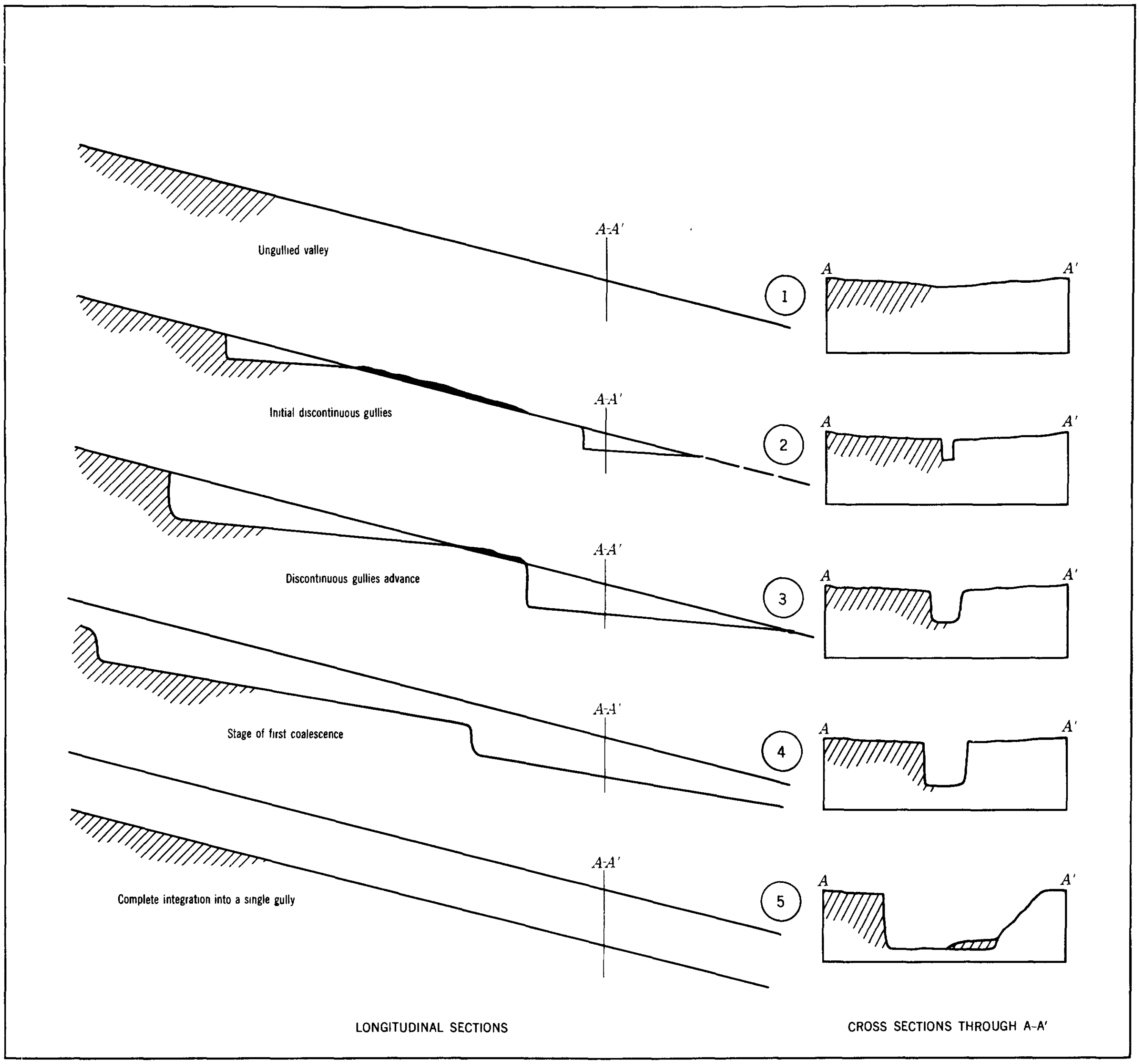

FIGURE 26.—Stages of development of an arroyo from discontinuous gullies.

This line of reasoning which follows the principles derived from interrelation of width, depth, velocity, and suspended load at constant discharge in natural rivers does provide an explanation for the small gradient of the channel bed in the early and narrow stage of the discontinuous gully. It indicates why an increase in bed slope could be expected to be associated with the channel widening.

