MEMOIR 4

High Mountain Streams: Effects of Geology on Channel Characteristics and Bed Material

By JOHN P. MILLER

Interpretation of quantitative measurements made in the Sangre de Cristo Range, north-central New Mexico

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Symbols

(Any exceptions to these designations are specified in the text.)

a	Coefficient in relation of width to discharge
A	Area of cross-section of channel
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	Sediment size (intermediate particle diameter)
f	Exponent in relation of depth to discharge
k	Coefficient in relation of velocity to discharge
L	Stream length
m	Exponent in relation of velocity to discharge
n'	A roughness parameter
Q	Discharge in volume per unit of time
$Q_{2.3}$	Mean annual flood (equaled or exceeded on average once every 2.3 years)
r	Coefficient in relation of roughness to discharge
s	Slope of water surface
S_0	Trask sorting coefficient
t	Coefficient in relation of slope to discharge
v	Mean velocity, defined as quotient of discharge divided by cross-sectional area
w	Channel width
y	Exponent in relation of roughness to discharge
z	Exponent in relation of stream slope to discharge
ϱ	Specific weight of water
σ	Standard deviation
τ	Tractive force
ψ	Intercept sphericity

Introduction

Recent quantitative studies of drainage basins and stream channels have yielded significant results which promise greatly to enlarge our knowledge of fluvial processes and erosional landforms. However, the tentative conclusions derived from these preliminary investigations have raised in turn a host of fascinating new problems which are now receiving attention by increasing numbers of research workers. Several current projects involve collection and analysis of new data, with reexamination and rigorous testing of certain basic concepts of geomorphology the immediate objective.

In recent years several different approaches to the study of drainage basins and stream channels have been used, and a considerable variety of areas has been investigated. Strahler (1952, 1954 etc.) and his students, following the lead of Horton (1945), have collected an impressive amount of information on morphology of drainage basins and characteristics of drainage nets. Areas studied thus far include portions of southern California, the badlands in South Dakota, and the Appalachians in Virginia and Tennessee. In most of these examples, morphometric data were obtained from topographic maps, though in several cases considerable field work was done to evaluate the effects of different geological conditions. Extensive use of statistical procedures for design of sampling programs and for tests of significance has contributed materially to the importance of work by Strahler's group. In contrast to this approach, certain other studies have dealt largely with hydraulic aspects of streams, especially channel dimensions, velocity, and suspended load. The study of Midwestern and Western streams by Leopold and Maddock (1953), the report by Wolman (1955) on Brandywine Creek in Pennsylvania, and the discussion of arroyos in New Mexico by Leopold and Miller (1956) all fall into this category. The data used in these investigations were derived from earlier engineering studies, from the huge backlog of available gaging- and sediment-station records, and from field measurements by the various authors. Hack (1957) investigated the effects of geology on channel dimensions, size and lithology of bed material, and longitudinal profiles of streams in the Appalachians of Virginia and Maryland, but he did not consider hydraulic properties of the streams. Brush (in press), working in the Appalachians of central Pennsylvania, made investigations which combined several aspects of previous studies; that is, he used field measurements and gaging-station data to discuss the effects of both hydraulic and geologic factors on properties of drainage basins, channels, and bed material.

From the results of these recent investigations, there can be little doubt that lithology and geologic structure affect the properties of drainage basins and stream channels in ways which are complex but nonetheless detectable by quantitative methods. However, the geologic and hydraulic data collected thus far are not sufficient in either number or areal scope to enable one to state very precisely what cause-andeffect relationships exist, or the extent to which they may differ in various geologic and geographic environments. Efforts at present are concentrated on accumulation of data and on sorting out the several independent and dependent variables. This is a slow process because of the relatively small number of persons involved, the complex interrelationships of the variables, and the large number of different kinds of areas to be studied.

This report adds new data from a high mountain area characterized by geologic and geographic conditions markedly different from those in areas previously studied. Besides the effects of different bedrock lithologies, extreme relief, and vertical changes in climate, the additional factor of glaciation might be expected to affect channel properties and the characteristics of bed material in high mountain streams. Furthermore, it has long been supposed that mountain streams are not graded; that is, they are not in equilibrium, but instead are actively downcutting. This implies that they differ in some possibly measurable way from equilibrium or graded streams. Hence, a principal objective in this study of mountain streams was to test the range of validity of certain conclusions reached in earlier investigations. Briefly, the procedure involved collection and analysis of stream data obtained in the high mountains, and comparison of the results with information of similar character obtained in other, mostly nonmountainous, areas.

Any attempt to segregate geologic controls of characteristics and lithologic composition of bed material requires detailed knowledge of the bedrock geology of the drainage basin. Because the author had recently spent several field seasons mapping the geology of a large area in the southern portion of the Sangre de Cristo Range, New Mexico, the choice of a site for stream studies was resolved by the purely practical consideration of expediency. However, one additional advantage of this area over other possible choices is the fact that the arroyos described by Leopold and Miller (1956) are located in the Rio Grande Depression, at the foot of the range, and afford a basis for local comparison.

The order of discussion will be as follows. A brief résumé of local geography and geology, emphasizing aspects of the physical setting which affect stream properties, will define the general framework of these studies. Next, field procedures will be described, and data obtained from measurements of channel dimensions, bed material, and other properties will be presented. The subsequent sections will describe the results of efforts to segregate specific effects of geology on stream characteristics. Finally, the problem of grade or equilibrium in mountain streams will be considered.

BACKGROUND OF THE PRESENT STUDY

This report is an outgrowth of a more general geologic investigation in the southern Sangre de Cristo Range carried on over a period of several years. Geologic mapping of a 3o-minute quadrangle (105°30' to 106° W. and 35°45' to 36° 15' N.) was begun by the author in 1949, but other duties during the period 1950-1954 interrupted this work. In 1952, Dr. Arthur Montgomery, Lafayette College, began a detailed petrologic study of the Precambrian igneous and metamorphic rocks which comprise the core of the range. The author resumed field work in 1955 and continued during 1956-57. Dr. P. K. Sutherland, University of Oklahoma, joined the group in 1956 and 1957 to study stratigraphic and paleontologic problems of the area. A joint report on the geology of the quadrangle is in preparation.

It was necessary that mapping be essentially complete before stream studies were begun, because collection of data on downstream changes in size and lithologic composition of bed material was one of the major objectives of this investigation. Most of the stream data were obtained in 1956. As it turned out, this was an optimum year for such work because exceptionally light snows, combined with the prolonged drought in northern New Mexico, reduced the flow to such an extent that all streams could be waded safely. Efforts to make additional measurements in 1957 were mostly unsuccessful, owing to extremely high flow.

Acknowledgments

Besides Arthur Montgomery and P. K. Sutherland, my collaborators on the geologic map, many other organizations and individuals have contributed in various ways to the project. Financial support was provided from the Sayles and Shaler funds, Harvard University, in 1949 and 1955, and by the National Science Foundation in 1956-57. Mr. Wallace Miller, of the U.S. Geological Survey office in Santa Fe, made available records of gaging stations located in or near map area.

The New Mexico Bureau of Mines and Mineral Resources financed the cost of final publication.

The following persons served as field assistants: Laura Miller in 1949, Emery Cleaves and Alfred Hoyt in 1955, Robert Rodden in 1956, and Dean Gerber in 1957. Field work was greatly facilitated through the interest and tactical assistance of Mrs. Katherine Kavanagh, Los Piños Ranch, Cowles, New Mexico.

This paper was written at the California Institute of Technology while the author was on sabbatical leave from Harvard University. During this period, Prof. Robert P. Sharp graciously made available office space and other facilities of the Institute's Division of the Geological Sciences.

Luna B. Leopold and Lucien Brush, of the U.S. Geological Survey, read a preliminary draft of the manuscript and made several valuable suggestions which are incorporated in the present text.

Physical Setting

GEOGRAPHICAL DESCRIPTION

The area considered here includes the southern end of the Sangre de Cristo Range and a portion of the Rio Grande Depression, which borders the mountains on the west. The position of the mapped 30-minute quadrangle in relation to physiographic features of the region is shown in Figure 1. An enlarged map of this quadrangle (fig. 2) shows details of the local geography. -The area includes portions of San Miguel, Santa Fe, Mora, Taos, and Rio Arriba Counties; approximately the eastern half of it lies within the boundaries of the Santa Fe and Carson National Forests.

Topographic maps on a scale of 1:24,000, with contour interval of 20 or 40 feet, have been prepared by the U. S. Geological Survey for the western half of the 30-minute quadrangle. The following 7½-minute sheets cover this section: Tesuque, Cundiyo, Chimayo, Velarde, Trampas, Truchas, Sierra Mosca, and Aspen Basin. For the eastern half, only planimetric maps prepared from aerial photographs at a scale of 2 inches to 1 mile are available through either the Soil Conservation Service or the New Mexico State Highway Department. Complete aerial photo coverage is available from either the Soil Conservation Service or the U. S. Geological Survey.

The map area is thinly settled, with most of the population concentrated in several small villages and the remainder on irrigated farms along stream courses. In addition to the major highways shown in Figure 2, there is a network of secondary roads and wagon trails, which make the lower portion of the area fairly accessible. At a few places, roads penetrate short distances into the high country, but for the most part travel in the mountains is by horse or on foot along trails maintained by the U. S. Forest Service.

TOPOGRAPHY AND DRAINAGE

Viewed at a distance and from the west, the Sangre de Cristo Range appears to rise abruptly above an extensive plain which slopes gently toward the Rio Grande. The mountain front, which is quite irregular in plan view, is sharply defined by low foothills. The topographic boundary between mountains and piedmont plain corresponds to the boundary between slightly consolidated sediments of the Rio Grande Depression and hard rocks of the mountains. Closer inspection of the plain shows that it is not a single smooth surface, as it appears from afar; rather, there are remnants of several sloping surfaces which are of both erosional and depositional origin. For the most part, these various surfaces are no longer smooth but have been dissected to such an extent that much of the piedmont lowland has a typical "badlands" type of topography.

In the portion of the Rio Grande Depression considered here, altitudes range from about 6,000 feet to approximately 7,500 feet at the mountain front. Local relief does not exceed a few hundred feet and is considerably less at most places. The mountains rise gradually from altitudes of 7,500 to 8,000 feet, at their western margin, to a maximum of about 13,300 feet, at the top of South Truchas Peak. The crest of the range from Lake Peak to Jicarilla Peak is marked by a group of rugged summits connected by serrate ridges. Northward for a considerable distance from Jicarita Peak, the crest is poorly defined and consists of a high, dissected, plateaulike ridge. The north-central portion of the map area includes part of the Picuris Range, which is a narrow prong extending roughly r5 miles westward from the main mass of the Sangre de Cristo Range. In the southeastern portion of the area (fig. 2), the divide east of the Pecos River is nearly as high as that to the west but consists of a single essentially smooth ridge instead of individual sharp-crested peaks. Differences in the extent of Pleistocene glaciation apparently are responsible for these topographic contrasts. Local relief is greatest in the most severely glaciated areas but amounts to 2,000 feet or more in a few of the nonglaciated canyons.

The western slope of the Sangre de Cristo Range in this area drains to the Rio Grande by way of Embudo Creek, the Truchas River, the Rio Santa Cruz, and the Pojoaque River. In the mountains, valleys of these streams are deep, narrow canyons, but beyond the mountain front, in the unconsolidated deposits of the Rio Grande Depression, valley depths decrease abruptly and widths increase markedly. Runoff from snowmelt and summer rains is sufficient to maintain appreciable perennial flows in the mountains, but many streams become ephemeral within a few miles from the mountain front, owing to percolation into the unconsolidated deposits. West of the mountain front, large flows in these streams occur only during the height of the snowmelt season or for a few hours after heavy rains. The characteristics of ephemeral streams, or arroyos, in this portion of the Rio Grande Depression have been described in considerable detail by Leopold and Miller (1956).

The southeastern part of the map area is occupied by the headwaters portion of the Pecos River drainage basin. All the tributaries shown in Figure 2 flow in deep canyons; as a matter of fact, the canyon of the main stem extends many miles farther downstream beyond the southern boundary of the map. In the area considered here, the Pecos maintains an appreciable perennial flow.

The eastern slope of the range in the area considered here drains to the Pecos and Canadian Rivers.

CLIMATE

The available data do not permit a detailed description of climate in this area. The only weather station in the vicinity where daily observations of temperature and precipitation are obtained is at Santa Fe (elevation 7,000 feet). Although there are nearly a dozen additional part-time precipitation stations within the quadrangle boundaries, only one of these is located above 9,000 feet.

The most striking feature of climate in this area is its variation with altitude. Average annual precipitation ranges from 10 to 20 inches, in the vicinity of the Rio Grande, and increases upward to an estimated 30 to 35 inches, on the highest peaks. Approximately one-half of the total precipitation falls during the summer months. Ordinarily, the rainy season begins about July r and continues through August. Precipita-

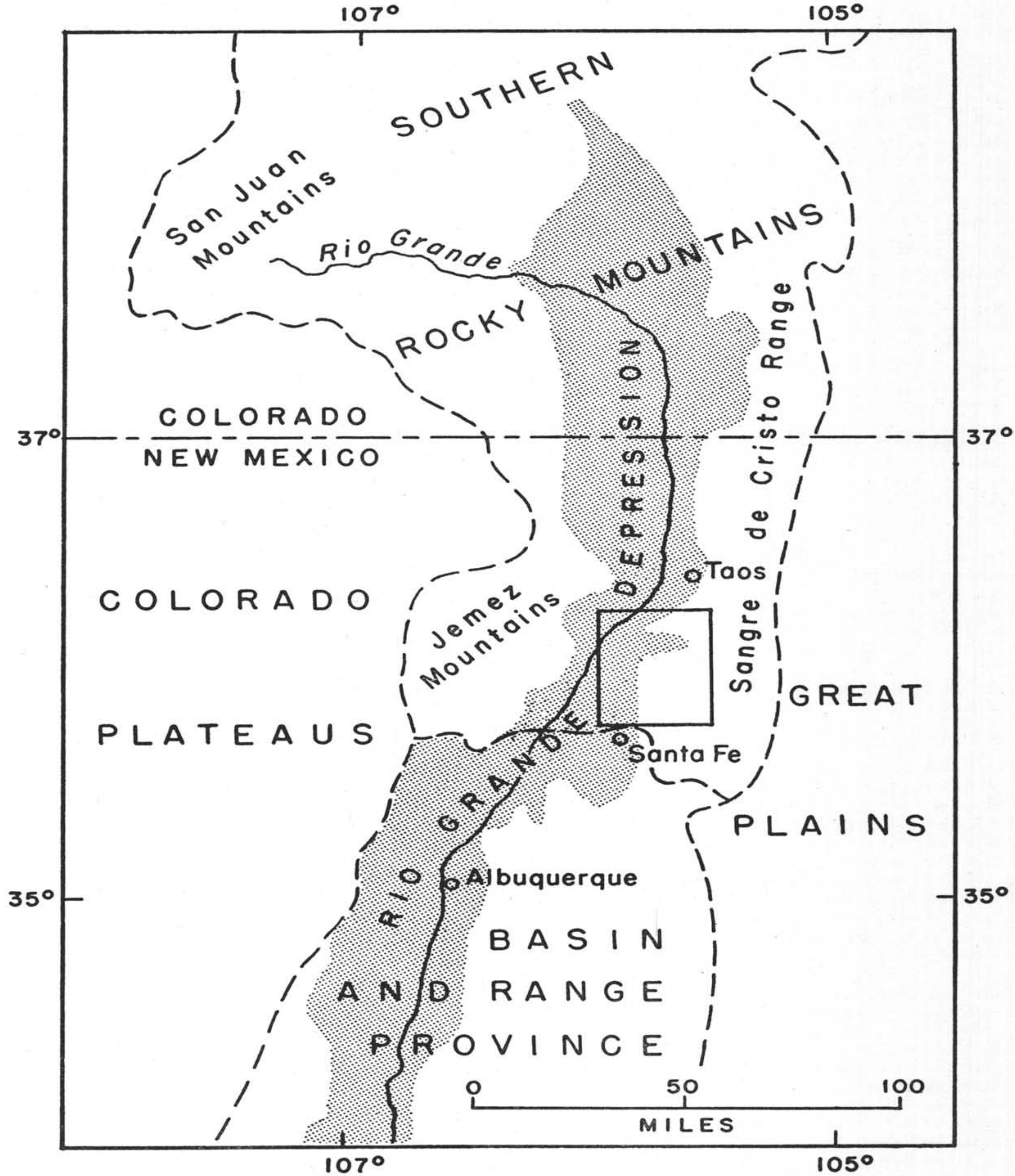


Figure 1

Location map, showing relation of area considered in this report to regional physiographic features

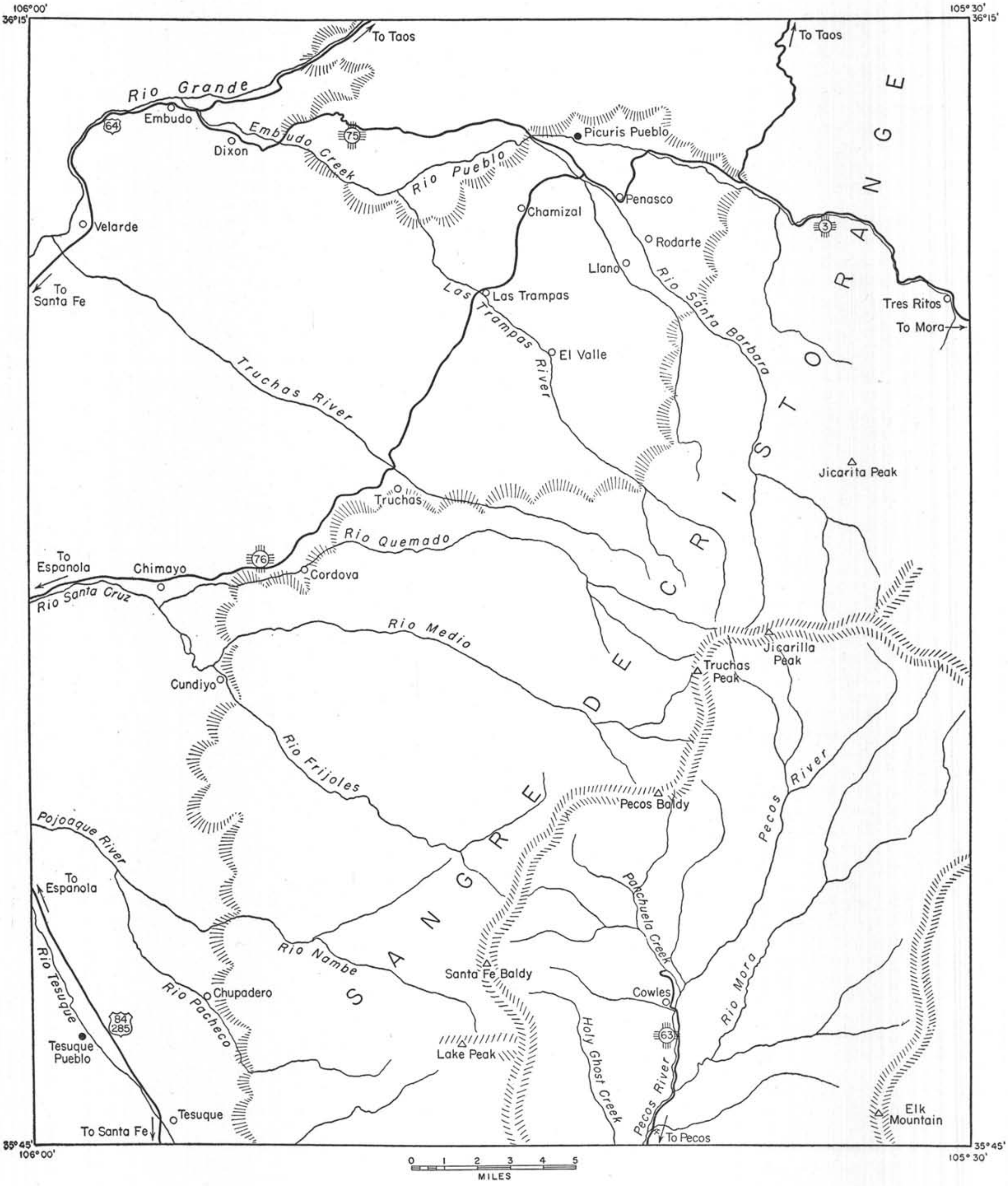


Figure 2

GEOGRAPHIC FEATURES OF THE MAP AREA

tion occurs usually as brief afternoon thunderstorms, but there are occasional extended rainy periods such as occurred during the summers of 1955 and 1957. Winter snowfall ranges from about 15 inches, near the Rio Grande Valley, to more than 60 inches, in the mountains. In summer, daytime temperatures may reach 70°F in the mountains and 90° to 95°F in the low country. The winter mean temperature is slightly below freezing in the lowlands and considerably colder in the high mountains. Humidity is low both winter and summer, and more than 75 percent of the days in the year are sunny. Gentle winds blow most of the time in the low country, and in the spring may reach sufficient velocities to cause severe dust storms. Strong winds prevail in the high country both winter and summer.

The direct effects of climate on stream flow in this area are similar to those in other high mountain ranges. Melting of snow produces a flood in the late spring or early summer of each year. Furthermore, replenishment of ground-water storage during the snowmelt season sustains flow through the summer months. Thunderstorms which occur during the rainy season supplement the base flow and occasionally may yield sufficient precipitation to cause flow in the arroyos of the lowlands. As discussed in the next section, climate also affects streams indirectly through its effect on vegetation.