The principles outlined apply to the changing relations at constant discharge, suspended load, and roughness. Before the individual discontinuous gullies coalesce to form a continuous, flumelike channel, there is still sufficient channel storage in the ungullied reaches of the valley to attenuate flood peaks originating far upstream. Thus, the most effective flows in a given reach probably are derived from storm rainfall in the immediate vicinity of the reach in question. The suspended load in the reach is derived from the bank caving and headcutting of the individual discontinuous gully. In the early stages of gully development these conditions justify the use of the assumption of constant discharge and suspended load. Constant load and discharge imply constant sediment concentration which is one of the important determinants of roughness. So, the assumption of constant roughness is logical.

In developing an argument to explain the low gradient associated with the narrow and deep channel of the early stages of a discontinuous gully, it is inappropriate to reason from the downstream relation of slope to width shown in figure 18, for there is implicit in that diagram an increase in discharge associated with increasing width downstream. The stages in development of the discontinuous gully must be explained in terms of changes in channel characteristics which can take place at constant discharge.

In summary, it is reasoned that mutual accommodation causes the small width of the initial discontinuous gully to be associated with the development of a small slope, in fact so much flatter than that of the original valley floor that the two gradients intersect (diagram 2, fig. 26). The depth rapidly decreases downstream to the point of extinction, and below that point a small fan develops where the sediment is spread on the original valley floor.

These same forces operate to expand the gully headward and to deepen it, the combination of which makes the gully increase in total length. Because there are usually several gullies along the valley floor, the toe of one gully tends to extend itself downstream while the head-cut of the succeeding one is extending itself up-valley, decreasing the total length of unchanneled valley floor between them. This change is indicated by diagram 3 of figure 26.

As the process continues through time, a stage will be reached when the head-cut of the downstream gully meets the toe of the one upstream and the two discontinuous gullies coalesce. This stage is indicated by diagram 4. The lengthening and deepening of the gully is accompanied, however, by a gradual increase in width. As both width and depth continue to become greater, the channel reaches a size sufficient to contain without overflow on the valley floor the largest discharge which the basin upstream can produce. At this stage the effect of the large amount of channel storage which characterized the flat-floored alluvial valley in its ungullied condition has been lost and the discharge in a given reach increases greatly.

Finally a condition is reached in which the total width between the arroyo walls is sufficient to allow the stream to wander back and forth between the walls and the development of local patches of flood plain can begin. At that stage the gradual increase in width has required such an increase in slope that the gully bed becomes almost parallel to the original valley floor (diagram 5). This is characteristic of most of the large arroyos throughout the West. There may still remain a difference, however, between the total width of even a large gully and the width of flowing water characteristic of the condition when the ungullied valley floor carried the same discharge. Probably it is for this reason that through long reaches in the down-valley direction the depths of some arroyos gradually decrease. The hypothesis implies that gullies in the process of development could generally show the features indicated diagrammatically on figure 26.

To test this hypothesis several gullies in the Santa Fe area were mapped. The profile of figure 27 is a typical example. This gully is discontinuous, as can be seen by the fact that its depth becomes progressively shallower downstream to point of extinction at a distance of 2,750 feet. Immediately upstream from this point the gradient of the gully floor was .014, a much smaller figure than the original gradient of the valley floor, .028. Some deposition in the form of a fan can be seen between 2,750 and 3,200 feet on the distance scale, but only a small proportion of the total evacuated materials is ever deposited in the fan below a discontinuous gully; most of it proceeds downstream. The situation downstream from point 2,300 feet indeed bears a marked similarity to that shown in diagram 2 of figure 26.

At 2,240 feet it can be seen that the gully bed drops more than 4 feet, and this is certainly the point of coalescense of two formerly separated discontinuous gullies, exactly as shown in diagram 4 of figure 26.