VEGETATION

Zonation of climate according to altitude is shown clearly in this area by the distribution of natural vegetation. The porous materials which underlie the semiarid Rio Grande Depression support only a sparse cover of juniper, sage, grasses, and cacti. In vertical order of their occurrence, zones of pinion, yellow pine, aspen, spruce, and fir dominate the luxuriant forest which covers the mountains up to the timberline at 11,000 to 11,500 feet. Because most of the forest zone lies within the boundaries of a national wilderness area, logging has been negligible. However, there are several large burns, most of which are several decades old. Above timberline, there are a few clumps of dwarf trees and, in addition, several kinds of arctic-type sedges and grasses.

As in any other forested area, geological field work was affected adversely by the dense cover of timber. Not only are the rocks covered at most places by soil and forest litter, but also freedom of movement through the forest is restricted greatly by incredible tangles of "deadfalls."

Properties of streams are influenced in several ways by the natural vegetation. The sustained flow of clear water in mountain streams reflects in part the capacity of vegetation to inhibit runoff and immobilize soil. Even after heavy rains, these streams carry small amounts of suspended-sediment load, generally only enough to make the water murky. By contrast, the arroyos of the Rio Grande Depression flow only after heavy rains and carry tremendous suspended-sediment loads derived mostly from lowland areas where the cover of vegetation is sparse. Also, as will be discussed later, vegetation may be partly responsible for very striking differences in channel width between perennial mountain streams and ephemeral arroyos of the lowlands.

Because of the climate and rugged topography, agriculture is confined to a few protected mountain valleys and to lowland areas where there is sufficient water for irrigation. Length of the growing season ranges from about l00 days, in the mountains, to 180 days, near the Rio Grande. Irrigation has been practiced in this area for several centuries, but at most places agriculture is still a precarious venture, owing to insufficient water during frequent droughts. Individual farms are small, with only a few acres devoted to crops, principally grain, hay, fruit, and vegetables, especially chile and beans. Cattle are grazed in the high meadows of the national forest lands and sheep in a few areas above timberline.

Studies of natural streams in this area are affected by the distribution of agricultural lands for the reason that raising crops requires removal of irrigation water from streams. Because artificial withdrawal of water introduces yet another variable into the vastly complex subject of stream behavior, the investigations reported in this paper were restricted to areas upstream from major irrigation diversions.

SUMMARY OF GEOLOGY

Figure 3 is a generalized geologic map and structure section showing the major features of lithology and structure. Outcrop patterns and geologic structure are shown in greater detail on the maps (pl. 1-4) of individual drainage basins.

The oldest rocks of the Sangre de Cristo Range in this area are Precambrian quartzites, schists, and granites, which underlie most of the mountain crest and also are exposed in several deep canyons of the Pecos River drainage system. A thick section of Paleozoic sedimentary rocks overlies the Precambrian basement complex. Up to 100 feet of poorly fossiliferous limestone of possible Mississippian age may be present at the base of the Paleozoic sequence. Except for this thin zone at the base, the remainder of the sedimentary column in the mountains is of Pennsylvanian age. Several thousand feet of thick nonmarine sandstones, alternating with thin marine limestones and shales, are exposed over wide areas, especially east of the mountain crest. Except for a few very small patches along the Rio Nambe, Paleozoic rocks are not exposed in the map area on the western slopes of the range south of the Rio Santa Barbara. The youngest rocks of the area are those in the Rio Grande Depression, where several thousand feet of calcareous, slightly consolidated sand and gravel deposits, together with interbedded basalt flows, comprise the Santa Fe formation of Miocene-Pliocene age.

Structurally, this portion of the Sangre de Cristo Range is a broad anticline broken at several places by major faults that trend north-south. These structures are considered to be of Laramide age. Deformation evidently continued intermittently throughout most of the Tertiary, because the Santa Fe formation is tilted and locally faulted. In addition, the Precambrian metamorphic rocks are complexly folded, with axes trending roughly east-west, transverse to the structures of Laramide age mentioned above. The geologic structure of Precambrian rocks in the Picuris Range has been described in detail by Montgomery (1953).

Pleistocene glaciation of the Sangre de Cristo Range is evident both from the numerous cirques and from the rugged character of the higher peaks. However, glacial and glaciofluvial deposits are remarkably sparse in most valleys. The approximate limits of glaciation are shown in the drainage basin maps (pl. 1-4). Patterned ground, block fields, and other features produced by Pleistocene frost action are abundant at altitudes above 10,000 feet.

Alluvial fan deposits, locally several hundred feet thick

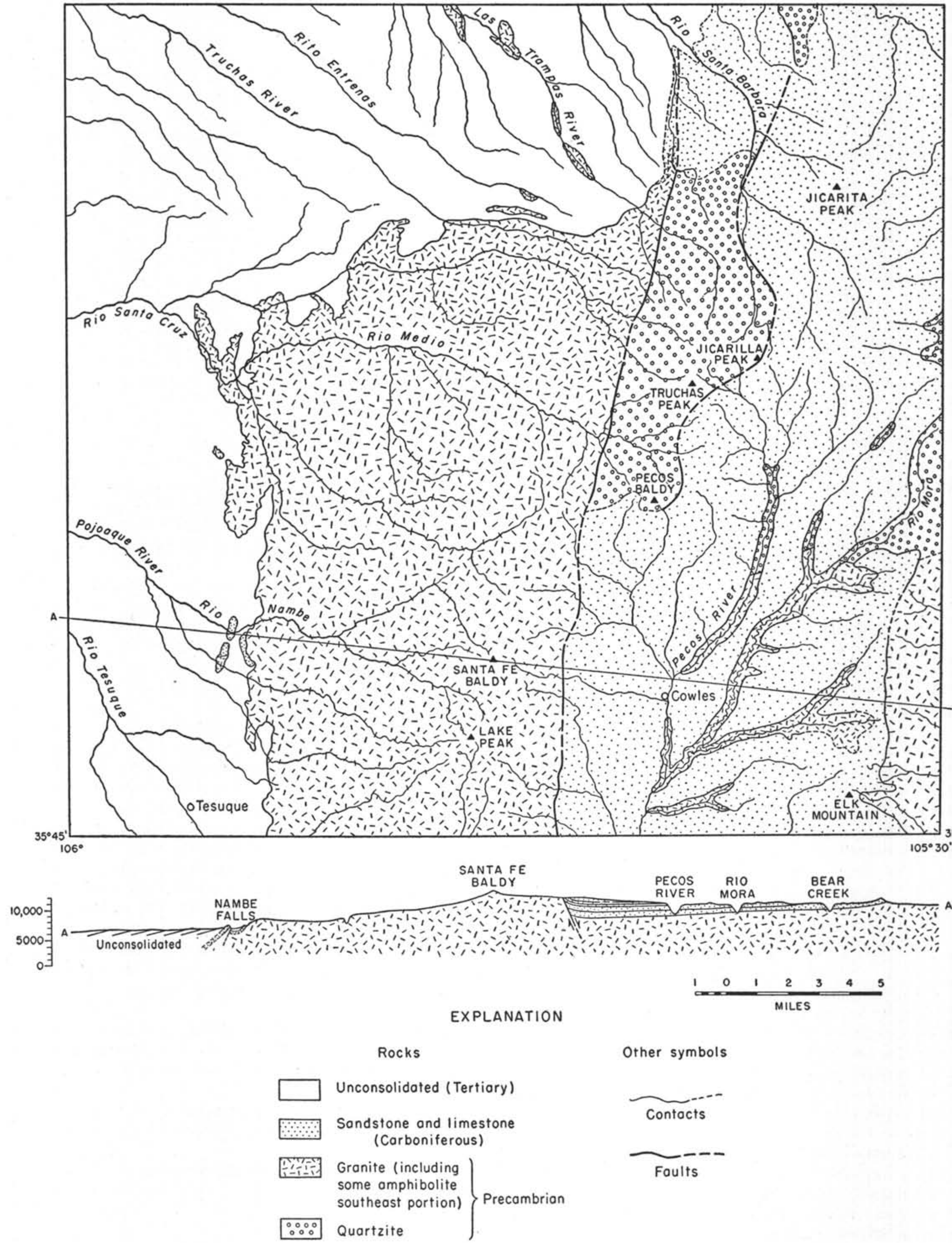


Figure 3

GENERALIZED GEOLOGIC MAP AND STRUCTURE SECTION OF THE AREA CONSIDERED IN THIS REPORT

(Based on preliminary mapping by Miller, Montgomery, and Sutherland.)

and probably of Pleistocene age, overlie the typical Santa Fe formation throughout much of the Rio Grande Depression considered here. In addition there are other gravel deposits which may be related to pediment surfaces, as suggested by Cabot (1938).

Arroyos in the Rio Grande Depression are flanked by terraced, fine-grained alluvial deposits which occupy valleys one-fourth to one-half mile wide. Radiocarbon and archeologic ages indicate that the bulk of this alluvium was deposited during the last two millennia (Miller and Wendorf, 1958). The recent alluvial deposits are absent in the mountain valleys.

By way of summary, it should be emphasized that the principal features of stratigraphy, structure, and geologic history in the area studied are typical of the entire Southern Rocky Mountains region. This fact is of importance in connection with the present investigation only to the extent that conclusions reached about the effects of geology on characteristics of mountain streams may be applicable outside the specific area studied.

Basic Data

The principal criteria used in choosing drainage basins for detailed study were accessibility and freedom from artificial influences, such as irrigation.

In the selection of individual measuring stations, an effort was made to pick reaches of channel which were essentially straight. Also, for reasons which will be discussed later, many of the stations were located as close as possible to stream junctions. Locations of other stations were determined by various other conditions, the most important being changes in bedrock lithology along the stream.

Measurements of channel dimensions, channel slope, and lithology and size of bed material were obtained at a total of 104 stations in 4 different drainage basins. The locations of sampling stations in each basin are shown in Plates 1 to 4. There are 53 stations in the Pecos River basin, 13 in the Rio Santa Cruz basin, 18 in the Rio Santa Barbara basin, and 2o in the Pojoaque River basin. In addition to these 104 stations, partial measurements were obtained at about two dozen other stations.

Elevations of sampling stations range from 6,400 to I 1,000 feet, distributed as follows:

6,000-7,000	feet — 11	stations
7,000-8,000	" — 12	"
8,000-9,000	" — 23	"
9,000-10,000	" — 28	"
10,000-11000	" — 30	"

Nearly 80 percent of the stations lie above 8,000 feet, and more than half of them above 9,000 feet.

At each station, the following kinds of data were obtained:

1) Bankfull width was measured by stretching a tag line from the edge of one channel bank to the other. Streams less than r o feet wide were measured to the nearest half foot, and all others to the nearest foot.

2) Bankfull depth was determined by measuring, to the nearest one-tenth of a foot, the distance from the taut tag line to the channel bottom. Depending on the width of the stream, 8 to 25 such measurements were made at intervals across the channel section. Mean bankfull depth is defined as the average of these individual measured depths.

3) Slope of the water surface at each station was determined by using a tape, hand level, and stadia rod. Horizontal distances were measured to the nearest foot, and vertical distances to the nearest one-tenth of a foot. Slope at each station was measured through a distance which corresponded to the length of stream segment where the sample of bed material was obtained (75 to 150 feet).

4) Bed material, which is composed of pebbles, cobbles, and boulders, was sampled systematically by picking up individual particles located at the intersections of a grid pattern. The method used in sampling is essentially like that described by Wolman (1954), Hack (1957), and Brush (in press). Instead of actually laying out a grid, the procedure was to make a series of equally spaced traverses upstream and downstream along each sampling reach and to pick up particles for measurement at one-step intervals. The number of such traverses ranged from 3 to 6, depending on stream width. The particle picked up at each point was the one touched (without looking) by reaching over the toe of the boot with the middle finger protruding. The intermediate diameter (b-axis) of each particle was measured to the nearest millimeter, and its lithology recorded. The number of particles measured at individual stations ranged from 35 to 500, but most commonly the sample size was 50 to 60 particles. Altogether, nearly 7,000 particles were measured.

More detailed consideration of the general problems involved in sampling streambed material, and a brief evaluation of the precision of the method used in this study, will be given in a later section.

5) Stream length, meaning the distance from the drainage divide to a particular station, and also the drainage area above each station, were determined by measuring these quantities on maps.

The various measurements were all made by the author, so as to eliminate possible complications due to operator variation.

The data obtained from the measurements, and the results of various computations involving these data, are summarized in Tables 15-18.

Characteristics of Drainage Basins

A rigorous description of drainage basins, including such characteristics as basin circularity, area-altitude relations, and other measurable properties, is not possible for this area as a whole because of inadequate maps. Hence, only the general properties of drainage basins which bear on the subsequent discussion of stream channels will be discussed here.

Geographic relations of the Rio Santa Barbara, the Rio Santa Cruz, the Pojoaque River, and the Pecos River are given in Figure 1. More detailed representations of drainage patterns and relations of channels to lithology and structure of the bedrock are shown in Plates 1-4. The Pecos River and the Rio Santa Barbara drain areas underlain dominantly by Pennsylvanian sandstone. The Rio Santa Cruz and the Pojoaque River flow westward from the mountain crest over granitic terrain, and beyond the mountains they both flow on the unconsolidated deposits of the Santa Fe formation.

As can be seen in Plates 1-4, the various drainage basins differ somewhat in shape, but all have dendritic drainage patterns. Although no diagrammatic or tabular summary is presented here, it should be mentioned that the several characteristics of drainage networks described by Horton (1945) apply to the mountain portions of these basins. Number, average length, average slope, and average drainage area of streams are related to stream order by geometric series. Leopold and Miller (1956) presented these same kinds of data for ephemeral streams in the Rio Grande Depression.

One convenient way of indicating the general form of drainage basins is by plotting stream length against drainage area, as shown in Figure 4. The points on these graphs are for sampling stations in the mountains only. Despite the fact that any line drawn through the scattered points obscures the abrupt changes which occur at stream junctions, it is obvious from the coefficients and exponents in the power functions relating length and drainage area that the four basins studied are quite similar. Furthermore, drainage basins of arroyos (Leopold and Miller, 1956) exhibit almost exactly the same relationship of stream length to drainage area as the mountain streams. From these results and the conclusions of Hack (1957) and Brush (in press), it appears that this method of description indicates general similarity of drainage-basin form in a wide range of geologic and geographic environments. This tentative conclusion will be the subject of further discussion in a later section (p. 40).

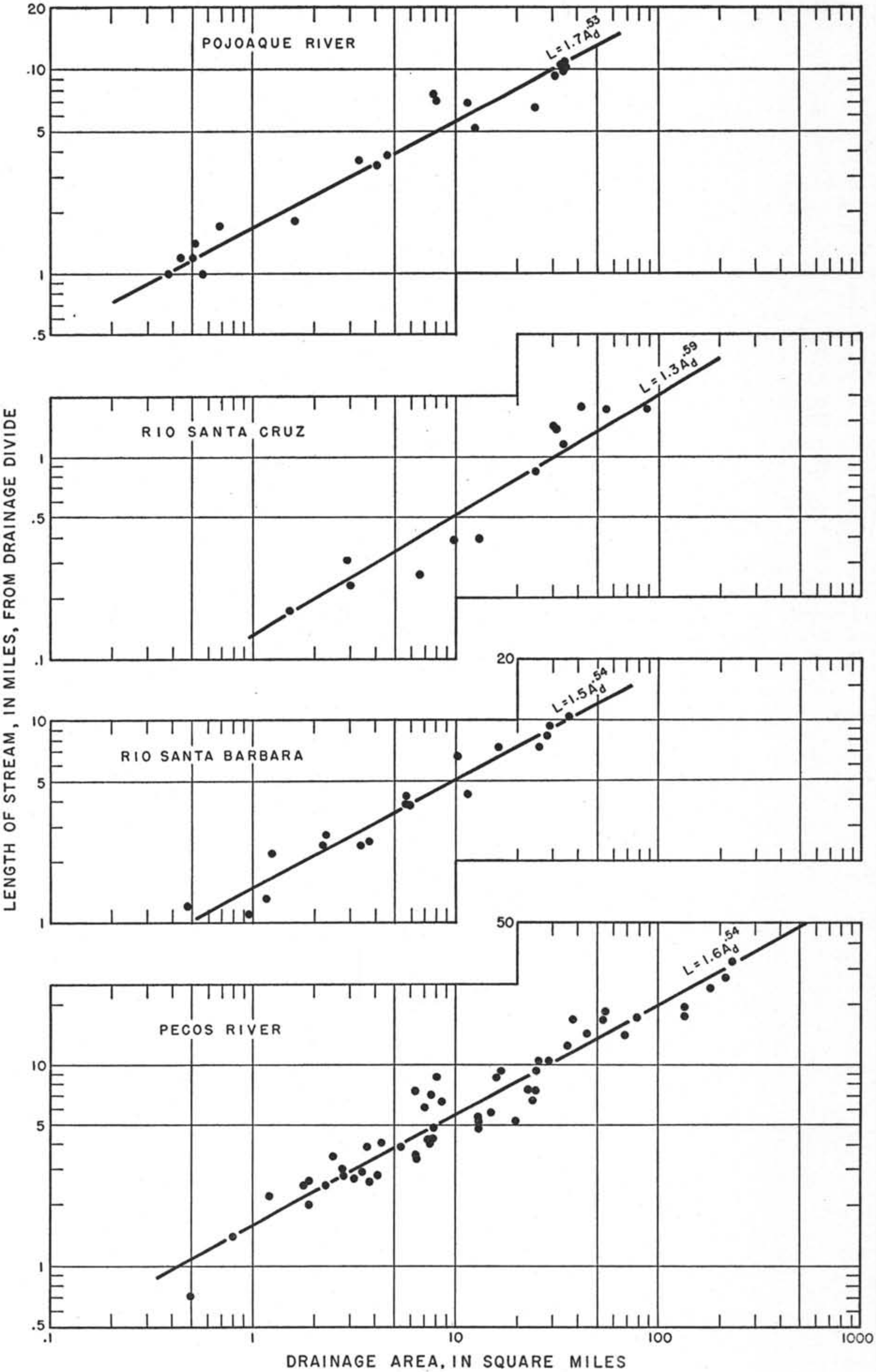
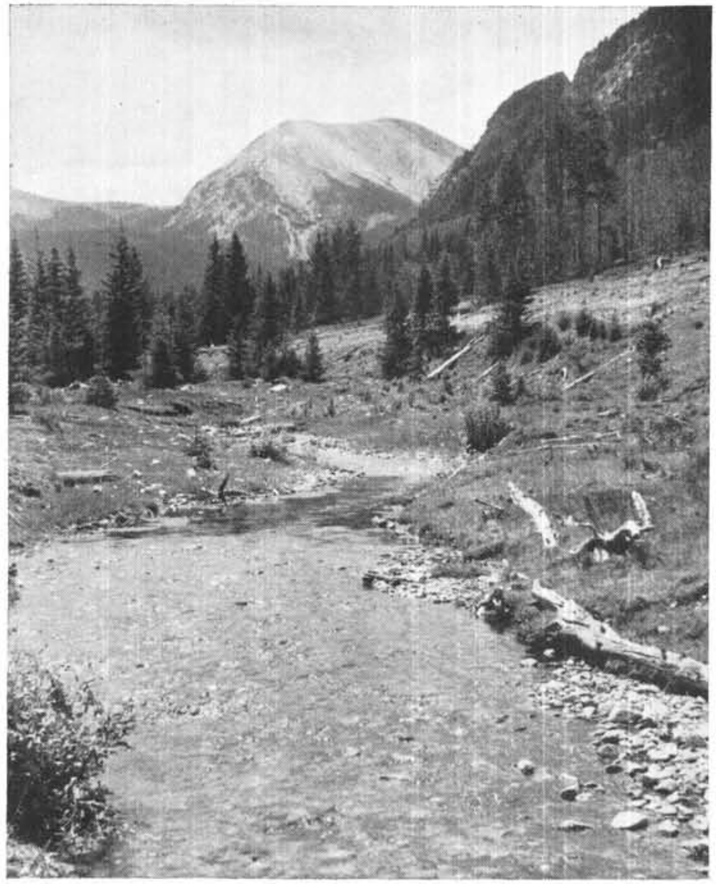


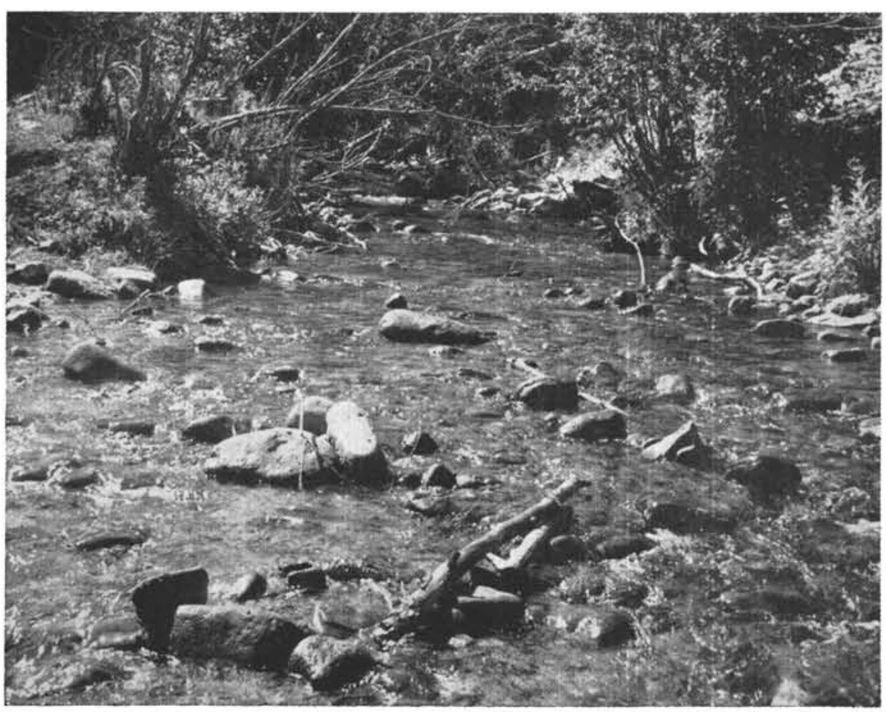
Figure 4

RELATION OF STREAM LENGTH TO DRAINAGE AREA IN EACH OF THE FOUR BASINS STUDIED

The data are for stations in the mountains.