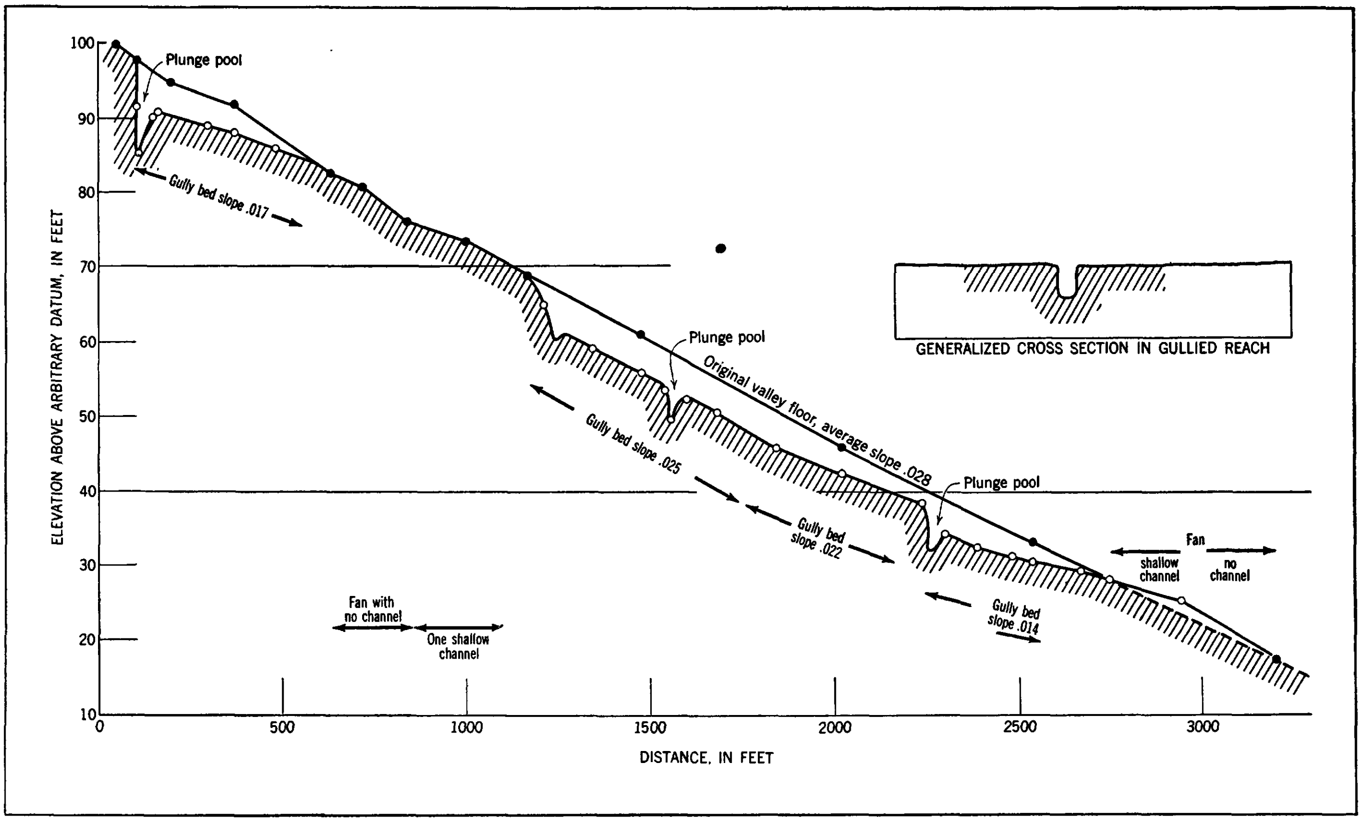

FIGURE 27.—Profile of a discontinuous gully, Arroyo Falta, near Las Dos, 6 miles northwest of Santa Fe, N. Mex.

To extend the analogy between field conditions and the postulated stages of development, it was noted that another gully nearby and parallel to the one shown in figure 27 could be characterized by diagram 5 (fig. 26) in that the depth remained nearly uniform and width had developed to a point where small patches of flood plain were beginning to appear within the gully walls.

So it appears that the discontinuous gully changes progressively through a series of stages to be transformed into a continuous gully of nearly uniform depth. The early stages are characterized by relatively flat slope of the discontinuous gully bed, but slope gradually increases with increasing channel width until the slope becomes nearly equal to that of the original ungullied valley floor.

The relation between the hydraulic variables and the tendency for establishment and maintenance of a quasi-equilibrium, even in headwater ephemeral channels, provide a logical explanation for the development of continuous gullies from discontinuous ones. Without the hydraulic argument, the transition from discontinuous to continuous channels could not be explained.

SUMMARY

The broad, sandy, and usually dry washes, and the trenchlike arroyos of the semiarid Southwest appear to be the antithesis of the perennial rivers of the humid East. Probably the most striking fact which emerges with the collection and analysis of hydraulic data pertaining to these apparently different channel systems is the similarity between them. Differing in detail, to be sure, the interrelations among hydraulic and physiographic factors in ephemeral and perennial channel systems are generally similar. Differences observed are in the direction which might be expected from the hydrologic and physiographic character of the areas, and for this reason the observed differences in values of hydraulic factors can be assumed valid.

The same forces or processes operating in a river channel which lead to the condition of grade also exist in channels which are not graded. This leads to a tendency for adjustment among hydraulic parameters toward a condition of quasi-equilibrium between transporting capability and load delivered to the channel system. Even when complete adjustment is not achieved, the disequilibrium is spread through relatively long reaches of channel. Furthermore, when one of the ordinarily adjustable hydraulic factors is fixed or invariant, the other factors mutually adapt toward quasi-equilibrium under the prevailing conditions.

The ephemeral streams studied provide examples of each of these general situations. These channels are probably downcutting slowly, for they are located in the headwater parts of their drainage areas where over short periods geologically, land reduction is in progress. Thus the streams do not fulfill the requirement of stability which the term "grade" usually implies (Wolman, 1954). Yet, the channels are adjustable. These ephemeral streams in the Southwest are characterized by increasing concentration of suspended sediment in the downstream direction, which results from loss of water by infiltration to the channel bed and to some lesser extent, from progressive pick-up of load from the stream bed. In response to this downstream change in sediment concentration, channel factors mutually adjust themselves toward quasi-equilibrium. One observable aspect of this adjustment is a greater downstream increase in velocity than is characteristic of normal perennial rivers.

These same processes of adjustment lead to a progressively increasing width and decreasing gradient downstream in a nearly uniform manner, not only in the larger streams but headward as far as the uppermost rills. As an example of adaptation to a temporarily fixed or predetermined value of one hydraulic factor, the width-slope relations in road-cut rills were investigated. It was shown that on a road-cut where slope was determined by a road grader rather than by natural processes, rills developed a width which fits the width-slope relations in natural ephemeral rills and channels in the same locality.

The same general principles were used to analyze the probable interrelation between hydraulic factors in gullies or discontinuous arroyos. This leads to the conclusion that the mechanics of gully development involve the same tendency for mutual adjustment of hydraulic factors, even in a channel undergoing progressive erosion. The analysis suggests that measurements of flash flows in active gullies and suspended-sediment load would be pertinent, even essential, to an improved understanding of the manner and rate of gully erosion and effective measures for gully control.