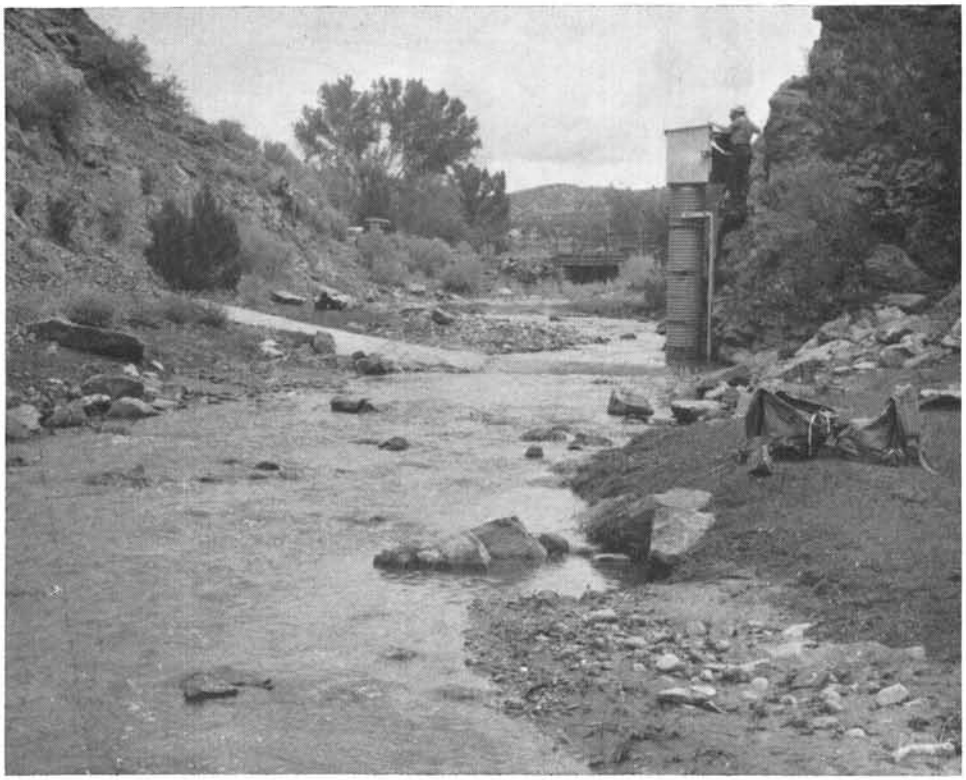
A. West fork at station 57 (alt. 9,670 ft). Jicarilla Peak in background and cliffs at right are composed of quartzite. Bed material is mostly quartzite of considerably finer size range than that occurring both upstream and downstream. Gradient of channel at this point is very slight.



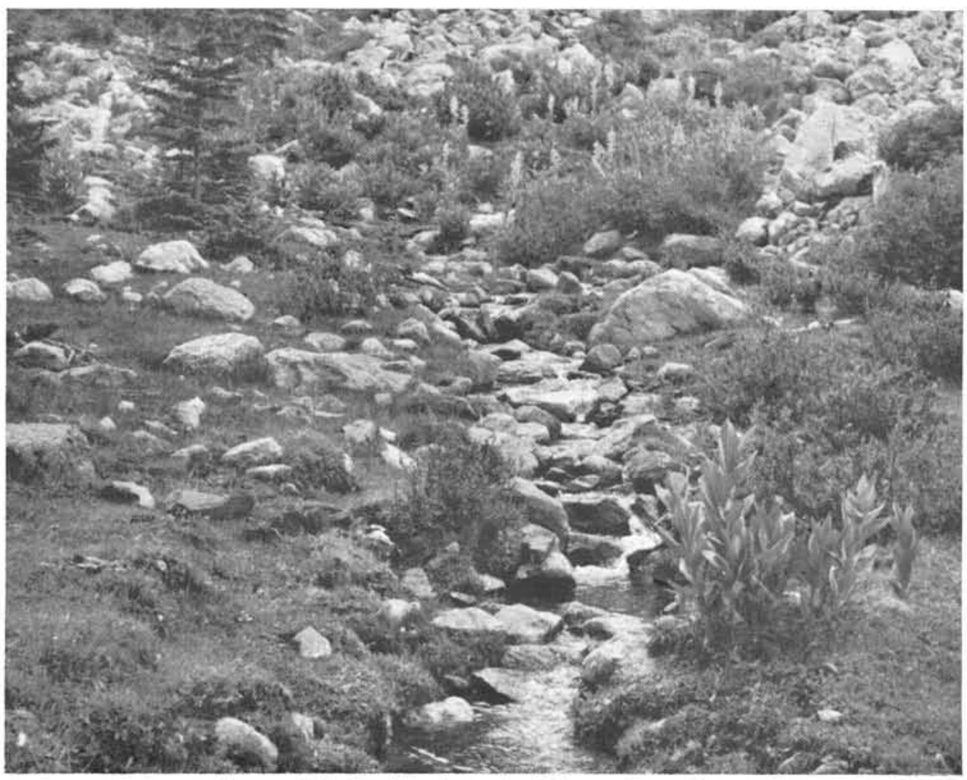
B. West fork at station 56C (alt. 9,160 ft), just above junction of East Fork, and about 2 miles downstream from station 57 shown above. Bed material is dominantly sandstone and is much coarser than at station 57.

Plate 5

RIO SANTA BARBARA



A. Rio Santa Cruz at the Cundiyo gaging station (station 53A; alt. 6,470 ft). Note boulders from nearby granite cliffs, in the channel.

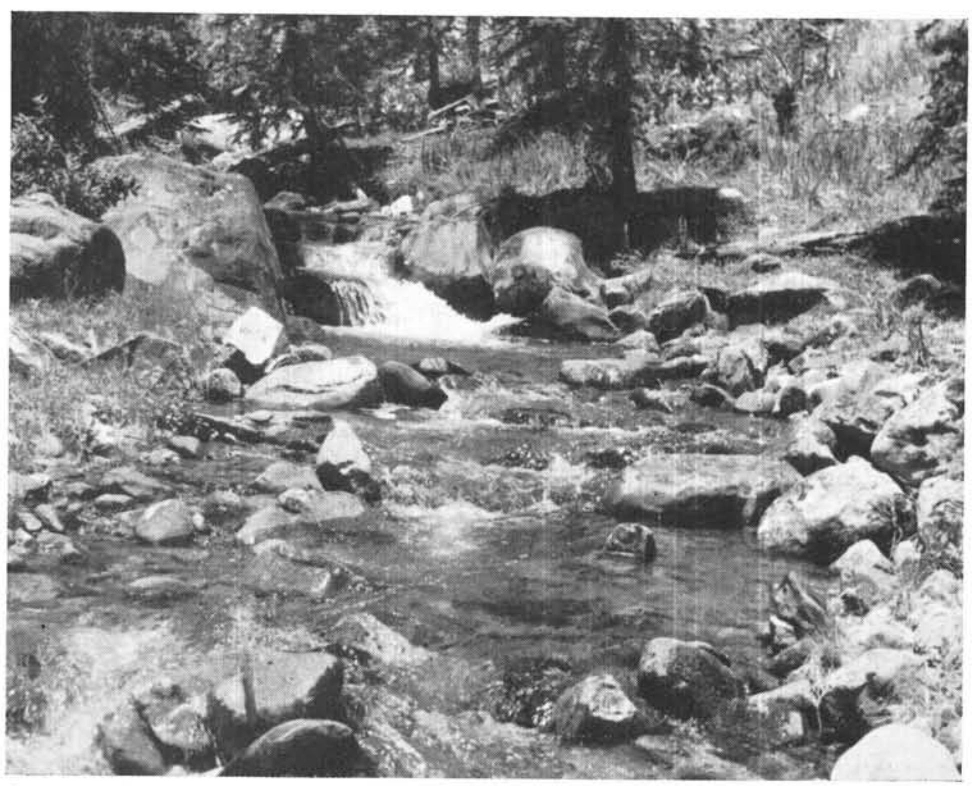


B. Rio en Medio, a tributary of the Rio Nambe in the Pojoaque River drainage basin. This location is about one-quarter mile upstream from the Santa Fe Basin ski lodge (station 19; alt. 10,600 ft). The channel is about 7 feet wide and 1 foot deep. Channel dimensions vary considerably from place to place, owing to the presence of very large granitic boulders which are largely of glacial origin.

Plate 6

RIO SANTA CRUZ AND RIO EN MEDIO

A. Rito de las Chimoyosos about 50 feet above junction of Rito Azul, at station 61C. Bed material is dominantly angular sandstone boulders from outcrop in background.

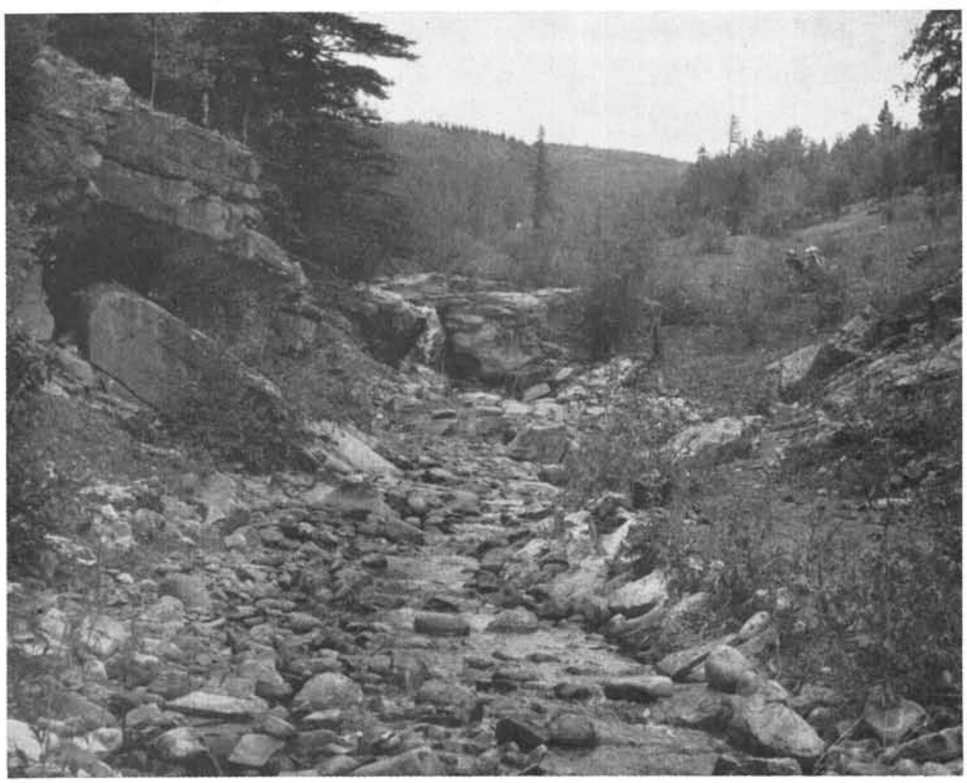


B. Rito de las Chimoyosos at station 10B (alt. 9,620 ft), about one-half mile downstream from station 61C shown above. Largest boulders are quartzite fragments from nearby outcrops.

Plate 7

Upper Pecos River Drainage Basin

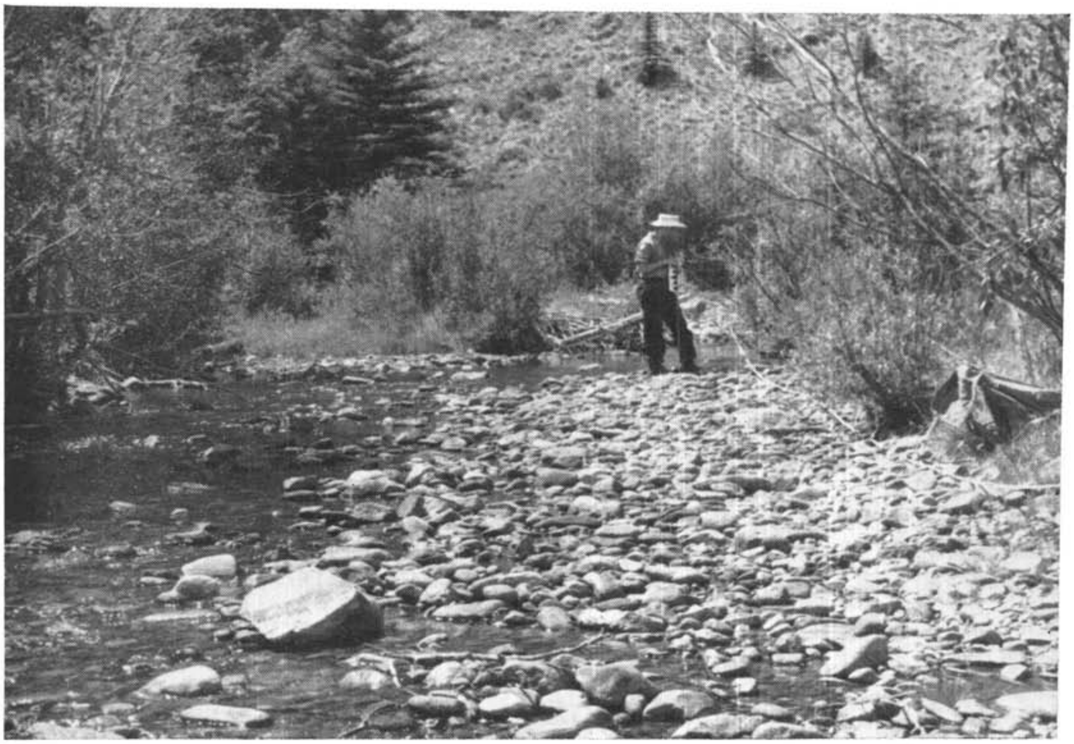
A. Rio Valdez at station 13, about 1 mile upstream from junction with the Rio Mora. Bed material is mostly sandstone.



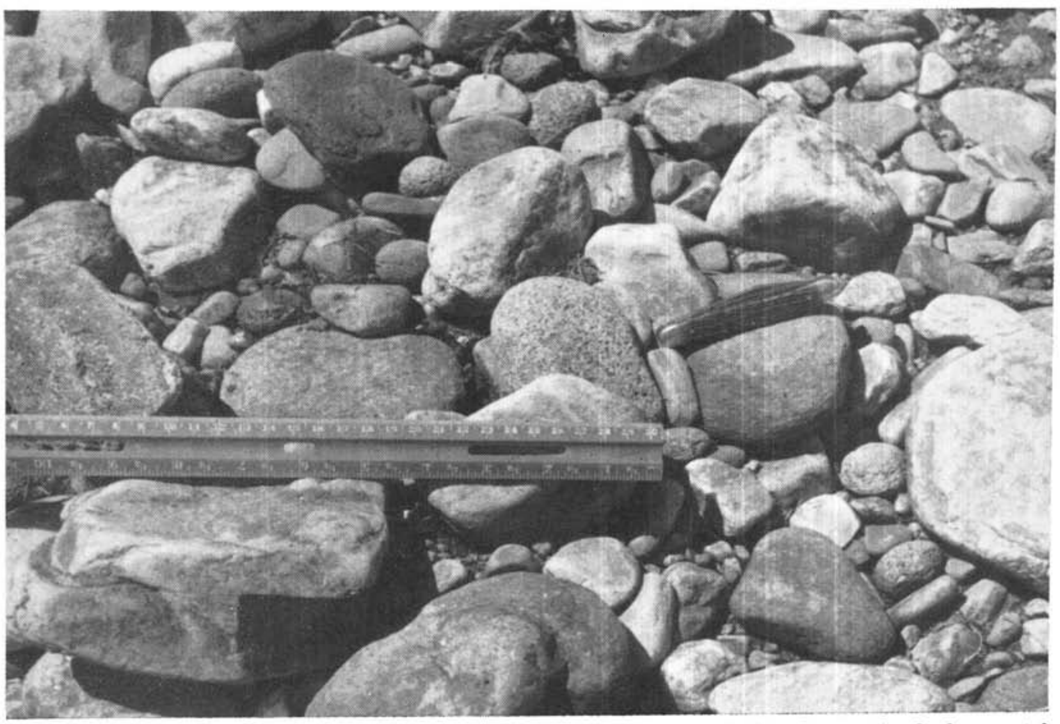
B. Jack's Creek at station 41 (alt. 8,290 ft), approximately 300 yards above mouth. Limestone ledge contributes about 12 percent of the bed material in reach immediately downstream.

Plate 8

UPPER PECOS RIVER DRAINAGE BASIN



A. Channel of the Rio Mora above the junction of the Rito los Steros (station 14B).



B. Closeup of stream bed at station 14B shown above. In order of abundance, the bed material consists of quartzite, sandstone, and granite.

Plate 9

Rio Mora

A. Station 8, located about 7 miles north of Pecos, New Mexico. Channel at this point is 60 feet wide. Large boulders are granite and amphibolite.

B. View upstream from Highway 84 bridge, 2 miles south of Dalia, New Mexico, about 60 miles downstream from station 8 shown above. Banks are composed of silty alluvium, and bed material is mostly sand, with some gravel on bar at center of photo.

Plate 10

MAIN STEM OF THE PECOS RIVER

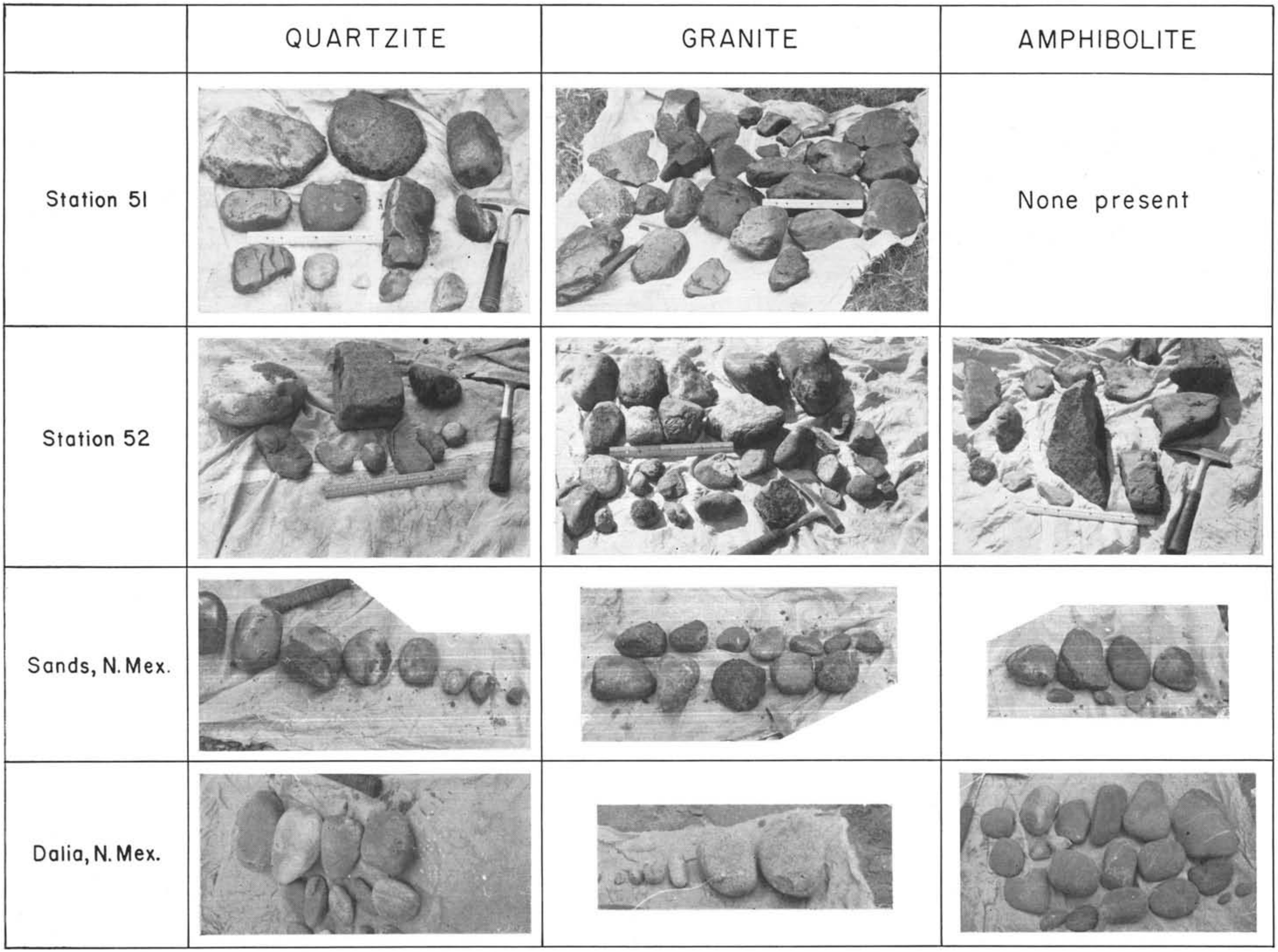


Plate 11

Downstream Changes in Size and Shape of Quartzite, Granite, and Amphibolite Fragments in the Bed Material of the Pecos River

Distances of the various stations from drainage divide are as follows: Station 51, 12 miles; station 52, 27 miles; Sands, New Mexico, 48 miles; Dalia, New Mexico, 84 miles.

A. Station 43, where width is 8 feet and depth is 0.9 foot.

B. Two miles downstream from station 43, where width is 24 feet and depth is 1.0 foot.

Plate 12

CHANGES IN WIDTH OF THE RIO CHUPADERO AT THE MOUNTAIN FRONT

Characteristics of Stream Channels

The streams referred to in this report are typical high-mountain torrents, characterized by clear, cold water flowing at appreciable velocities in gravel-covered channels which have steep and rather irregular gradients. There are a few waterfalls, most of which do not exceed io to 15 feet in height. These streams flow in deep, narrow canyons, and there are only a few spots in the entire area which have valley flats of alluvial origin large enough to be called flood plains. Valley side slopes rise steeply only a few feet, or at most a few tens of feet, from the channel. However, except for some of the steeper canyon stretches, relatively little bedrock is exposed on the valley walls, because of the luxuriant cover of grass and trees. Furthermore, there are very few places where gravel is absent and bedrock is exposed in stream channels.

At most places, the channels do not have obvious well-defined pools and riffles. Indeed, almost the entire length of each stream might more appropriately be considered as a continuous riffle. Reasons for this important difference from streams at lower elevations will be discussed in a later section of this paper (p. 46).

Although the channel banks stand nearly vertical at many places, they are not easily seen, because of water in the channel and the presence of vegetation near the channel edge. At all points examined, bank materials consist mostly of gravel, with sand, silt, and clay filling the interstices.

An impression of the general appearance of streams in the Sangre de Cristo Range may be obtained by examining Plates 5-8. These photographs were taken in 1956 when the streams were carrying smaller discharges than at any time during the previous 3o years. In addition to the typical boulder-strewn character of the channels, the presence of grass, brush, and trees along the edges of channel banks should be noted.

Data obtained from measurements of channel dimensions at the various sampling stations in the four drainage basins will be presented in this section. The effects of geological conditions on channel properties will be discussed in a later section (p. 40).

BANKFULL WIDTH AND DEPTH

Despite many irregularities caused by the presence of large boulders, Sangre de Cristo streams of all sizes have trapezoidal to rectangular cross-sections, as shown by the examples in Figure 5.

Because the data for channel width and depth were to be used in estimating bankfull velocity, it was necessary to determine the amount of local variation in channel dimensions. Measuring errors in the method used for determining width and depth are negligible; that is, once the tag line is stretched from bank to bank, replicate sets of measurements give essentially identical results. If the tag line, however, is relocated between repeated measurements at the same section, differences amounting to less than 0.5 foot for bankfull width, and less than o.t foot for mean bankfull depth, may be obtained. The two main reasons for these slight differences are (1) uncertainty in locating the edge of the channel bank, especially if the bank is gently sloping, and (2) taking individual depth measurements at slightly different places in the cross-section.

Several sets of measurements were made for the purpose of determining the variations in width and depth along short (less than one-fourth mile), straight segments of channel where no tributaries enter. The results of these studies are summarized in Table i, and one set of data is shown in Figure 6. It is apparent that both bankfull width and mean bankfull depth vary appreciably over short distances. Local differences in height of the channel banks on opposite sides of the stream, as shown in the profile of Figure 6, would affect measured values of both width and depth, and may account for a considerable portion of the variation.

THE THALWEG

Practically all measured cross-sections show a well-defined zone of maximum depth (the thalweg). Because the measuring stations were located in straight reaches, it might be expected that the channels would be symmetrical, with the deepest point in the middle. Analysis of the data indicates that this is not true in most cases. Dividing the horizontal distance from the bank to the thalweg by the width of the channel defines the relative position of the thalweg, and affords a basis for comparison of thalweg positions in channels with different widths. The channels considered here range in width from 3.5 to 76 feet. Computed values of thalweg positions were tabulated according to categories defined by dividing the channel cross-section into five parts of equal width. The results are summarized in Table 2.

There are appreciable differences between drainage basins, but these are difficult to evaluate because of the different number of measuring stations in each basin. At slightly less than a third of all the stations, the thalweg is in the center one-fifth of the channel, and in about one-sixth of the cases, the thalweg is located in the zone nearest the bank. Obviously, the categories labeled *intermediate* and *outermost* in Table 2. may occur on either side of the center of the channel. The average value for 126 stations places the thalweg to the right or left of the center line by an amount equal to 0.4 of the distance from the center to the bank; that is, exactly in the middle of the intermediate one-fifth of the channel. There is no relationship between channel size and the position of the thalweg in the channel.

It should be noted (table 2) that the thalweg does not occur in the outermost one-fifth of the channel at any of the 13 stations in the Rio Santa Cruz drainage basin. Furthermore, the percentage of cases with the thalweg in the middle one-fifth is more than twice that for stations in other basins. No obvious differences in bed or bank material which might account for this condition were observed.

Leopold and Wolman (1957) have described meandering thaiwegs in straight segments of channel. The data from individual stations discussed above suggest the same

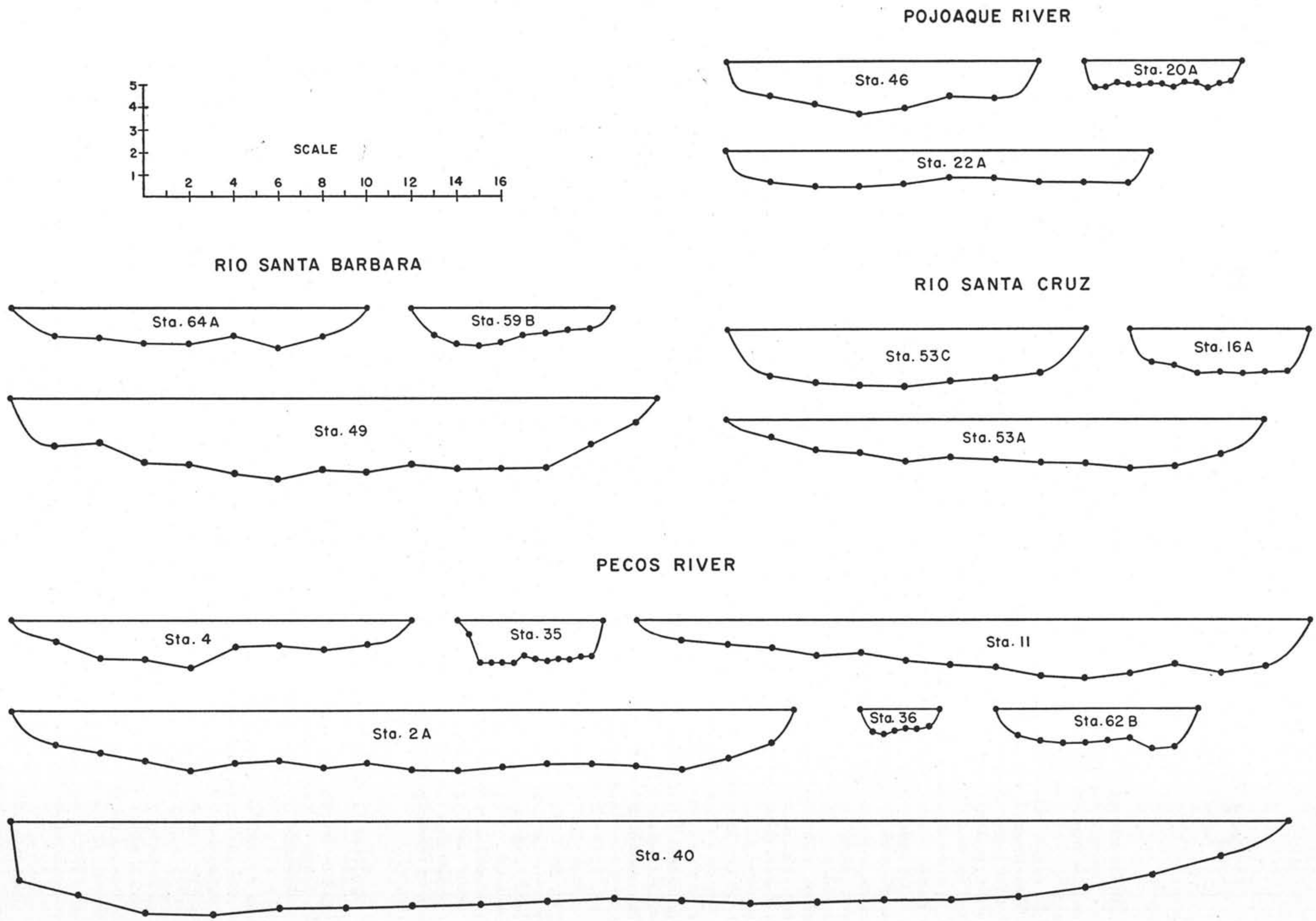


Figure 5

EXAMPLES OF CHANNEL CROSS-SECTIONS AT SEVERAL STATIONS IN EACH OF THE FOUR DRAINAGE BASINS STUDIED

No vertical exaggeration.

TABLE 1. VARIATION IN BANKFULL WIDTH, MEAN DEPTH, AND CROSS-SECTIONAL AREA OF STREAM CHANNELS ALONG SEGMENTS ONE-FOURTH MILE OR LESS IN LENGTH, WHERE NO TRIBUTARIES ENTER

STREAM AND STATION NUMBER	BANKFULL WIDTH	BANKFULL DEPTH	CROSS-SECTIONAL AREA
Pecos River Stations 5A and 6	25, 22, 21, 22; $\bar{w} = 22.5$, $\sigma = 1.73$, $\sigma = 8\%$ of $\bar{w}$	1.81, 1.48, 1.64, 1.84; $\bar{d} = 1.69$, $\sigma = 0.17$, $\sigma = 10\%$ of $\bar{d}$	45, 33, 34, 40; $\bar{A} = 38.0$, $\sigma = 5.6$, $\sigma = 15\%$ of $\bar{A}$
Pecos River Station 38	15, 11, 14, 16; $\bar{w} = 14.0$, $\sigma = 2.16$, $\sigma = 15\%$ of $\bar{w}$	1.20, .85, 1.10, 1.15; $\bar{d} = 1.08$, $\sigma = 0.14$, $\sigma = 13\%$ of $\bar{d}$	18, 9, 15, 18; $\bar{A} = 15.0$, $\sigma = 4.2$, $\sigma = 28\%$ of $\bar{A}$
Pecos River Station 40	57, 54, 54, 57, 59, 53; $\bar{w} = 55.7$, $\sigma = 2.32$, $\sigma = 4\%$ of $\bar{w}$	3.11, 3.42, 2.64, 2.60, 3.41, 2.75; $\bar{d} = 2.99$, $\sigma = 0.38$, $\sigma = 13\%$ of $\bar{d}$	177, 185, 142, 148, 201, 146; $\bar{A} = 166.7$, $\sigma = 24.5$, $\sigma = 15\%$ of $\bar{A}$
Rio Santa Barbara Station 57	20, 17, 19, 17, 18, 20; $\bar{w} = 18.5$, $\sigma = 1.38$, $\sigma = 7\%$ of $\bar{w}$	1.13, 1.60, 1.74, 1.17, 1.06, 1.12; $\bar{d} = 1.30$, $\sigma = 0.29$, $\sigma = 22\%$ of $\bar{d}$	23, 27, 33, 20, 19, 22; $\bar{A} = 24.0$, $\sigma = 5.2$, $\sigma = 22\%$ of $\bar{A}$
Rio Santa Barbara Station 71			
Section 1*	31	1.41	44
2	34	1.37	47
3	39.5	1.62	64
4	38	1.24	47
5	30.5	1.26	38
6	26	1.71	44
7	32	1.27	41
	$\bar{w} = 33.0$, $\sigma = 4.63$, $\sigma = 14\%$ of $\bar{w}$	$\bar{d} = 1.41$, $\sigma = 0.18$, $\sigma = 13\%$ of $\bar{d}$	$\bar{A} = 46.4$, $\sigma = 8.4$, $\sigma = 17\%$ of $\bar{A}$

* Section numbers on map (fig. 6). For explanation of symbols, see p. vi.

phenomenon. At a few places, measuring stations were spaced close enough so that meandering of the thalweg could be mapped. One such example is shown in the map of Figure 6.

TABLE 2. RELATIVE POSITION OF THE THALWEG, OR DEEPEST POINT IN CHANNEL CROSS-SECTION

(Percent of cases)

	CENTER 1/5 OF CHANNEL	INTERMEDIATE 1/5 OF CHANNEL	OUTERMOST 1/5 OF CHANNEL (NEAREST BANK)
Pecos River (68 stations)	26	50	24
Rio Santa Barbara (23 stations)	22	65	13
Rio Santa Cruz (13 stations)	54	46	0
Pojoaque River (22 stations)	23	64	13
All stations (126)	28	55	17

CHANGES AT JUNCTIONS

Many of the measuring stations were located above and below stream junctions so as to determine the nature of changes in channel dimensions caused by the joining of tributaries. The data for mountain streams are summarized in Table 3. Comparisons of channel dimensions above and below stream junctions are based on the coefficient k in the expression $a = k\,(b + c)$, where b and c are tributaries, and a is the main stream below the junction. Thus, if dimensions were simply additive, k would be equal to one. Data from stations at 19 stream junctions indicate that bankfull width below a junction ranges from 1/2 to 5/6 and averages 2/3 of the sum of the tributary widths. Despite a few cases of appreciable departures from the mean value of 0.66, the overall dispersion is low.

Considering individual sets of stations at junctions, it should be noted (table 3) that in 80 percent of the cases width below the junction is equal to or greater than the larger

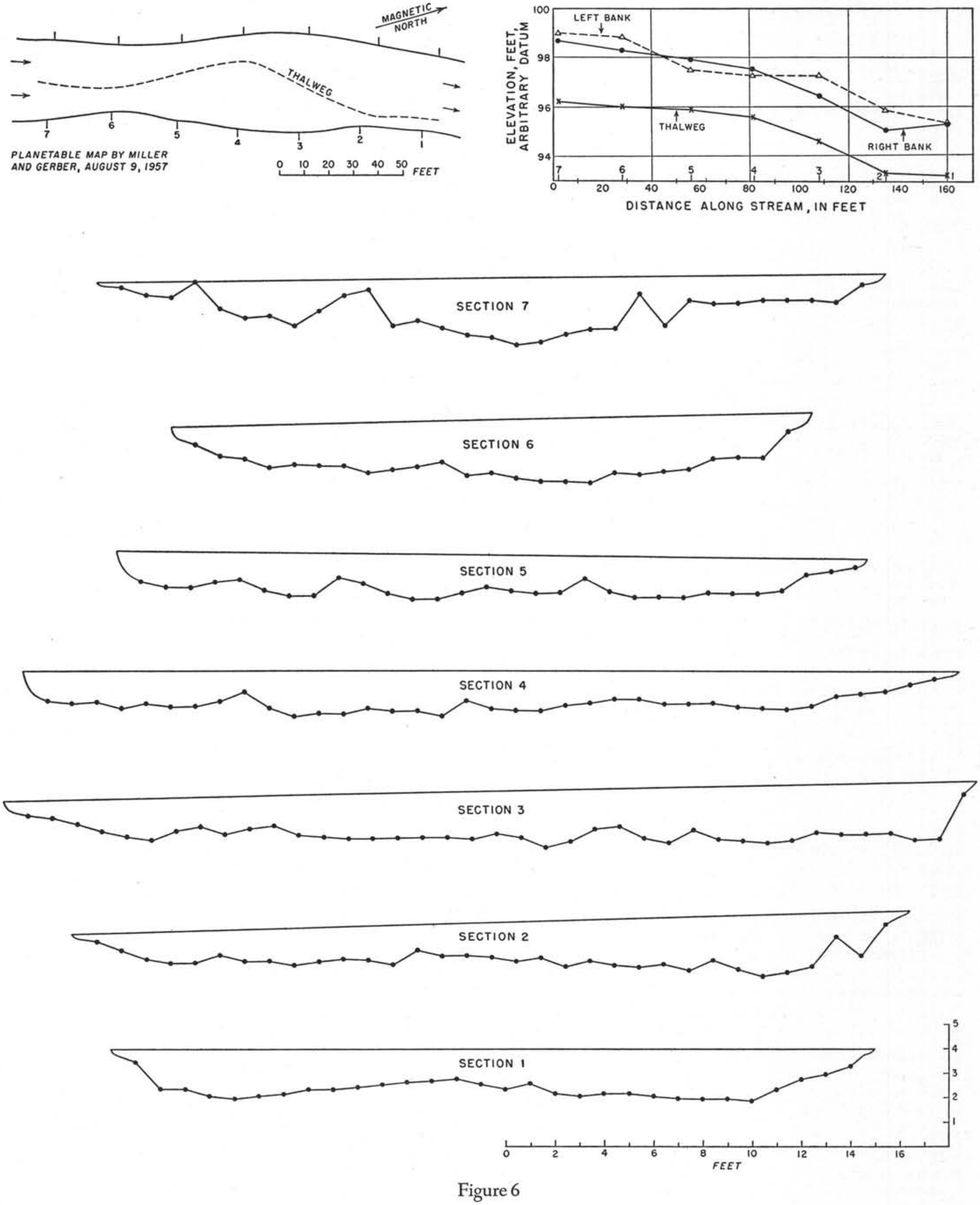


Figure 6

PLAN, PROFILE, AND CROSS-SECTIONS OF THE RIO SANTA BARBARA AT STATION 71

The magnitude of local variations in channel width and depth is shown by the cross-sections. Most irregularities in the bed of the channel are due to the presence of large boulders. Meandering of the thalweg can be seen on the plan. Also, the profile shows that the stream banks are of unequal height at most places.

TABLE 3. CHANGE IN BANKFULL WIDTH, MEAN DEPTH, AND CROSS-SECTIONAL AREA AT STREAM JUNCTIONS

The value k is a coefficient in the expression $a = k\,(b + c)$, where b and c are tributaries, and a is the main stream below the junction.

STREAM AND STATION NUMBER	BANKFULL WIDTH				BANKFULL DEPTH				CROSS-SECTIONAL AREA			
	a	b	c	k_w	a	b	c	k_d	a	b	c	k_A
Pecos River												
1	76	43	48	0.83	2.2	2.8	2.2	0.44	167	120	106	0.74
2	35	15	28	0.81	2.1	2.1	1.9	0.52	73	31	53	0.87
3	19	20	17	0.51	1.6	1.7	1.8	0.64	30	34	31	0.46
4	18	12	14	0.69	1.1	1.3	1.4	0.41	20	16	20	0.55
5	25	28	18	0.54	1.8	2.0	1.8	0.47	45	56	32	0.51
9	37	39	19	0.64	1.5	1.5	1.2	0.56	55	58	23	0.68
10	25	18	22	0.62	2.1	1.8	1.7	0.84	52	32	37	0.75
12	23	15	24	0.59	1.7	1.1	1.6	0.63	39	16	38	0.72
14	24	24	9	0.73	1.3	1.1	1.2	0.56	31	26	11	0.84
37	16	15	9	0.67	0.8	0.8	1.1	0.42	13	12	10	0.59
39 and 50	17	9	15	0.71	1.0	1.0	1.4	0.42	17	9	21	0.57
61	14	9	13	0.64	1.6	1.2	1.1	0.70	22	11	14	0.88
62	14	9	8	0.82	1.4	1.2	0.9	0.67	20	11	7	1.11
	$\bar{k}_w = 0.68$, $\sigma = 0.10$				$\bar{k}_d = 0.56$, $\sigma = 0.13$				$\bar{k}_A = 0.71$, $\sigma = 0.18$			
Rio Santa Barbara												
56	29	24	24	0.60	1.6	2.2	2.1	0.37	46	53	50	0.44
59	12	9	10	0.63	1.0	1.0	0.9	0.53	12	9	9	0.67
64	16	12	12	0.64	1.1	1.2	1.0	0.50	18	14	12	0.69
	$\bar{k}_w = 0.62$, $\sigma = 0.02$				$\bar{k}_d = 0.47$, $\sigma = 0.09$				$\bar{k}_A = 0.60$, $\sigma = 0.14$			
Rio Santa Cruz												
15	13	9	11	0.65	1.3	1.3	1.3	0.50	17	12	14	0.65
53	24	23	16	0.61	1.4	1.3	1.8	0.45	34	30	29	0.58
	$\bar{k}_w = 0.63$				$\bar{k}_d = 0.48$				$\bar{k}_A = 0.62$			
Pojoaque River												
22	19	10	18	0.68	1.2	1.4	1.2	0.46	23	14	22	0.64
	Grand mean: $\bar{k}_w = 0.66$, $\sigma = 0.09$				$\bar{k}_d = 0.53$, $\sigma = 0.12$				$\bar{k}_A = 0.68$, $\sigma = 0.16$			

tary width. For the remaining cases, width below the junction is intermediate between the widths of the tributaries.

Data for changes in width at 28 arroyo junctions located at various places in the Rio Grande Depression are given in Table 4. The mean value of k for arroyos is nearly the same as for mountain streams, but the dispersion is slightly greater. In nearly 80 percent of the cases for arroyos, width below a junction is equal to or greater than the larger tributary width; furthermore, nearly all the exceptions are very small arroyos.