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APPENDIX

APPENDIX A.—*Basic data on ephemeral channels near Santa Fe, N. Mex.*

Point No.	Stream and location	Drainage area (square miles)	Width (feet)	Slope (feet per foot)	Order (adjusted)	Length from divide (miles)
1	Rio Galisteo, at mouth	672	615	0.0055	11	54
2	at Hwy. 85	540	275	.0041	11	46
3	at Lamy	--------	240	.0074	--------	--------
4	at Galisteo	--------	160	.0085	--------	--------
5	at Cerrillos	--------	180	.0059	--------	--------
6	Arroyo San Cristobal, near Galisteo	--------	180	.007	--------	--------
7	at ruin (above Hwy. 285)	--------	100	.0071	--------	--------
8	Arroyo La Jara, 2½ miles south of Galisteo	--------	90	.007	--------	--------
9	Apache Canyon, near mouth at Canoncito	17	42	.0132	9	9.9
10	2.2-square-mile wash, 1 mile west of Canoncito	2.2	36	.0150	--------	--------
11	Rio Santa Fe, at mouth	240	230	.0062	--------	--------
12	at La Bajada	210	125	.0069	--------	--------
13	½ mile below mouth of Arroyo de los Chamisos	--------	120	.0054	--------	--------
14	at mouth of Arroyo de los Chamisos	--------	50	.0072	--------	--------
15	at mouth Arroyo Calabasas	--------	325	.0087	--------	--------
16	at Old Albuquerque Road	--------	--------	.0151	--------	--------
17	Arroyo de los Chamisos, at Old Pecos Road (near Mt. Carmel Chapel)	2.5	40	.025	8	--------
19	Cañada Ancha at Hwy. 84, near El Rancho Montoso, Seton Village quadrangle	1.52	90	.0175	8	2.2
20	Ancha Chiquita, near Hwy. 84, 0.4 mile north of El Rancho Montoso	.093	22	.0286	5	.7
21	Arroyo de los Frijoles, in NW¼ SW¼ sec. 17, T. 17 N., R. 9 E.	--------	100	0.0177	--------	--------
22	at Connell Ranch ½ mile above mouth of Arroyo de las Trampas	--------	80	.0125	--------	--------
23	at "Short of Sand Reach" at mouth of Arroyo de las Trampas	--------	90	--------	8	--------
24	Pojuaque Creek, near San Ildefonso	--------	300	.008	--------	--------
25	Ancha Arroyo, near Cuyamungue	8.85	298	.0155	--------	--------
26	Cañada Ancha, ½ mile below mouth Las Dos Arroyo	37.4	290	.0185	10	10.2
27	Las Dos Arroyo, near mouth, lat. 35°48′, long. 106°06′30″	15.1	219	.019	9	7.7
28	Alamo Creek, at NE¼SE¼ sec. 33, T. 18 N., R. 8 E.	9.72	98	.014	8	6.1
29	Cañada Ancha, in NW¼ sec. 33, T. 18 N., R. 8 E.	.27	18	.031	7	.75
30	Hermanas Arroyo, near Las Dos, SE¼SW¼ sec. 31, T. 18 N., R. 9 E.	1.16	32	.018	7	2.85
31	Tributary to Hermanas Arroyo, near Las Dos	.063	10	.032	6	.3
32	Lagunitas Arroyo, at Las Dos, SE¼ sec. 30, T. 18 N., R. 9 E.	.79	20	.019	8	1.3

APPENDIX B.—*Arroyo Caliente* [1]*: channel data*

Point No.	Location	Drainage area (square miles)	Width (feet)	Slope (feet per foot)	Order (adjusted)	Length from divide (miles)	Length from divide (feet)
33	Station 19	0.027	16	0.0195	5	0.272	--------
34	Do 17	.025	6.5	.020	--------	.238	--------
35	Do 16	.024	11	.0255	--------	.218	--------
36	Do 15	.0018	2.6	--------	3	--------	--------
37	Do 13	.019	8.4	--------	--------	--------	--------
38	Do 12	.013	7.2	.0405	--------	.189	--------
39	Do 11	.0019	2.8	--------	--------	--------	--------
40	Do 9	.011	10	.0395	--------	.121	--------
41	Do 8	.0076	5.6	.048	--------	.102	--------
42	Do 6	.0026	5.4	--------	4	--------	--------
43	Do 4	.0022	3.8	--------	--------	--------	--------
44	Do 3	.00043	2.8	.08	2	--------	--------
45	Do 2	.00072	2.0	.071	2	.042	--------
46	Do 1	.00058	2.0	--------	--------	--------	250
47	Headwater rill at random location in arroyo basin.	.00021	1.0	--------	--------	--------	130
48	Headwater rill at ran-	0.00021	0.5	--------	--------	--------	380
49	dom location in arroyo-	.00021	1.6	--------	--------	--------	450
50	basin—Continued	.00040	.8	--------	--------	--------	535
51	Do	.00054	1.5	--------	--------	--------	650
52	Do	.00047	.8	--------	1	--------	--------
53	Do	.0016	2.0	--------	--------	--------	810
54	Do	.0018	2.2	--------	--------	--------	1,010
55	Do	.00093	2.3	--------	--------	--------	1,190
56	Do	.0013	1.5	--------	--------	--------	--------
57	Do	--------	.9	--------	1	--------	--------
58	Do	--------	.8	--------	1	--------	--------
59	Do	--------	1.4	--------	3	--------	--------
60	Do	--------	3.1	--------	3	--------	--------
	Do						
	Do						
	Do						

[1] Arroyo Caliente is a tributary (not named on Geological Survey maps) to Arroyo de los Frijoles, in SW¼ NE¼ sec. 17, T. 17 N., R. 9 E., within the limits of Aqua Fria quadrangle, New Mexico.