A value of k for widths only slightly different from the measured value could have been predicted from the rate of downstream increase in channel width with increasing discharge. Available data indicate that for streams in a great variety of physiographic and climatic situations $w = Q^{0.5}$ approximately (Leopold and Maddock, 1953; Wolman, 1954; Leopold and Miller, 1956; Brush, in press).

At any junction,

$$Q_a = Q_b + Q_c, \qquad (1)$$

$$w_a = k(w_b + w_c). \qquad (2)$$

Assuming

$$w = Q^{0.5}, \qquad (3)$$

and substituting in (2),

$$Q_a^{0.5} = k\,(Q_b^{0.5} + Q_c^{0.5}), \qquad (4)$$

$$k = \frac{Q_a^{0.5}}{Q_b^{0.5} + Q_c^{0.5}} = \frac{(Q_b + Q_c)^{0.5}}{Q_b^{0.5} + Q_c^{0.5}}. \qquad (5)$$

If $Q_b = Q_c$, then $k = 0.71$. If $Q_b = 2\,Q_c$, which is the maximum difference in tributary discharges for data in Tables 3 and 4, then $k = 0.72$.

Mean bankfull depths of mountain streams below junctions range from 1/3 to 5/6 and average about 1/2 the sum, or in other words, the mean of the two tributary depths. Dispersion about this mean value of it for depth is somewhat

TABLE 4. CHANGES IN CHANNEL WIDTHS OF ARROYOS AT JUNCTIONS

The value k is a coefficient in the expression $a = k\ (b + c)$, where b and c are tributaries, and a is the stream below the junction.

a	b	c	k
41.5	35	21	0.74
31	28	15	.72
13	12	6	.72
10	7	6.5	.74
39	26	22	.81
1.9	2.0	1.6	.53
7.5	5.0	3.0	.94
13	16	13	.45
62	40	53	.67
18	16.5	6.5	.78
24	33	11	.54
40	27	41	.59
9	5.5	7	.72
32	21	17	.84
803	556	428	.82
6	4	5.5	.63
9	10	5	.60
5	4	5	.55
2.1	2.2	1.2	.62
1.8	2.2	2.5	.38
172	62	268	.52
72	46	50	.75
49	28	29	.86
14	6	13	.74
4.0	4.0	3.5	.53
7.5	5.0	4.5	.79
7.0	5.5	5.0	.67
16	10	12	.73
	$k = 0.68;$	$\sigma = 0.13$	

greater than in the case of width. The character of changes in depth at individual sets of stations at junctions is summarized below.

BANKFULL DEPTH BELOW JUNCTIONS

= OR > LARGER TRIBUTARY DEPTH	INTERMEDIATE BETWEEN TRIBUTARY DEPTHS	< DEPTH OP EITHER TRIBUTARY
42%	50%	8%

It has been determined that for many streams the rate of increase of mean depth with downstream increases in discharge can be expressed as $d \propto Q^{0.4}$ approximately. Proceeding as in the example for width above, the computed value of k is much higher than the measured value. Even if the exponent relating depth to discharge were 0.25 instead of 0.4, the value of k would be about 0.60.

The data presented above suggest that systematic kinds of changes in channel dimensions occur at stream junctions. However, the relatively small samples considered include several examples of appreciable departures from these idealized relationships.

Results of width and depth measurements at junctions show fairly good agreement with data on divided and undivided segments of channels in braided reaches obtained by Rubey (1952) and Leopold and Wolman (1957).

Relation to Drainage Area

Plots of bankfull width and mean depth versus drainage area for each of the basins studied are shown in Figure 7. Although there is considerable scatter, the line of best fit through the plotted points of each graph is a straight line, which on double logarithmic paper defines a power-function relation. Needless to say, the scatter of points in these graphs for entire drainage basins could be reduced by plotting each tributary separately, but doing this would not alter the general character of the relationships involved.

It should be noted (fig. 7) that in all four basins width increases more rapidly than depth; in other words, the width-depth ratio becomes greater downstream.

The rate of increase of both width and depth for the Rio Santa Barbara and the Pecos River is much greater than for the Rio Santa Cruz and the Pojoaque River. As will be discussed later (p. 40), this difference may be related in part to geologic factors, for the Pecos and Santa Barbara flow mostly on sandstone, whereas the other two streams flow across terrain underlain by granite.

ESTIMATION OF BANKFULL DISCHARGE

Having described the cross-sectional dimensions of channels at the various measuring stations, the next step is to consider the bankfull discharge at these points. It should be obvious that there is no practical alternative to estimating bankfull discharge, because its actual measurement would require a gaging-station record covering a period of several years at many of the 104 sampling stations considered here. Following a procedure commonly used in engineering practice and described in the geological literature by Leopold and Miller (1956) and Brush (in press), available gaging-station records can be used to establish a relation between drainage area and discharge of a particular frequency. In this case the frequency sought is the one which corresponds to the bankfull stage. It is not definitely proved that the bankfull stage occurs with the same frequency throughout the same drainage basin. Available evidence indicates, however, that the mean annual flood (recurrence interval 2.3 years) approximates the bankfull stage in a wide variety of geographical situations (Wolman, 1954; Wolman and Leopold, 1957; Brush, in press). Tentatively, this relation is assumed to be applicable to Sangre de Cristo streams.

Although nearly two dozen gaging stations with long records are located within a 50-mile radius of the area considered here, most of them are in the lowlands rather than the mountains. Furthermore, several of the mountain stations, including some of those in the Taos Mountains a short distance to the northeast, are unsatisfactory for flood-frequency studies because they are downstream from major irrigation diversions.

Eight gaging stations, located in or near the mountains and above irrigation diversions, were selected for frequency analysis by the annual-flood[1] method (Langbein, 1949). Four of these stations are located in the portion of the Sangre de Cristo Range considered in the present study; the others are in the Taos Mountains a short distance to the north. The discharge, which has a recurrence interval of 2.3 years, was determined for each station, and the results are summarized in Table 5. These values of discharge are plotted against drainage area in Figure 8. The solid line, which is drawn through the points for stations actually in the area studied, has the equation $Q_{2.3} = 18\ A_d^{0.72}$. It should be noted that this relation

1. The annual flood is the maximum discharge recorded during each year. The mean annual flood (average of all annual floods in the record), which has an average recurrence interval of about 2.3 years, should not be confused with the mean annual discharge.

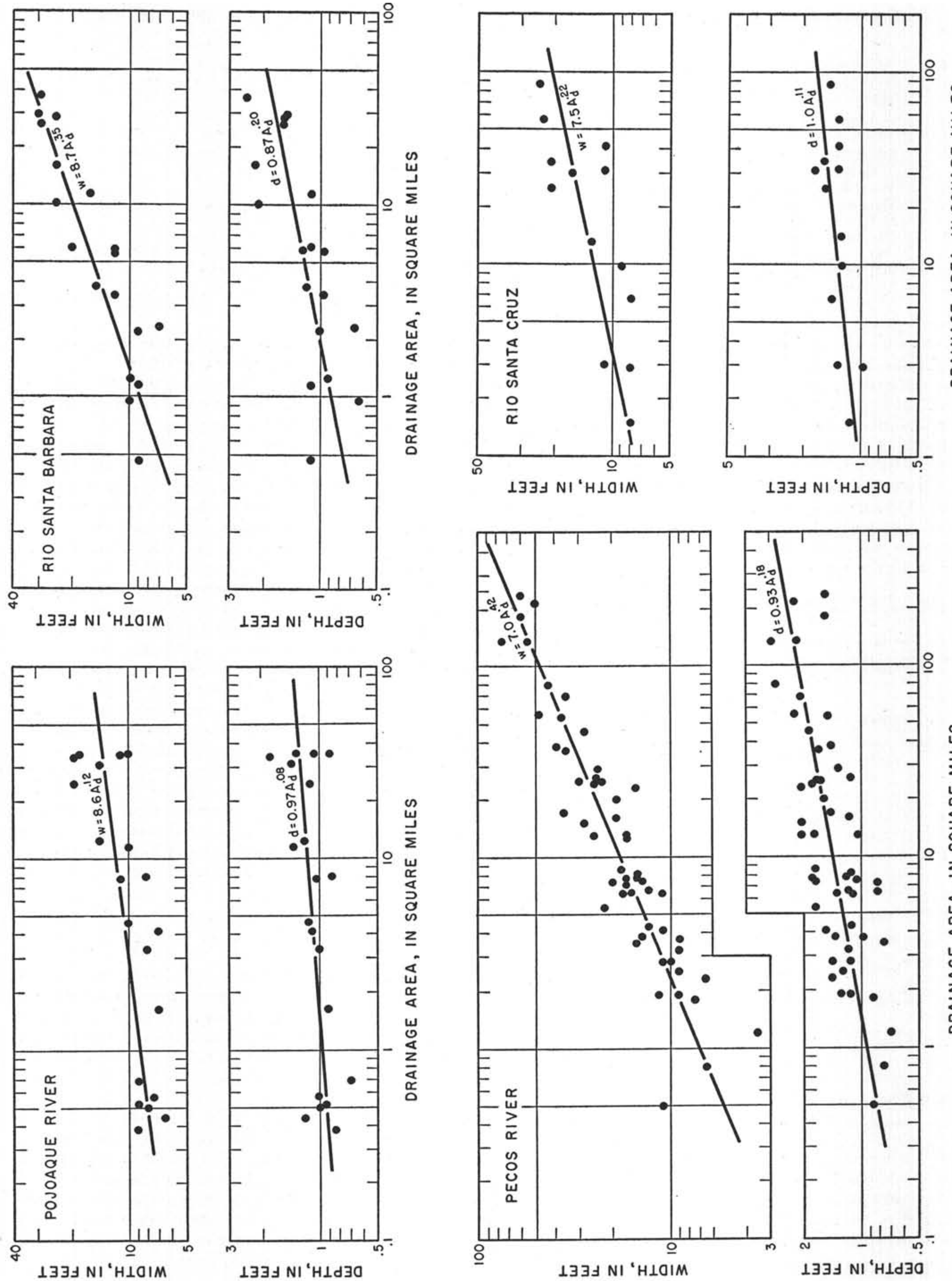


Figure 7

Relation of bankfull width and mean depth to drainage area for all stations in each of the four drainage basins studied

TABLE 5. FLOOD-FREQUENCY DATA FOR GAGING STATIONS IN THE SANGRE DE CRISTO RANGE

NUMBER	STREAM AND LOCATION	ALTITUDE (feet)	DRAINAGE AREA (sq mi)	DISCHARGE (2.3-YEAR FREQUENCY) (cu ft/sec)	YEARS OF RECORD
1	Rio Tesuque, near Santa Fe	7,100	11	112	15
2	Rio Nambe, near Nambe	6,200	37	210	17
3	Rio Santa Cruz, at Cundiyo	6,460	86	490	25
4	Pecos River, near Pecos	7,505	189	690	25
5	Rio Lucero, near Arroyo Seco	8,000	17	152	26
6	Red River, near Red River	9,395	19	113	15
7	Rio Hondo, near Valdez	7,700	38	177	19
8	Rio Pueblo de Taos, near Taos	7,110	56	305	16

between discharge and drainage, based on data from only four stations, conforms closely with results obtained by using other combinations of available gaging records (Leopold and Miller, 1956, p. 24).

Bankfull discharge at each sampling station was estimated by measuring drainage area above the station and reading the discharge for that drainage area off the curve (fig. 8). It should be emphasized that there are several uncertainties involved in this method. As was indicated earlier, discharge of 2.3-year frequency may not correspond exactly to the bankfull stage, nor is there definite proof that bankfull discharges occur with exactly the same frequency throughout a drainage basin.

TABLE 6. ESTIMATED BANKFULL DISCHARGES AT STREAM JUNCTIONS

Because *a* is the main stem below the junction, and *b* and *c* are tributaries, $(b + c)$ must equal *a* in order for continuity to be maintained.

STREAM AND STATION NUMBER	ESTIMATED BANKFULL DISCHARGE *a*	*b*	*c*	$(b + c)$
Pecos River				
1	617	420	325	745
2	376	173	280	453
3	156	76	114	190
4	68	42	40	82
5	178	128	85	213
9	318	247	130	377
10	115	79	61	140
12	183	82	138	220
37	69	44	40	84
39 and 50	113	46	88	134
61	77	41	52	93
62	47	29	28	57
Rio Santa Barbara				
56	190	135	95	230
59	44	32	21	53
64	105	64	63	127
Rio Santa Cruz				
15	114	94	40	134
53	450	325	209	534
Pojoaque River				
22	180	104	115	219

In this particular case, there is another difficulty: the gaging stations, on which the relationship between discharge and drainage area is based, have considerably lower elevations than most of the sampling stations at which it is desired to estimate discharge. Furthermore, the extent to which geologic factors, together with increased precipitation and more dense cover of vegetation at higher altitudes, may affect the frequency of bankfull flooding cannot be evaluated.

Another kind of problem associated with estimating discharge from a line drawn through scattered points on the drainage area-discharge graph is demonstrated by the data in Table 6. Discharges obtained by this method do not add up at stream junctions as required by continuity considerations; that is, the discharge of a main stem below a junction is about 17 percent less than the sum of discharges in the two tributaries. This inequality comes about because the slope of the discharge-drainage area curve is less than one. Continuity can be maintained at stream junctions only if the slope of this curve is equal to one (discharge exactly proportional to drainage area). This means that in detail the discharge-drainage area curve is not a straight line, as drawn, but goes up as a series of steps. The short risers on this sloping stairlike line occur at stream junctions, where discharge is proportional to drainage area, and the treads of this stairlike curve are essentially flat, because little discharge enters the channel in the reaches between tributary junctions. The average curve drawn through the stair-step arrangement of points defines discharge as being proportional to the 0.7-0.8 power of drainage area, as shown in Figure 8. Adjusting the values of discharges at stream junctions so as to fulfill continuity requirements would involve arbitrary procedures, and for this reason will not be attempted. However, the discrepancies at junctions will not seriously affect the types of analyses which are to made.

DOWNSTREAM CHANGES IN BANKFULL WIDTH, DEPTH, AND VELOCITY

Using bankfull discharges estimated by the method described in the previous section, and measured values of bankfull width and mean depth, the mean bankfull velocity at each sampling station was computed from the continuity equation

$$w\,d\,v = Q. \qquad (6)$$

The values obtained are given in Tables 15-18.

Downstream changes in width, depth, and velocity which accompany downstream increases in discharge (of constant frequency) constitute one part of the hydraulic geometry, described by Leopold and Maddock (1953). The other portion of the hydraulic geometry, at-a-station relations for the same variables, requires gaging information which is not available for the sampling stations considered here.

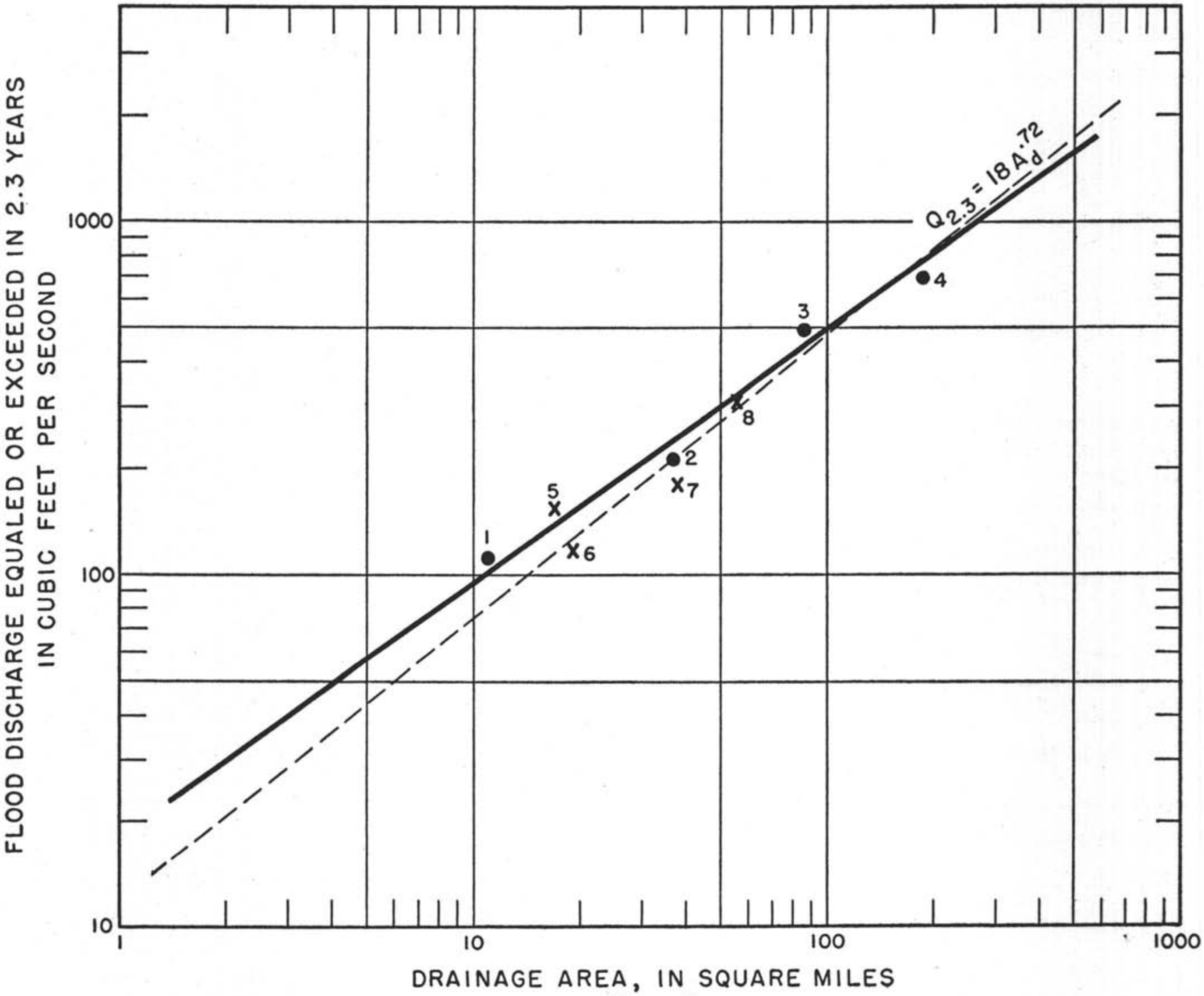


Figure 8

RELATION OF DISCHARGE TO DRAINAGE AREA, BASED ON DATA FROM GAGING-STATION RECORDS

Points 1-4, through which the solid line is fitted, are stations in the area studied. The other four stations are in the Taos Mountains. The dashed line, which has the equation $Q_{2.3} = 12\ A_d^{0.72}$, shows the relation found by Leopold and Miller (1956), using both mountain and lowland stations. Recurrence interval of discharge, 2.3 years.

Values of bankfull width, depth, and velocity are plotted against bankfull discharge in Figure 9, which includes a separate graph for each of the four drainage basins. The amount of scatter among the plotted points for width and depth is about the same as for cases in which discharge was actually measured (Leopold and Maddock, 1953; Wolman, 1954). Scatter among the points of velocity is somewhat greater, as might be expected from the method by which velocity was obtained. There can be little doubt, however, that straight lines best fit the plotted points in each graph.[2] These lines define power functions with the following general equations:

$$w = a\,Q^b, \qquad (7)$$
$$d = c\,Q^f, \qquad (8)$$
$$v = k\,Q^m. \qquad (9)$$

2. Curves were fitted by eye. Notice in each case that the sum $b + f + m \simeq 1$, and the product $a\,c\,k \simeq 1$.

This means that the general nature of downstream relations of width, depth, and velocity to discharge for high mountain streams is similar to that reported previously for streams in less rugged terrain (Leopold and Maddock, 1953; Wolman, 1954; Leopold and Miller, 1956; Brush, in press.)