APPENDIX C.—*Flood-frequency data for streams in central New Mexico*

Point No.	Stream and location	Drainage area (square miles)	Mean annual flood (cfs)	Years of record
61	Embudo Creek, at Dixon	305	[1] 1,800	28
62	El Rito Creek, near El Rito	52	[1] 350	20
63	Rio Ojo Caliente, at La Moderna	344	1,450	16
64	Rio Hondo, at Arroyo Hondo	70	[1] 320	42
65	near Valdez	38	[1] 235	36
66	Rio Santa Cruz, at Cundiyo	86	[1] 300	31
67	Willow Creek, at Park View	193	[1] 1,550	20
68	Rio Chama, near Chamita	3,200	5,000	17
69	Latir Creek, near Cerro	10	[1] 62	20
71	Nambe Creek, near Nambe	37	[1] 270	20
72	Rio Tesuque, near Santa Fe	11	[1] 120	20
73	Rio Pueblo de Taos, near Taos	56	[1] 200	42
74	Santistevan Creek, near Costilla	3.8	[1] 11	20
75	Rio Puerco, at Rio Puerco	5,160	12,600	18

[1] Data from unpublished compilations by H. H. Hudson.

APPENDIX D.—*Distribution of bed-particle size in selected channels*

Point No.	Stream and location	Percent finer than (millimeters)— 0.25	0.50	1	2	4	8	16	32	64	128	256	50 percent finer than (feet)—	84 percent finer than (feet)—
20	Ancha Chiquita northeast of Hwy. 84, 0.4 mile north of El Rancho Montoso	2	14	48	78	87	93	97	98	98	100	--------	0.0034	0.010
17	Arroyo de los Chamisos, at Old Pecos Rd. (near Mt. Carmel Chapel)	--------	6	26	52	72	86	94	97	99	100	--------	.0066	.023
19	Cañada Ancha, at Hwy. 84 nr El Rancho Montoso	9	17	39	59	83	93	96	98	100	--------	--------	.0048	.013
5	Rio Galisteo, at Cerrillos	27	42	46	60	67	79	86	94	101	--------	--------	.0039	.043
31	Tributary to Hermanas Arroyo, near Las Dos	1	11	37	54	70	81	89	97	99	100	--------	.0052	.033
16	Rio Santa Fe, at Old Albuquerque Rd	1	4	34	61	68	76	90	96	100	--------	--------	.0046	.039
76	Rio Galisteo, at Domingo	11	36	71	91	96	96	100	--------	--------	--------	--------	.0022	.005
33	Arroyo Caliente, 5th order, near mouth	1	7	23	56	82	90	94	96	100	--------	--------	.0057	.015
42	4th order	--------	5	23	46	54	59	72	79	90	99	100	.0083	.140
36	3rd order	--------	--------	33	52	61	63	71	78	88	96	100	.0066	.150
52	1st order	--------	7	21	34	45	61	72	88	99	100	--------	.0174	.085
47	Arroyo de los Frijoles, at mouth of Arroyo Caliente	4	11	28	62	80	92	95	98	100	--------	--------	.0050	.016
2	Rio Galisteo, at Hwy. 85	6	22	70	89	91	92	94	97	99	100	--------	.0022	.005

APPENDIX E.—*Arroyo discharge measurements during flash-flood flow at particular cross sections in New Mexico and Wyoming*