Values for rates of change of width, depth, and velocity (the exponents in the power functions) shown in Figure 9 can be checked by considering the interrelationships of these variables with drainage area. Using the width of the Rio Santa Barbara as an example,

from Figure 7, $w \propto A_d^{0.35}$,
from Figure 8, $Q \propto A_d^{0.72}$;
therefore, $w \propto Q^{0.49}$

By this method, exponents expressing the downstream rates of change of width and depth can be computed directly. The

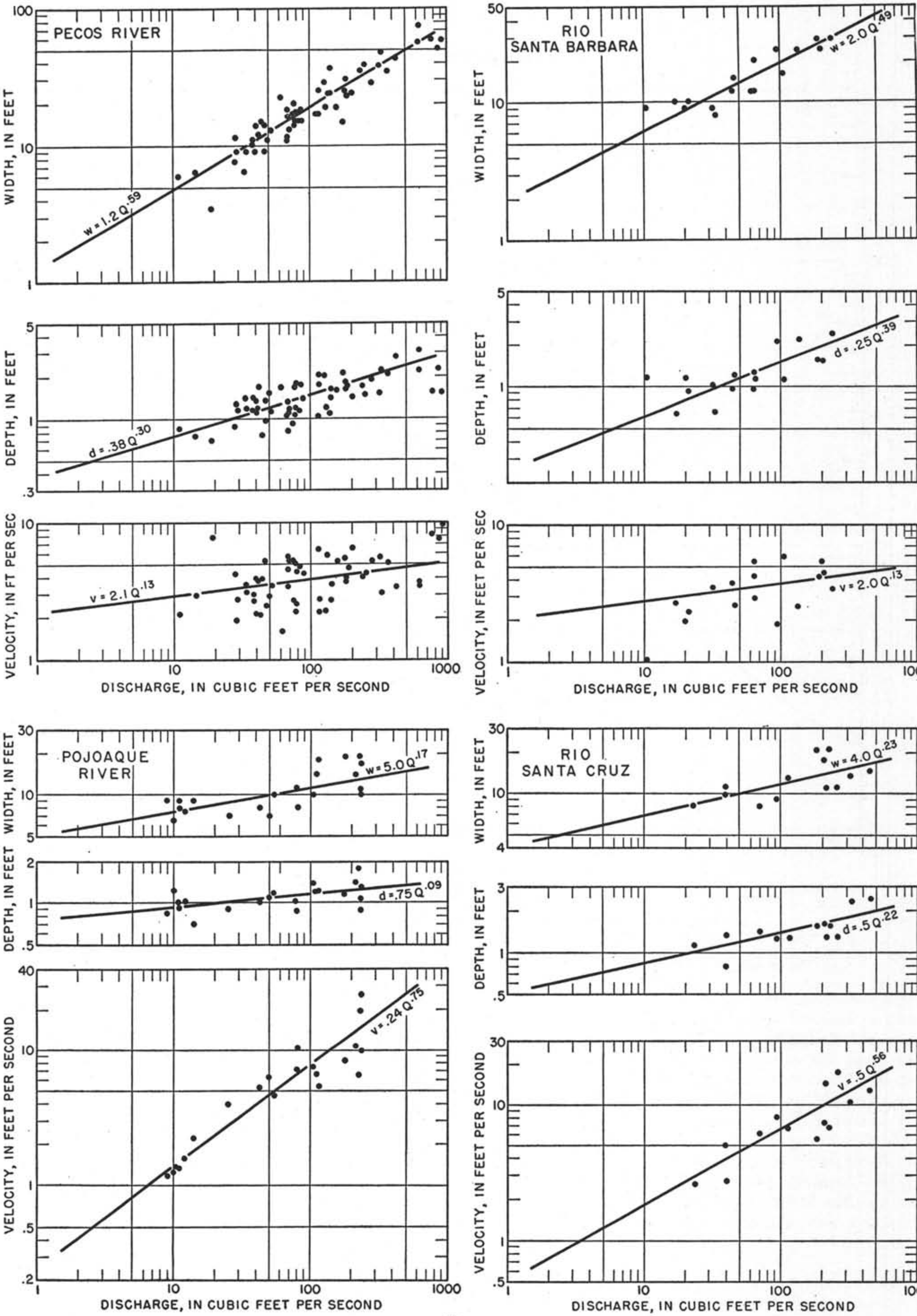


Figure 9

DOWNSTREAM CHANGES IN BANKFULL WIDTH, MEAN DEPTH, AND VELOCITY ACCOMPANYING DOWNSTREAM INCREASES IN DISCHARGE OF CONSTANT FREQUENCY

Width and depth were measured. Discharge was estimated from Figure 8, and velocity was computed from continuity requirements.

exponent for velocity can be obtained by difference because of the relations required for continuity:

$$w\,d\,v = Q.$$

Substituting from equations (7), (8), and (9),

$$(a\,Q^b)(c\,Q^f)(k\,Q^m) = Q,$$
$$b + m + f = 1$$

It is possible to check the exponents for downstream changes of width, depth, and velocity in another way because of the fact that width and depth are related to stream length by power functions. Graphs showing this relation were plotted but are not included here. Again using the width of the Rio Santa Barbara as an example,

$$W \;\alpha\; L^{0.58};$$
from Figure 4, $L \;\alpha\; A_d^{0.54}$;
from Figure 8, $Q \;\alpha\; A_d^{0.72}$;
therefore, $w \;\alpha\; Q^{0.44}$;

For this particular case, both checks involving interrelationships of other variables indicate that the line drawn through the points in Figure 9 is approximately correct.

The most interesting feature of Figure 9 is that downstream changes of width, depth, and velocity in the Rio Santa Barbara and the Pecos River are quite different from changes of the same variables in the Rio Santa Cruz and the Pojoaque River. Comparison with data from earlier studies (table 7) indicates that the values of exponents expressing rates of downstream changes in the Rio Santa Cruz and the Pojoaque River are quite anomalous. Unfortunately, the significance of these differences cannot be evaluated satisfactorily. If the assumptions which underlie the method used for determining bankfull discharge are valid, the slight downstream increase in channel dimensions of the Rio Santa Cruz and the Pojoaque River requires a very large rate of increase in velocity. An alternative, but unproved, explanation is that the bankfull stage is not attained with equal frequency in different drainage basins. However, it is significant in this connection that gaging measurements of bankfull discharge and velocity at stations in the Rio Santa Barbara basin agree remarkably well with values estimated by the methods described in this report. No gaging measurements of the Rio Santa Cruz and the Pojoaque River at bankfull stage are available for comparison with estimated values of discharge and velocity, but the calculated velocities definitely seem too high, in that some of them exceed maximum point velocities measured in larger rivers. However, despite the fact that some absolute values of velocity and discharge plotted in Figure 9 may be in error, the relative rates of downstream changes in width, depth, and velocity probably are approximately correct.

TABLE 7. COMPARISON OF EXPONENTS IN EQUATIONS RELATING WIDTH, DEPTH, AND VELOCITY TO DOWNSTREAM INCREASE IN DISCHARGE:
$w = a\,Q^b$; $d = c\,Q^f$; $v = k\,Q^m$

		b	f	m
Leopold and Maddock (1953), western streams; mean annual discharge				
	Range	0.46–0.76	0.09–0.44	0.06–0.15
	Median	0.50	0.40	0.10
Wolman (1954), Brandywine Creek, Pennsylvania; bankfull stage = $Q_{2.3}$				
	Average	0.42	0.45	0.05
Leopold and Miller (1956), arroyos in New Mexico; frequency of discharge unknown				
	Average	0.50	0.28	0.22
Brush (in press), Appalachian streams in Pennsylvania; bankfull stage = $Q_{2.3}$*				
	Range	0.30–0.89	0.29–0.70	0.51–0.29
	Average	0.55	0.36	0.09
Miller (this paper); bankfull stage = $Q_{2.3}$*				
	Rio Santa Barbara	0.49	0.39	0.13
	Pecos River	0.59	0.30	0.13
	Rio Santa Cruz	0.23	0.22	0.56
	Pojoaque River	0.17	0.09	0.75

* Discharges estimated from drainage area-discharge relation.

As was mentioned earlier, the Rio Santa Barbara and the Pecos River flow over sandstone, whereas the Rio Santa Cruz and the Pojoaque River flow on granite. The possibility that hydraulic properties of the various streams are affected by these lithologic differences or other geological factors will be discussed in a later section of this paper (p. 40).

ANGLE OF TRIBUTARY JUNCTIONS

The angle between tributaries at 19 junctions was measured in the field and checked on air photos. These measurements refer only to the distances visible from a junction, generally only a few hundred feet. The data are summarized below:

	ANGLE BETWEEN TRIBUTARIES (DEGREES)			
	51-60	61-70	71-80	81-90
Pecos River	2	3	2	6
Rio Santa Barbara	2	1	—	—
Rio Santa Cruz	—	—	1	1
Pojoaque River	—	1	—	—
Totals	4	5	3	7

On the basis of this sample, it seems evident that tributaries meet at rather large angles. The range of values obtained was 56-88 degrees. Considering all four basins together, the values of angle are distributed fairly evenly among the various classes.

The data for junction angles presented above are for tributaries of roughly equal size. Field observations and studies of air photos indicate that small tributaries consistently enter the main streams at angles between 80 and 90 degrees.

CHANNEL SLOPE

It was mentioned previously that streams in the Sangre de Cristo Range have steep and rather irregular gradients. Generally, channel slopes are steepest near the headwaters divides and flatten gradually downstream, but there are many local exceptions. For example, exceedingly flat gradients are associated commonly with high mountain meadows (pl. 5A, station 57). In general, the most abrupt changes in channel slope occur at stream junctions, at lithologic contacts, and especially at erosional and depositional features produced by glaciation.

In their studies of Appalachian streams, both Hack (1957) and Brush (in press) noted a power-function relation between channel slope and stream length. Most commonly, they found that slope was exactly inversely proportional to stream length; that is, $s \propto L^{-1}$. Graphs of channel slope plotted against stream length for Sangre de Cristo streams were prepared but are not included in this paper. Despite considerable scatter, slope is related to length by power functions, but in most cases the exponents are considerably less than The smaller exponent means that Sangre de Cristo streams, which have steeper gradients because of the greater relief, have profiles which are less concave than Appalachian streams.

Wolman (1954) expressed downstream change in slope as a function of discharge:

$$s = t\,Q^{z}. \qquad (10)$$

For Brandywine Creek, Pennsylvania, the exponent z for bankfull discharge is equal to —1. Values of z for Sangre de Cristo streams were obtained by plotting slope against bankfull discharge for all the stations in each drainage basin.[3] These graphs are not shown, but the results are summarized below:

	z
Rio Santa Barbara	- 0.36
Pecos River	- 0.63
Rio Santa Cruz	- 0.30
Pojoaque River	- 0.48

It should be remembered that these are slope measurements of the water surface at considerably less than bankfull stage. However, despite slightly increased slopes at higher discharges, the values of z for Sangre de Cristo streams are considerably smaller than those reported by Wolman for Brandywine Creek.

Changes in channel slope at stream junctions are summarized in Table 8. Slope below a junction ranges from 1/5 to 1/2 and averages 1/3 of the sum of the tributary slopes. In slightly more than half the cases, slope below the junction is less than the lesser tributary slope; in no case is it greater than the slope of the steeper tributary.

TABLE 8. CHANGES IN CHANNEL SLOPE AT STREAM JUNCTIONS

The value k is a coefficient in the expression $a = k\ (b + c)$, where b and c are tributaries, and a is the main stream below the junction.

STREAM AND STATION NUMBER	CHANNEL SLOPE *a*	*b*	*c*	*k*
Pecos River				
1	0.013	0.009	0.028	0.35
2	.008	.018	.016	.23
3	.032	.038	.022	.53
5	.026	.048	.048	.27
9	.019	.030	.042	.26
10	.020	.054	.039	.21
12	.027	.047	.017	.42
37	.048	.084	.042	.38
61	.026	.103	.036	.19
62	.058	.091	.058	.39
Rio Santa Barbara				
56	0.041	0.053	0.032	0.48
59	.048	.060	.032	.52
64	.036	.044	.043	.41
Rio Santa Cruz				
15	0.038	0.034	0.039	0.52
53	.022	.032	.030	.35
Pojoaque River				
22	0.036	0.052	0.057	0.34

Grand mean: $\bar{k} = 0.34$; $\sigma = 0.14$

3. If possible, values obtained graphically were checked, using the interrelationships between slope, stream length, drainage area, and discharge. Using the Pecos River as an example, $s \propto L^{-0.84}$; $L \propto A_d^{0.64}$; $Q \propto A_d^{0.72}$; $s \propto Q^{-0.63}$.

Bed Material

Several general characteristics of the streambed material are apparent even from a superficial examination. At many places along most streams, it is obvious that rock fragments which comprise the bed material range widely in size and commonly include several different lithologies. Rock fragments from nearby sources constitute a large fraction of the bed material at most places, and particles derived from a particular outcrop become less abundant downstream, because of dilution by local contributions. Where rock is exposed in or very near the channel, coarsening due to enrichment of the bed material with large fragments from this source is generally noticeable. The effects of tributaries which deliver bed material appreciably different in size or composition from that of the main stem are qualitatively apparent.

Although a downstream decrease in particle size of bed material can be recognized qualitatively in several streams, this trend generally undergoes several reversals, owing to coarsening at bedrock outcrops or tributary junctions. Downstream increases in roundness of particles recognized as having come from a particular distinctive outcrop are easily seen.

PROBLEMS IN SAMPLING

Because of its heterogeneous character with respect to size and lithology, precise description of bed material is rather difficult. The problem of how to obtain a truly representative sample from streambeds has received little attention. By comparison with difficulties in selecting the sample, measurement of particle size is exceedingly simple. Whatever the method adopted for measuring size, a large volume of raw data must be reduced to summary statistics; most commonly, the mean or median and some measure of dispersion are used. Size distribution of each sample also may be shown graphically by means of histograms, or cumulative curves, but if large numbers of samples are involved, this method becomes impractical. Lithologic composition of bed material seems most conveniently expressed as percentages of the various rock types in the samples.

The method used for sampling and measuring bed material was described in the section on basic data. It should be added here that the precision of this method has been thoroughly tested (Miller and Brush, in preparation). Repeated measurements of the same sample of 100 cobbles and boulders by 5 different persons yielded values of mean size which are not significantly different at the 5-percent level. Furthermore, 5 persons sampled the same reach of stream with no differences in sample means at the 5-percent level of significance. All measurements in the present study were made by the same person, but the results of these preliminary tests indicate that several individuals might have assisted without seriously affecting the results obtained.

It is obvious that the procedure used provides a sample of only the uppermost portion of the bed material. To obtain a sample through a thickness equal to a few particle diameters would require a power shovel, because at many stations the particles are wedged tightly together and a few boulders are too heavy to be lifted or rolled. The kind of sample obtained seems appropriate, however, for the purposes of this investigation, because the bed material which is in contact with the water should exert the greatest influence on channel properties.

The beds of streams at all stations included in this investigation are covered with particles of pebble size or larger (see pl. 9, for example). Measurements of bed material in the Rio Santa Cruz and the Pojoaque River were terminated at the mountain front, inasmuch as sampling of the dominantly sandy sediment in arroyos cannot be handled by the procedures used for coarse material.

Typical size distributions of bed material in Sangre de Cristo streams are shown in the cumulative curves of Figure 10. Although the specific ranges of size in the various samples differ appreciably, the distributions in all cases are approximately log-normal; that is, they plot as nearly straight lines on logarithmic probability paper.

Tests were made to determine the optimum number of particles to measure at each station. The results, which are treated in detail by Miller and Brush (in preparation), are summarized briefly here. As would be expected, increasing the sample size decreases dispersion around the mean value. The variance (standard deviation squared) for size is approximately inversely proportional to the cube root of the number of particles measured. This is true for samples ranging from 50 to 250 particles. These sediments, however, are poorly sorted, and dispersion cannot be reduced below a certain minimum, regardless of sample size.

For purposes of this study, the number of particles included in samples of bed material obtained at the various measuring stations was based in part on results of the tests discussed above, as well as on practical considerations. The time factor is an important one, as from 30 to 45 minutes is required for measuring each sample of 50 particles. Also, differences in stream width introduce complications, because it is impossible to collect a very large sample from narrow headwaters by the methods described. The decision to take samples consisting of from 50 to 100 particles represents a necessary compromise between achievement of desirable precision and coverage of a reasonably large number of stations.

DOWNSTREAM CHANGES IN SIZE AND LITHOLOGIC COMPOSITION

Mean particle size shows no well-defined relation to stream length or drainage area (and, therefore, discharge) if all stations in the drainage basin are plotted. The effects of bedrock outcrops in or near stream channels, plus differences in the sizes of fragments derived from various rock types and from the numerous tributaries, combine to cause a very large amount of scatter in such diagrams.

In different streams, mean particle size of bed material at the station farthest downstream may be larger, smaller, or approximately the same, as compared with the station closest to the drainage divide. Whatever the character of this overall trend, any detailed picture of downstream change includes numerous fluctuations of both particle size and lithologic composition, which generally occur either at stream junctions or lithologic contacts.

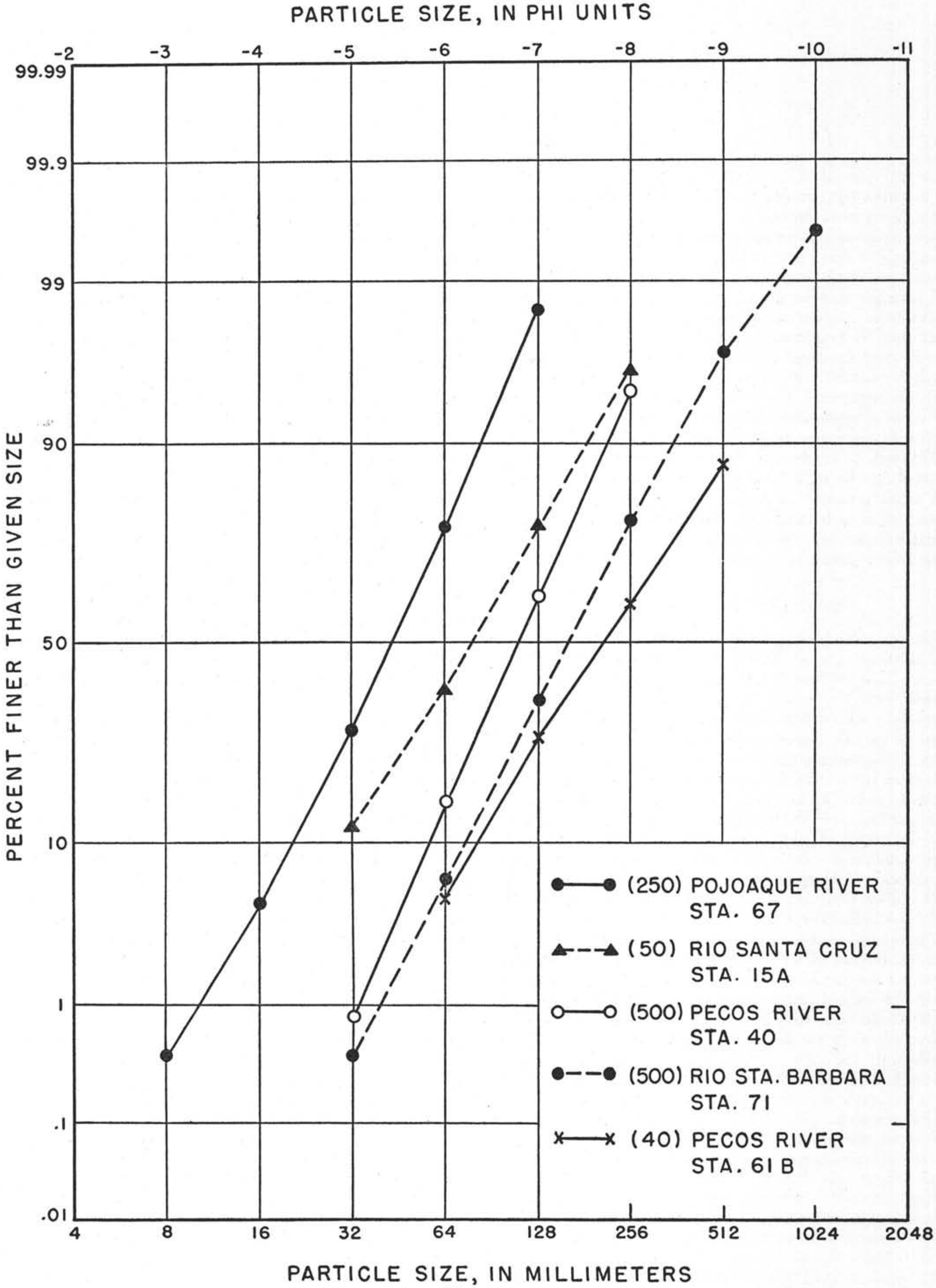


Figure 10

Size-distribution curves for typical samples of bed material in channels of Sangre de Cristo streams

The number of particles measured at each station is shown in parentheses.

CHANGES AT JUNCTIONS

Changes in mean size of bed material at stream junctions are summarized in Table 9. Mean particle size below junctions ranges from 1/3 to 2/3 and averages 1/2 of the sum of mean sizes in the two tributaries. In 40 percent of the cases, mean size below the junction is greater than in either tributary, and in about 25 percent of the cases, it is less than in either tributary.

TABLE 9. CHANGES IN MEAN PARTICLE SIZE (B-AXIS) AT STREAM JUNCTIONS

The value k is a coefficient in the expression $a = k\ (b + c)$, where b and c are tributaries, and a is the main stream below the junction.

STREAM AND STATION NUMBER	MEAN SIZE a	b	c	k
Pecos River				
1	115	123	136	0.43
2	112	117	97	.52
3	227	204	217	.54
5	160	158	146	.53
9	165	110	122	.71
10	191	195	151	.55
12	193	161	106	.72
14	99	106	139	.40
61	143	263	180	.32
62	171	189	158	.49
Rio Santa Barbara				
56	190	165	128	0.48
59	108	77	101	.61
64	163	182	144	.50
Rio Santa Cruz				
15	96	107	79	0.51
53	98	112	160	.36
Grand mean: $\bar{k} = 0.51$; $\sigma = 0.11$				

Changes in lithologic composition at junctions are summarized in Table 10. In a majority of the cases, the value of k is close enough to 0.5 to suggest that lithologic composition of bed material below junctions is roughly equal to the average of compositions in the two tributaries. There are a few cases, however, especially station 1 on the Pecos River, for which there is no apparent relationship between lithologic compositions above and below junctions.

Although the relations are not very clear, data for mean particle size and lithologic composition presented in both Tables 9 and 10 seem to indicate that in many cases bed material from tributaries is thoroughly mixed together only a short distance below the stream junction. Cases in which the mean size or lithologic composition of bed material for a station below a junction differs appreciably from the average for the tributaries are probably the result of selective sorting and varying degrees of mobility of bed material (that is, some is too coarse to move).

CHANGES ALONG SPECIFIC CHANNEL SEGMENTS

The detailed nature of interrelations which characterize downstream changes in bed material can be shown conveniently by considering a continuous segment of channel rather than including all stations located on the various tributaries of the drainage basin. Figures 11, 12, and 13 are examples of this kind, in which several aspects of particle size, lithologic composition, and channel slope are plotted against stream length. It should be noted that bedrock lithologies traversed by the stream are shown graphically in each figure; this facilitates comparisons of lithologic composition of stream-bed material with bedrock sources. The principal features of each diagram are described briefly below.

Rio Santa Barbara

The various graphs in Figure II refer to a channel segment which includes the West Fork and main stem of the Rio Santa Barbara. As shown in the diagram, and also on the map of Plate 1, the stream successively crosses quartzite, sandstone (with thin limestone), another belt of quartzite, and finally limestone[4] (with some sandstone) near the downstream end of the segment considered here. It also should be noted in Plate 1 that the West Fork follows or is near the trace of a major fault.

4. This limestone is part of the same formation as the sandstone and does not appear as a separate unit in Plate 1.

TABLE 10. CHANGES IN LITHOLOGIC COMPOSITION (PERCENTAGES OF MAJOR CONSTITUENTS ONLY) OF BED MATERIAL AT STREAM JUNCTIONS

The value k is an exponent in the expression $a = k\ (b + c)$, where b and c are tributaries, and a is the stream below the junction.

	GRANITE a	b	c	k	AMPHIBOLITE a	b	c	k	QUARTZITE a	b	c	k	SANDSTONE a	b	c	k
Pecos River																
1	49	14	14	1.75	20	11	14	0.80	14	11	12	0.61	12	56	56	0.11
2	20	0	32	.63					10	10	14	.42	62	86	50	.46
3													98	98	94	.51
5									34	28	12	.85	58	64	84	.39
9	36	34	40	.49	36	22	50	.50	14	18	2	.70	14	26	8	.41
10									36	38	24	.50	54	44	68	.48
12	24	36	12	.50					48	12	60	.67	16	51	22	.22
61									28	15	16	.90	48	67	72	.35
62													92	90	98	.49
Rio Santa Barbara																
56									20	0	24	0.84	80	100	74	0.46
59									84	86	85	.49	16	14	15	.55
64													98	100	100	.49
Rio Santa Cruz																
15	18	16	20	.50					74	80	72	0.49				
63	86	82	94	.49					6	12	2	.43				

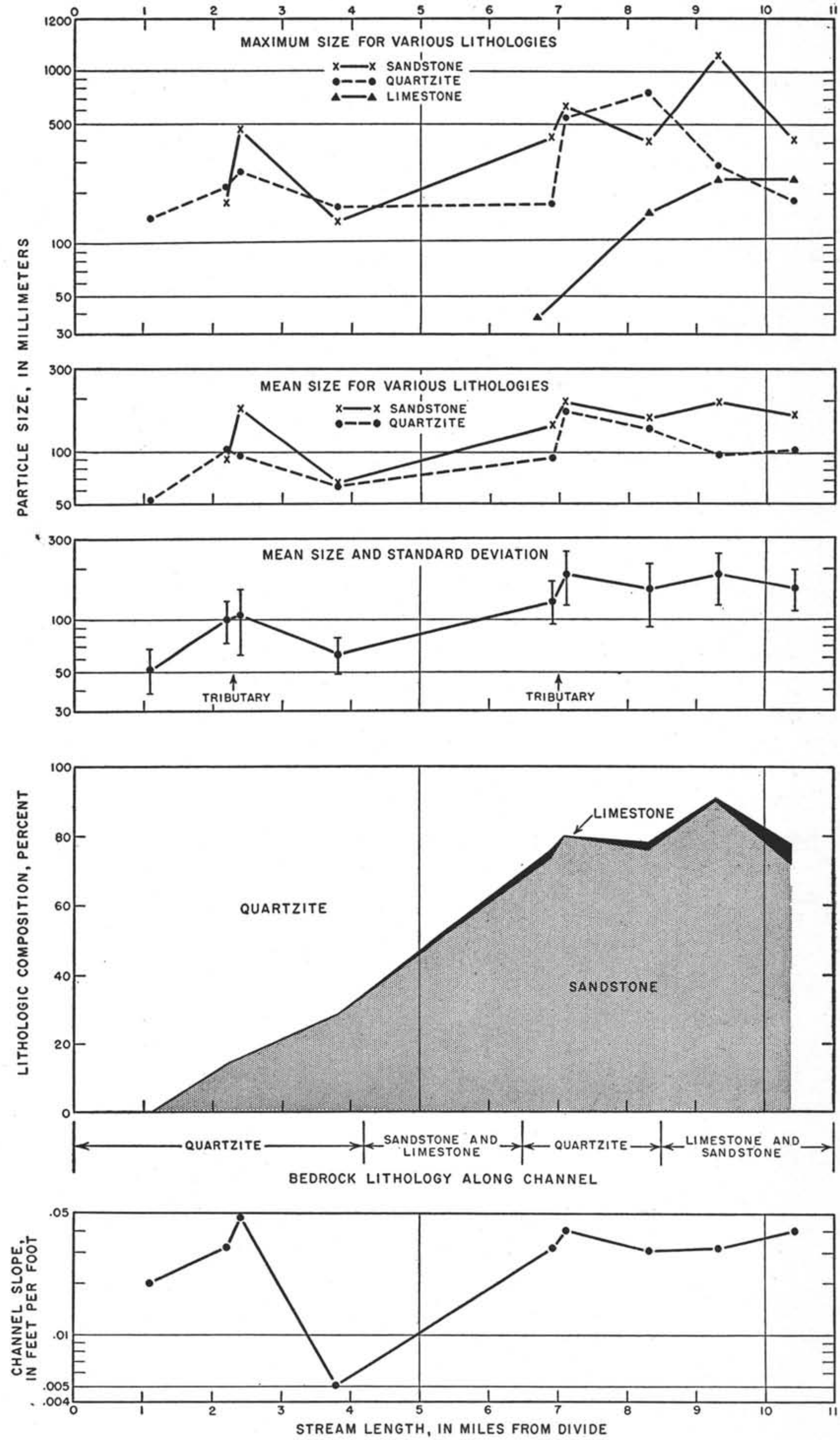


Figure 11

CHANGES IN CHANNEL SLOPE, AND MEAN SIZE AND LITHOLOGIC COMPOSITION OF BED MATERIAL, ALONG THE RIO SANTA BARBARA

Stations plotted in order downstream from the divide are: 1′, 59C, 59A, 57, 56C, 56A, 65, 71, and 49.

The most striking feature of Figure 11 is that stations 1' and 57 (at 1.1 and 3.8 miles, respectively) are characterized by much lower channel slope and smaller mean particle size of bed material than the other stations. Both these stations are located in mountain meadows produced by glaciation. The small particle size is related in part to fracture of rock in the fault zone, but it probably also reflects an effect of the gentle gradient on the carrying capacity of the stream. Except for these two stations, slope and mean particle size follow fairly consistent patterns; slope decreases slightly downstream, and mean particle size actually increases.

Quartzite and sandstone are the dominant constituents of the bed material. The presence of a particular lithology upstream from its outcrop along the course of the channel (for example, sandstone beginning at 1.2 miles) is due to tributary effects. In general composition, the bed material is related fairly closely to the lithology of bedrock exposed along the channel. A notable exception is the segment where limestone exposed along the stream contributes only a few percent of the bed material.

Changes in mean size and lithologic composition of bed material, as well as changes in slope, are greater at stream junctions than at lithologic contacts.

Quartzite particles decrease in size downstream from the limit of outcrop (at 8.5 miles), but a compensating increase in the size of limestone and sandstone particles causes values of mean size to remain essentially constant.

Rio Santa Cruz

The graphs of Figure 12 refer to a stream segment which includes the entire length of the Rio Medio and one additional station on the Rio Santa Cruz, just below the junction of the Rio Medio and the Rio Frijoles.

The uppermost 3.4 miles of the stream course traverses quartzite; the remainder is on granite except for a narrow belt of thin unconsolidated deposits just above the junction of the Rio Medio and the Rio Frijoles (pl. 2). It should be added that on the headwaters divide there is a very narrow strip of sandstone not shown in Figure 12 or Plate 2.

Channel slope decreases rapidly downstream from the divide to station 15B (at 3.9 miles), with a much smaller rate of decrease below this point. An exception to the general trend is the very gentle gradient at station 69 (8.4 miles). This station is in a deep canyon, and there is no obvious explanation for the small value of slope.

Mean particle size of bed material remains remarkably constant except at station 70 (11.5 miles), where there is a large increase apparently due to the large size of blocks coming off the canyon wall. It should be noted that the low slope at station 69 is not associated with small particle size. The nearly constant value of mean size results from the fact that a downstream increase in size of granite particles compensates for the decrease in size of quartzite particles.

Lithologic composition of bed material is rather closely related to bedrock exposed along the stream channel. The percentage of quartzite, for example, decreases rapidly downstream from its limit of outcrop (at 3.4 miles) to about 8.4 miles, below which point there is a smaller rate of decrease. The items in the category labelled "other" deserve additional explanation. Amphibolite is locally associated with the granite, whereas the phyllite and schist are minor constituents of the quartzite. The small amount of sandstone in the bed material evidently was derived from the headwaters divide area; much of the sandstone present probably was moved a considerable distance downstream during glaciation.

Pecos River

The 33-mile segment of channel referred to in the diagrams of Figure 13 includes the entire length of the Rito de las Chimoyosos and the main stem of the Pecos River below its junction with the Chimoyosos.

As can be seen in Figure 13 and Plate 4, bedrock exposed along this channel segment includes a greater variety of lithologic types than either the Rio Santa Barbara or the Rio Santa Cruz. Some of the boundaries of lithologic units shown in Figure 13 are necessarily generalized. Actually, in this case, the bedrock types traversed by the channel provide a rather incomplete picture of the sources from which streambed material may be derived. This is true because throughout most of the distance considered here the Pecos flows in a deep canyon which has been cut through the Paleozoic sediments and into the underlying Precambrian quartzite, granite, and amphibolite (pl. 4). The result is that debris of several different lithologies is shed from the canyon walls and into the stream channel. Also noteworthy (pl. 4) is the distribution of Mississippian limestone; this unit is a separate formation distinctly different from the thin limestones interbedded with Pennsylvanian sandstones.

The most striking features on the graph of channel slope are the marked changes which occur at stream junctions, especially upstream from mile 14. By comparison, slope changes downstream from mile 14 are relatively minor. The abrupt change in gradient between miles 12 and 14 occurs at the lower end of a very deep canyon section and roughly corresponds to the first of several amphibolite zones associated with the granite.

Except for minor changes at tributary junctions, mean particle size remains roughly constant downstream to about mile 12. Between miles 12 and 14, mean size decreases from about 180 to roughly 110 millimeters. As noted above, there is also an abrupt change in slope in this same reach. Downstream from mile 14, mean particle size remains essentially constant except at mile 23 (station 8), where a small ledge contributes exceptionally large fragments of amphibolite and granite. Another important fact is that the large slope changes at stream junctions (above mile 14) are not accompanied by major changes in mean particle size.

All lithologies locally yield large particles to the stream, and no one lithology consistently contributes larger particles than the others. In general, however, the largest particles derived from thin-bedded Pennsylvanian limestones (above mile 18) tend to be smaller than those derived from other lithologies.

Although the diagrams of downstream changes in lithologic composition of bed material are rather complicated, several interesting relations can be seen. The effect of local bedrock lithology is apparent all along this segment of channel, but is especially evident below mile 18, where the percentage of limestone starts to increase. As would be expected, the percentage of rock of any particular lithologic type decreases progressively downstream from its limit of outcrop, owing to dilution by contributions from the local rock. This relation is especially well shown in the case of quartzite. Schist and phyl-

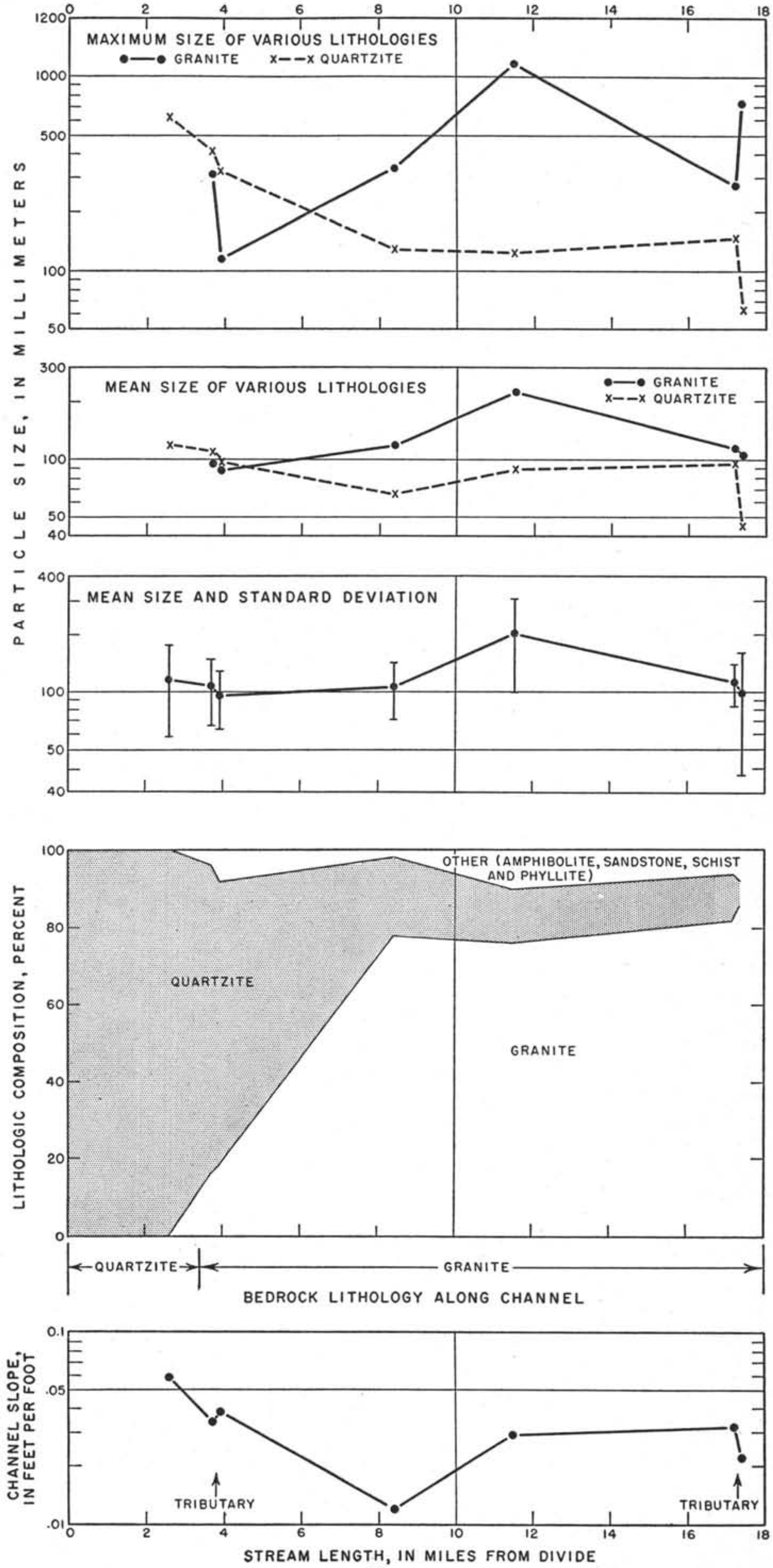


Figure 12

CHANGES IN CHANNEL SLOPE, AND MEAN SIZE AND LITHOLOGIC COMPOSITION OF BED MATERIAL, ALONG THE RIO SANTA CRUZ

Stations plotted in order downstream from the divide are: 16A, 15B, 15A, 69, 70, 53B, and 53A.

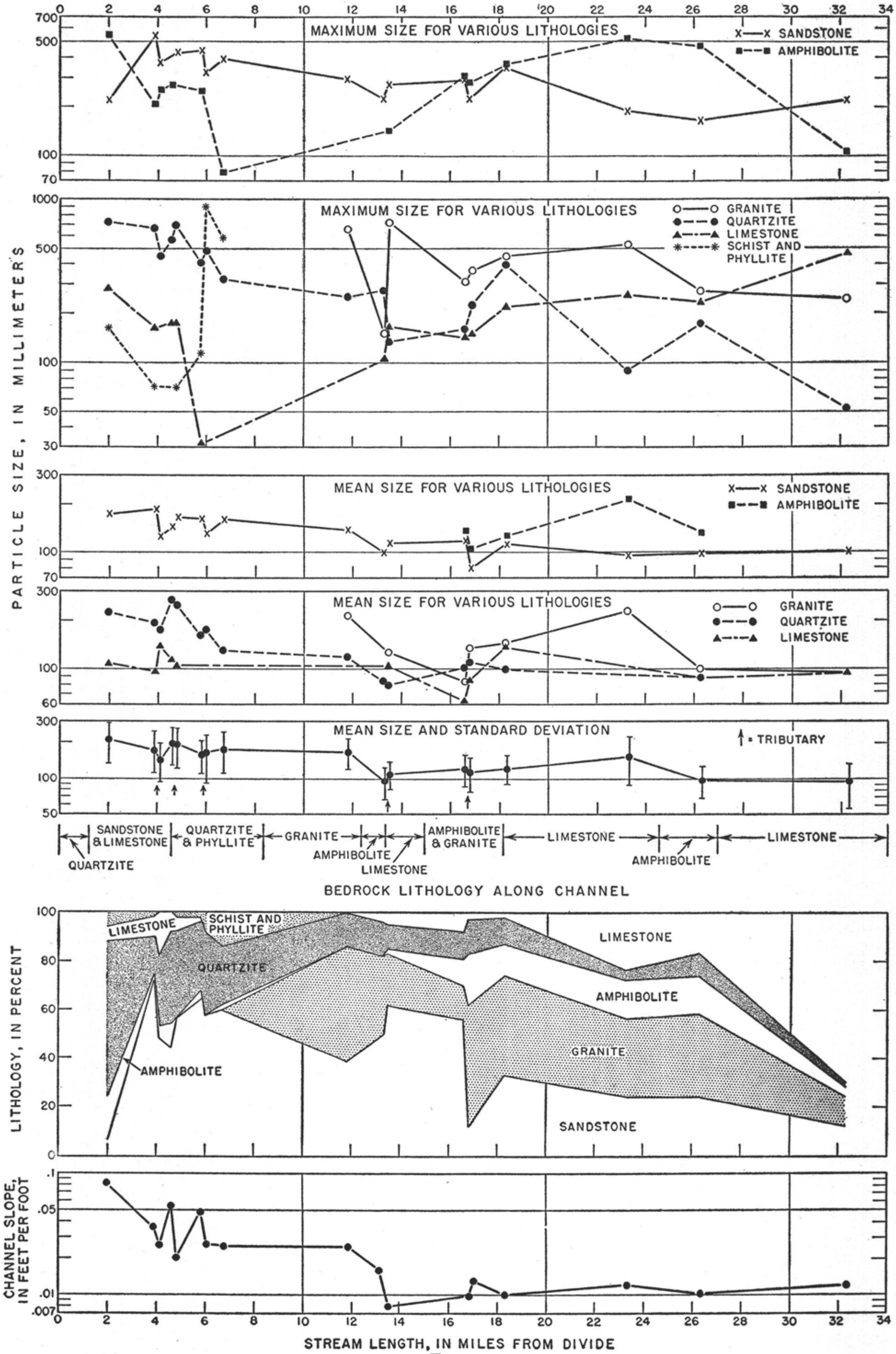


Figure 13

CHANGES IN CHANNEL SLOPE, AND MEAN SIZE AND LITHOLOGIC COMPOSITION OF BED MATERIAL, ALONG THE PECOS RIVER

Stations plotted in order downstream from the divide are: 60A, 61C, 61A, 10B, 10A, 5B, 5A, 11, 51, 2C, **2A, 1B,** 1A, 40, 8, 52, and 7.

lite particles evidently are broken up by abrasion and disappear completely within a distance of 2 miles below the outcrop.

In summary, the data on specific channel segments presented in Figures 11, 12, and 13 indicate several different kinds of relations between the variables considered. Mean particle size of bed material may increase, decrease, or remain approximately constant in the downstream direction. Major deviations from these trends are in most cases related to easily recognized geologic factors. Channel slope remains approximately constant in one case and decreases slightly downstream in the other two. The most abrupt changes in slope occur in most cases at tributary junctions. In general, the streambed material at any point includes a fraction derived from nearby bedrock sources and a fraction transported from upstream. It would be reasonable to suppose that modern processes are involved, and that the relative abundance of these two fractions depends entirely on differences between lithologies in the amounts of rock debris delivered to the stream, as well as on differential resistance to abrasion during transport. However, as will be discussed in detail later, modern processes probably have had less effect on the lithologic composition of bed material than have the events of recent geologic history.

CHANGES AT THE MOUNTAIN FRONT

Remarkable changes in channel properties occur at the western margin of the range, where mountain streams carrying perennial flow encounter the unconsolidated Santa Fe formation and are transformed into ephemeral arroyos. Within a few hundred feet, channel width increases severalfold, and bed material changes in size from coarse gravel to dominantly sand with scattered pebbles and cobbles. A short distance south of Pecos, New Mexico, bed material in the channel of the Pecos River undergoes a similar but much more gradual change from cobbles and boulders to dominantly sand with gravel concentrated on bars. The contrast between these two kinds of channels is shown in Plate 10. The Pecos downstream from the mountains encounters consolidated rocks and does not increase greatly in width. Furthermore, it maintains appreciable flow except during very dry years, when nearly all the water is removed for irrigation.