Stream and location	Date	Time [1]	Discharge Q (cfs)	Width w (feet)	Cross-sectional area (square feet	Mean depth d (feet)	Mean velocity v (fps)	Suspended load L (tons per day)	Suspended load concentration C_s (ppm)	At-a-station values b	f	m	j	Slope (feet per foot)	Remarks
Meet John Wash, near Casper, Wyo.[2]	Aug. 2, 1953	5:26	3.4	4.5	1.35	0.3	2.5	86	8,000	0.22	0.33	0.45	1.4	--------	L computed for Q interpolated between individual measurements.
		5:33	2.3	4.0	.9	.22	2.6	38	4,100						
		5:35	1.14	3.6	.61	.17	1.9	--------	--------						
		5:38	.8	3.5	.51	.15	1.6	10	3,270						
		5:41	.31	2.7	.38	.14	.82	5.3	2,460						
		5:46	.14	2.1	.22	.10	.64	.9	1,700						
Cañada Ancha, near El Rancho Montoso, N. Mex.	July 26, 1952	1:30	40	58	9.0	.16	4.4	5,800	--------	.39	.23	.28	1.27	0.0175	Estimated peak.
		--------	173	100	35.5	.35	4.9	62,000	--------						
		1:40	2.8	18	1.6	.09	1.8	260	--------						
Ancha Chiquita, near Rancho Montoso, N. Mex.	Aug. 9, 1952	3:05	21.5	14	4.7	.34	4.6	--------	--------	.13	.61	.32	--------	.0286	
		3:08	11.2	13	2.9	.22	3.9	--------	--------						
		3:15	3.1	11	1.3	.11	2.4	--------	--------						
Rio Santa Fe at Old Albuquerque Road, N. Mex.[3]	July 28, 1952	2:50	126	80	34	.42	3.7	17,500	--------	.26	.45	.36	1.34	.0151	
		3:07	79	68	26	.39	3.0	8,500	--------						
		3:49	40	57	16	.28	2.5	3,600	--------						
Rio Galisteo at Cerrillos, N. Mex.	July 7, 1952	8:30	66	79	26	.32	2.6	5,200	--------	.13	.56	.29	1.23	.0058	Estimated from partial measurement.
		3:15	800	110	--------	1.3	6.5	110,000	--------						
Tributary to Hermanas Arroyo, near Las Dos, N. Mex.	June 29, 1952	3:35	2.1	7	.9	.12	2.4	155	--------	.32	.24	.43	1.36	.032	
		--------	.14	2.8	.18	.06	.7	3.9	--------						
Arroyo de los Chamisos, near Mt. Carmel Chapel, N. Mex.	July 3, 1952	5:55	120	40	22.4	.56	5.4	--------	--------	.09	.61	.26	1.08	.025	
		6:05	57	36	13.6	.34	4.2	10,500	--------						
		6:20	9.5	32	--------	.12	2.8	1,600	--------						
Rio Galisteo at Domingo [4]	--------	--------	--------	--------	--------	--------	--------	--------	--------	.32	.39	.24	1.28	.0055	Slope measured near mouth.
Rio Puerco at Cabezon [4]	--------	--------	--------	--------	--------	--------	--------	--------	--------	.34	.35	.34	1.58	--------	
Rio Puerco, near Cabezon [4]	--------	--------	--------	--------	--------	--------	--------	--------	--------	.27	.35	.36	--------	--------	

[1] All times are postmeridian except that for Rio Galisteo., 8:30 a. m.
[2] Measurements made by M. G. Wolman and Leopold.
[3] Measurements made with assistance of C. E. Stearns.
[4] U. S. Geological Survey gaging stations.

APPENDIX F.—*Arroyo discharge measurements along length of stream during individual flood (considered to be downstream relations)*

Stream and location	Date	Discharge Q (cfs)	Width w (feet)	Mean depth d (feet)	Mean velocity v (fps)	Suspended load L (tons per day)	Downstream values b	f	m	j	Slope s (feet per foot)
Sedalia Gully, near Sedalia, Colo., 30 miles south of Denver.	July 30, 1952	0.14	2.2	0.10	0.57	6.7	0.29	0.15	0.58	1.85	
		.25	2.2	.12	.92	2.9					
		.3	2.5	.11	1.1	6.8					
		1.2	3	.14	2.8	33					
		1.6	5	.12	2.5	63					
		4.1	5	.17	4.8	440					
Sowbelly Creek, near Hat Creek near Harrison, Nebr.	July 11, 1951	.03	1.1	.08	.33	2.4	.31	.20	.49	1.2	0.017–0.011
		.07	1.4	.09	.54	5.0					
		.44	2.4	.13	1.4	38					
		.74	3.0	.14	1.7	82					
		1.15	2.5	.18	2.4	140					
		1.18	4.3	.21	1.4	170					

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