Sediment derived from the highly erodible Santa Fe formation accounts in part for the abrupt changes in size of streambed material which occur at the western margin of the range. Several different sandstones contribute large amounts of sand to the bed material of the Pecos River downstream from the mountains. In both cases, the fact that local sources can, in a very short distance, dilute the gravel fraction to such an extent implies that relatively little gravel is transported by the streams.

DILUTION

As was noted in the descriptions of specific channel segments (fig. 11, 12, and 13), the percentage of a particular lithology in the bed material decreases rather rapidly downstream from the limit of outcrop of that lithology. Dilution by contributions from local sources probably is the principal factor involved, but loss due to abrasion doubtless has an effect also, especially for nonresistant lithologies.

In the area considered here, quartzite is the only lithology with outcrops confined to the headwaters portions of several drainage basins. Not only is the outcrop distribution appropriate for dilution studies, but quartzite is also an ideal rock type for such purposes because of its extreme resistance to abrasion. The decrease in percentage of quartzite in bed material downstream from the limit of quartzite outcrop was studied in four basins, the Rio Santa Cruz, the Rio Mora and the Rito de las Chimoyosos (both tributaries of the Pecos River), and the main stem of the Pecos River. The data are plotted in Figure 14.

It should be pointed out that data given in Figure 14 for the main stem of the Pecos River are not exactly comparable with those for the other streams, because there is no zone where quartzite is the only lithologic type contributing to the bed material. Rocks of other lithologies are delivered to the bed material, both directly from the valley slopes and also from tributaries, with the result that quartzite comprises no more than 34 percent of the bed material even in belts of quartzite outcrop along the Pecos. Another difference is that some of the quartzite in the bed material at various points along the main stem of the Pecos may have come from one or both of the Pecos River tributaries considered separately in Figure 14. Also, the data for the Rio Mora are not completely satisfactory, because there is a small contribution of quartzite from the Rio Valdez. The data for the Rio Santa Cruz and the Rito de las Chimoyosos have no such disadvantages, but those for the Rio Santa Cruz probably are more easily interpreted because of the fact that only one other lithology is exposed downstream from the limit of quartzite outcrop.

Despite certain inequalities in the data, it is evident in Figure 14 that the downstream decrease in percentage of quartzite can best be described by an exponential function of the form $y = 100e^{-ax}$, where y is percentage and x is distance from the outcrop. The value of a, which is the coefficient of decrease, ranges between 0.2 and 0.4 for most of the examples plotted. According to these two exponential functions, the percentage decreases from 100 to 50 percent in 1.7 to 3.5 miles from the outcrop, and to 25 percent in 3.5 to 7 miles. Although this rate of decrease seems very great, there are no published data from comparable streams for comparison.

It is reasonable to suppose that the rapid decrease in percentage of quartzite is due largely to dilution. However, as will be discussed in the next section, size of the quartzite particles decreases somewhat in the downstream direction. The extent to which this trend is the result of particle wear, as opposed to selective sorting, is unknown. Actually, it is doubtful that changes in size, either by abrasion or sorting, are of sufficient magnitude to produce the effects observed. Nonetheless, uncertainty about these factors makes it impossible to evaluate the importance of dilution as a process controlling downstream changes in lithologic composition of bed material.

PARTICLE WEAR

Decrease in size of particles of a particular lithology downstream from the limit of outcrop may be due to particle wear or to selective sorting, and the two effects are not separable in natural stream gravels. Experimental results of Krumbein

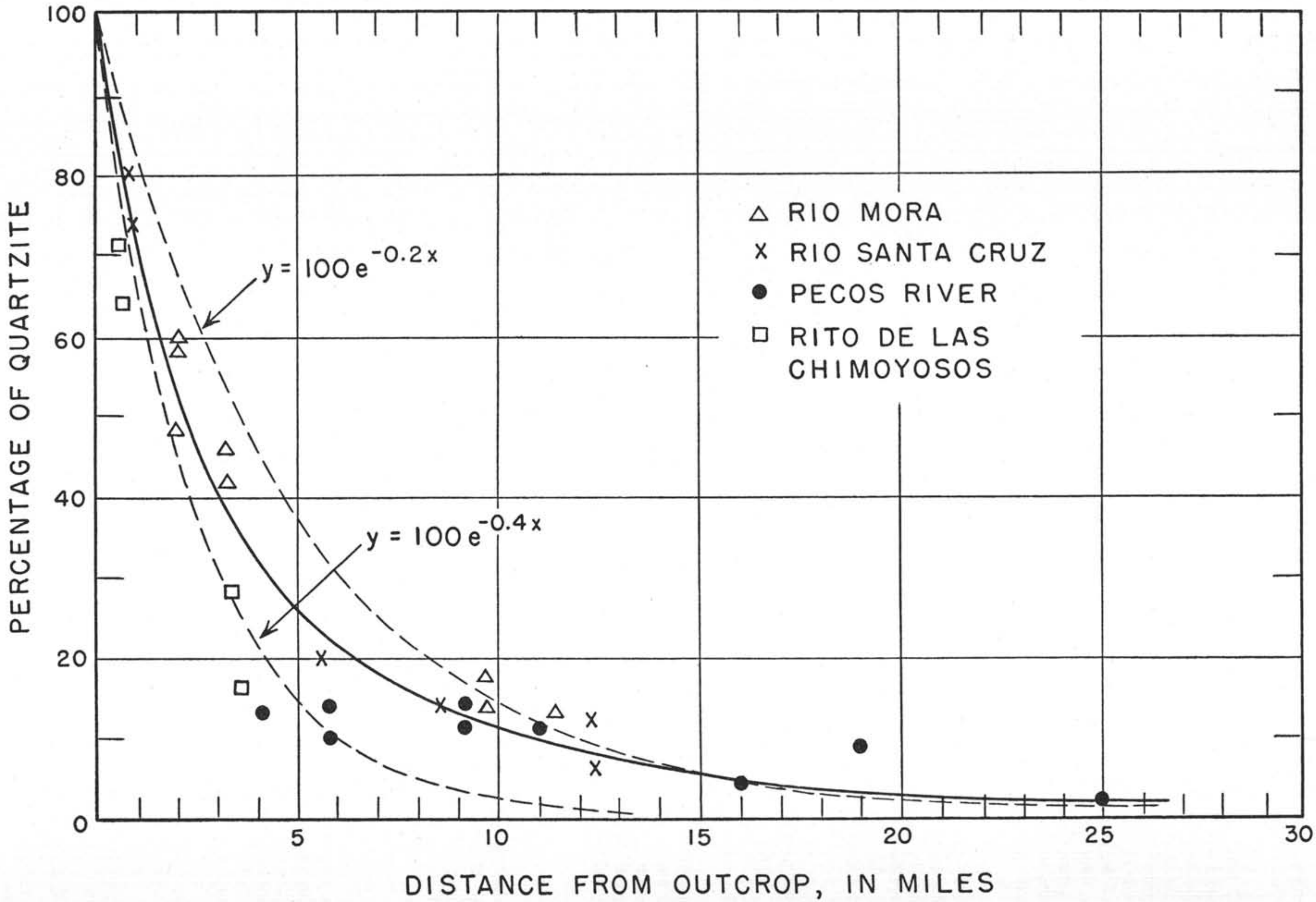


Figure 14

DECREASE IN PERCENTAGE OF QUARTZITE IN BED MATERIAL DOWNSTREAM FROM THE LIMIT OF QUARTZITE OUTCROP IN VARIOUS DRAINAGE BASINS

Stations plotted are as follows: Rito de las Chimoyosos, 60B, 60A, 61C, 61A; Pecos River, 51, 2C, 2A, 1B, 1A, 40, 8, 52, 7; Rio Mora, 12C, 12A, 14B, 14A, 9B, 9A, 1C; Rio Santa Cruz, 15B, 15A, 69, 70, 53B, 53A. The solid line defines a general trend through most of the points, and the dashed lines are graphs of two different exponential functions.

(1941b) and others indicate that particle-size decrease due to abrasion proceeds very slowly as a function of transport distance.

Data which may indicate particle wear of granite and quartzite in the Pecos River are plotted in Figure 15. The two stations farthest downstream (Sands and Dalia) are a considerable distance from the mountains and are not perfectly comparable with the other stations plotted, in that the channel is mostly covered with sand, with gravel restricted to bars. Several of the mountain stations are in locations which are affected to some degree by tributary contributions. Also, the graph for granite omits one station along this segment of the Pecos River, where a small granite-amphibolite outcrop contributes large granite blocks to one edge of the stream; almost certainly, most of these blocks are too big to be transported.

Despite the inequalities noted above and the additional factor of varying degrees of dispersion around the mean values plotted, the data in Figure 15 seem to indicate clearly that particle size decreases as negative power functions of distance transported. Furthermore, the rate of decrease of both mean and maximum size is similar for both granite and quartzite, suggesting roughly equal degrees of resistance to abrasion.

Although data for the various tributaries with quartzite outcrops are not plotted here, they suggest relations almost identical to those shown in Figure 15. Specific particle size at a given distance from the outcrop differs in the various streams, but the rate of decrease is approximately the same in all cases.

The conclusions on size decrease presented above should be regarded as tentative. More rigorous proof of the inferred relations requires larger samples, especially at stations far downstream on the Pecos. Plans to obtain additional data during the summer of 1957 were thwarted by exceptionally large floods.

SORTING

Values of the Trask sorting coefficients for bed material at the various stations are given in Tables 15-18. These data are summarized in Table 11. It should be noted in Table 11 that both the means and ranges in values are similar for all four drainage basins. In many cases, the extremes in sorting are related to obvious geologic causes. Along the Rio Santa Barbara, for example, the best sorted bed material occurs at stations in high mountain meadows where particle size is also small, and the poorest sorting occurs at a station located near the foot of a high quartzite cliff.

The amount of published information on sorting of stream gravels is remarkably small, and none of it refers to high mountain streams comparable to those considered here. Emery (1955) summarized published data on sorting of 35 samples with a median size of 19 mm and a size range from 10.4 to 355 mm. The Trask sorting coefficient for these samples ranged from 1.34 to 5.49, with a median of 3.18. Brush (in press) measured bed material of streams in the Appalachians of central Pennsylvania; his 119 stations were located on riffles, where particle size ranged from 5 to approximately 700 mm, though the median size in most cases was between 50 and l00 mm. He obtained values of the Trask sorting coefficient which range from 1.29 to 3.32, with a mean of 1.57.

TABLE 11. SUMMARY OF TRASK SORTING COEFFICIENTS FOR BED MATERIAL IN SANGRE DE CRISTO STREAMS

Particle size ranges from 12 to 1230 mm.

STREAM	TRASK SORTING COEFFICIENT RANGE	MEAN
Rio Santa Barbara (18 stations)	1.32–1.77	1.46
Rio Santa Cruz (13 stations)	1.46–1.70	1.59
Pojoaque River (20 stations)	1.32–1.98	1.59
Pecos River (53 stations)	1.29–2.05	1.60
All stations (104)	1.29–2.05	1.57

It is surprising that Sangre de Cristo streams, which have very coarse bed material, are better sorted than the finer bed material considered by Emery (1955) and measured by Brush (in press). Actually, if entire size distributions are considered, the Sangre de Cristo samples have the poorest sorting, though this relation is obscured by the Trask coefficient, which involves only the central half of the size distribution.

No consistent relation between size and sorting was found, though in many cases fine bed material is better sorted than coarse. Furthermore, sorting does not improve in the downstream direction. For example, see Figures 11, 12, and 13, in which dispersion (standard deviation) about the mean particle size at each station is plotted. Dispersion, which is a measure of sorting, varies from place to place along each channel segment but does not show a decreasing trend. Considering the fact that particle size and lithologic composition do not change systematically in the downstream direction, there is no reason to expect a consistent downstream improvement in sorting. Brush (in press) reached essentially the same conclusions for Appalachian streams.

PARTICLE SHAPE

The most commonly used measure of particle shape, Krumbein's (1941a) intercept sphericity,

$$\psi = (^{bc}/_{a^2})^{1/3}$$

requires that dimensions of the longest (a), intermediate (b), and shortest (c) axes of each particle be determined. Because of the considerable time element involved, it was possible to obtain sphericity data at only four stations along a 72-mile segment of the Pecos River. For these four stations, sphericity of particles with particular lithologies (granite, amphibolite, and quartzite) rather than entire samples of bed material were considered. This procedure was dictated by the fact that the two stations farthest downstream (Sands and Dalia) are characterized by dominantly sandy bed material, and particles of gravel size could be obtained only from the bars. All four stations are downstream from bedrock exposures of quartzite, but the two upstream stations (51 and 52) are at sites where granite or amphibolite are exposed near the channel.

5. Defined as $S_o = \sqrt{Q_3 / Q_1}$ where Q_3 = diameter in mm, of which 75 percent of the sample is finer, and Q_3 = diameter in mm, of which 25 percent is finer.

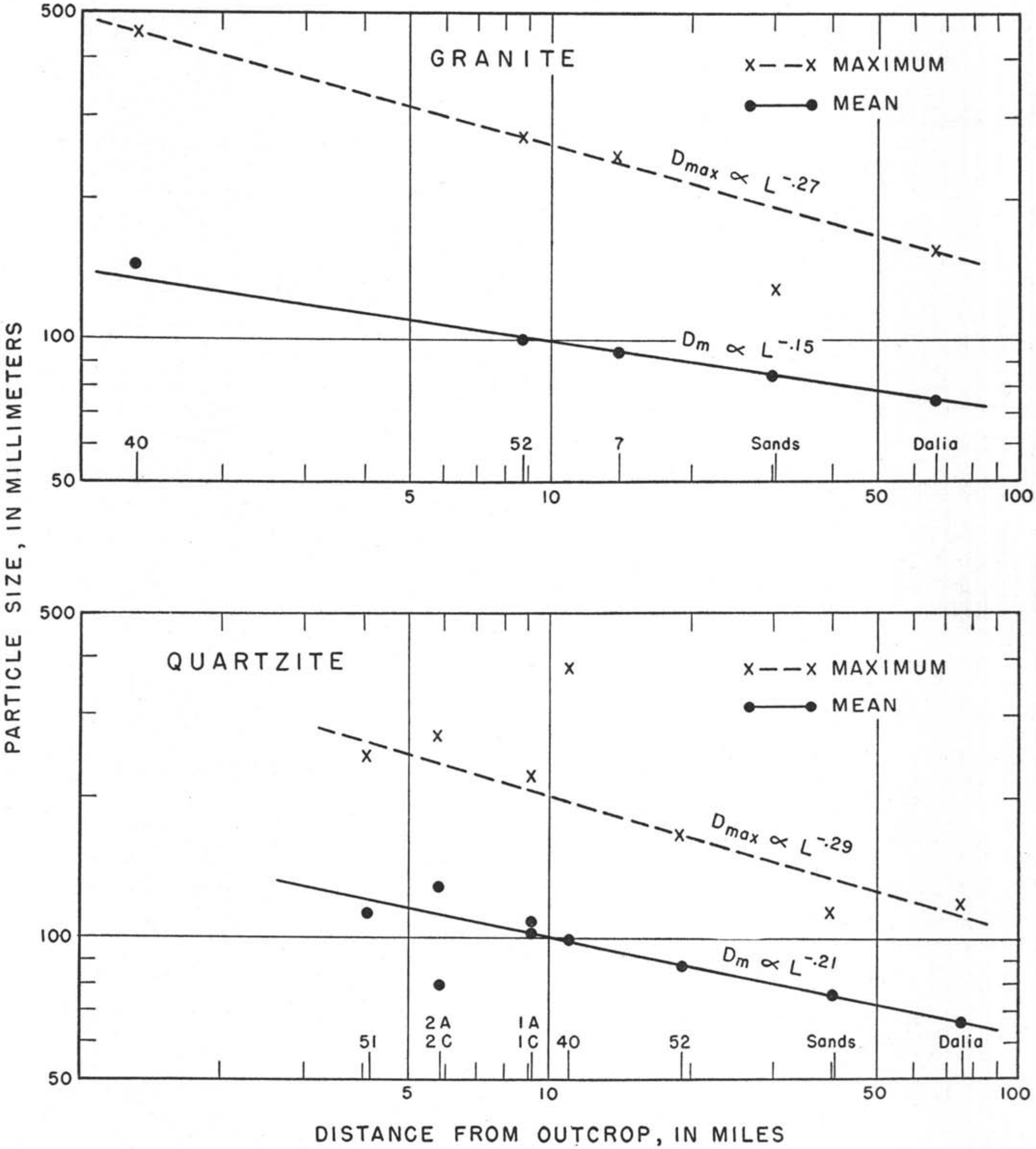


Figure 15

Decrease in mean and maximum particle size of quartzite and granite downstream from outcrops of these lithologies along the main stem of the Pecos River

Stations plotted are indicated along the base of each graph.

The qualitative character of downstream changes in shape can be ascertained from Plate it. Particle size obviously decreases, and roundness increases, for all three lithologies. The fact that a high degree of roundness is achieved within a relatively short distance is in accord with the observations of Kuenen (1956) and others.

Data obtained from sphericity measurements are summarized in Table 12. Disk shapes are considerably more common for all three lithologies than the other Zingg classes; however, quartzite at station 52 and at Sands affords exceptions to this general relation. Despite different degrees of bedrock and tributary influence, sphericity at each station is approximately the same for all three lithologies. Mean sphericity values for each of the three lithologies change slightly downstream, but the samples are not large enough for the small apparent changes to be significant at the 5-percent level.

The various lithologies considered separately or together show no well-defined correlation between particle size and sphericity, though in general the smaller particles tend to have somewhat greater values of sphericity.

It is generally supposed that sphericity increases downstream, especially in headwaters regions, owing to particle wear. Actually, the data bearing on this relationship for gravels are remarkably few in number (Pettijohn, 1957, p. 550-551). Brush's (in press) study of streams in the Valley and Ridge Province of Pennsylvania indicated no downstream change in sphericity of bed material on riffles. Results from the present investigation show no significant downstream change in sphericities of individual lithologies.

RELATION OF CHANNEL SLOPE TO PARTICLE SIZE

In recent years, several writers have discussed the possibility that channel slope is controlled to some degree by particle size of bed material. Shulits (1941) suggested that downstream wear of bedload constitutes an important aspect of any rational explanation of stream profiles; he indicated, however, that such a relation may not apply to steep mountain streams. Culling (1957, p. 263) is of the same opinion. Leopold and Wolman (1957) imply a relation between channel slope and particle size but do not present supporting data or a detailed discussion. Hack (1957) found good correlations between slope and median particle size for Appalachian streams, provided that the data were grouped according to bedrock lithology. For these cases, particle size decreased, increased, or remained constant as slope decreased downstream. Brush (in press) shows fairly close relationships between slope and particle size for some streams in the Appalachians of central Pennsylvania; however, there are several cases in which slope is definitely independent of particle size.

Along portions of the channel segments shown in Figures 11, 12, and 13, channel slope appears to change with variations in mean particle size of bed material. The trends of these examples indicate that slope may decrease or remain constant, accompanied by decreasing, increasing, or constant particle size. However, if all stations in each drainage basin are considered, as in Figure 16, it seems apparent that slope does not depend primarily on particle size. In contrast to the findings of Hack (1957), this is true even for streams on the same bedrock lithology. Nearly the entire basins of the Rio Santa Cruz and the Pojoaque River are underlain by granite. Also, roughly the upper halves of the graphs for the Rio Santa Barbara and the Pecos River refer to stations underlain by generally similar lithologies.

The conclusion to be drawn from the Sangre de Cristo data seems to be that channel slope and particle size are not caused one by the other, but rather that both are related in

TABLE 12. SUMMARY OF SPHERICITY MEASUREMENTS AT FOUR STATIONS ALONG THE PECOS RIVER

Each sample consisted of 100 particles; percentages of the various lithologies are indicated in parentheses.

	ZINGG CLASS (PERCENT)				SPHERICITY CLASS (PERCENT)						SPHERICITY		SIZE (B-AXIS)	
	I Disks	II Spherical	III Blades	IV Rods	0.40 –.49	0.50 –.59	0.60 –.69	0.70 –.79	0.80 –.89	0.90 –1.0	Mean	Std. Dev.	Mean	Std. Dev.
Station 51 (12 mi from divide)														
Granite (48%)	44	26	19	11	0	26	44	26	4	0	0.66	0.078	210	110
Quartzite (13%)	50	8	8	34	0	9	58	33	0	0	0.66	0.054	118	63
											0.66	0.071	190	107
Station 52 (27 mi from divide)														
Granite (33%)	36	17	17	30	0	17	40	40	3	0	0.68	0.105	96	55
Amphibolite (16%)	36	21	14	29	7	15	42	29	0	7	0.65	0.112	123	103
Quartzite (9%)	22	22	22	34	0	22	45	22	11	0	0.67	0.095	87	44
											0.67	0.094	102	70
*Pecos River at Sands, N. Mex.** (48 mi from divide)														
Granite (12%)	75	17	0	8	0	9	33	58	0	0	0.70	0.055	85	21
Amphibolite (7%)	57	0	29	14	0	29	42	29	0	0	0.64	0.092	73	34
Quartzite (8%)	12	63	0	25	0	0	37	50	13	0	0.73	0.056	75	32
											0.69	0.072	78	28
*Pecos River at Dalia, N. Mex.** (84 mi from divide)														
Granite (6%)	49	17	17	17	0	33	34	33	0	0	0.64	0.069	75	58
Amphibolite (20%)	50	10	25	15	15	35	35	5	5	5	0.61	0.105	69	29
Quartzite (14%)	65	7	14	14	0	14	43	43	0	0	0.67	0.075	66	32
											0.64	0.092	69	34

*** Samples collected on bars.**

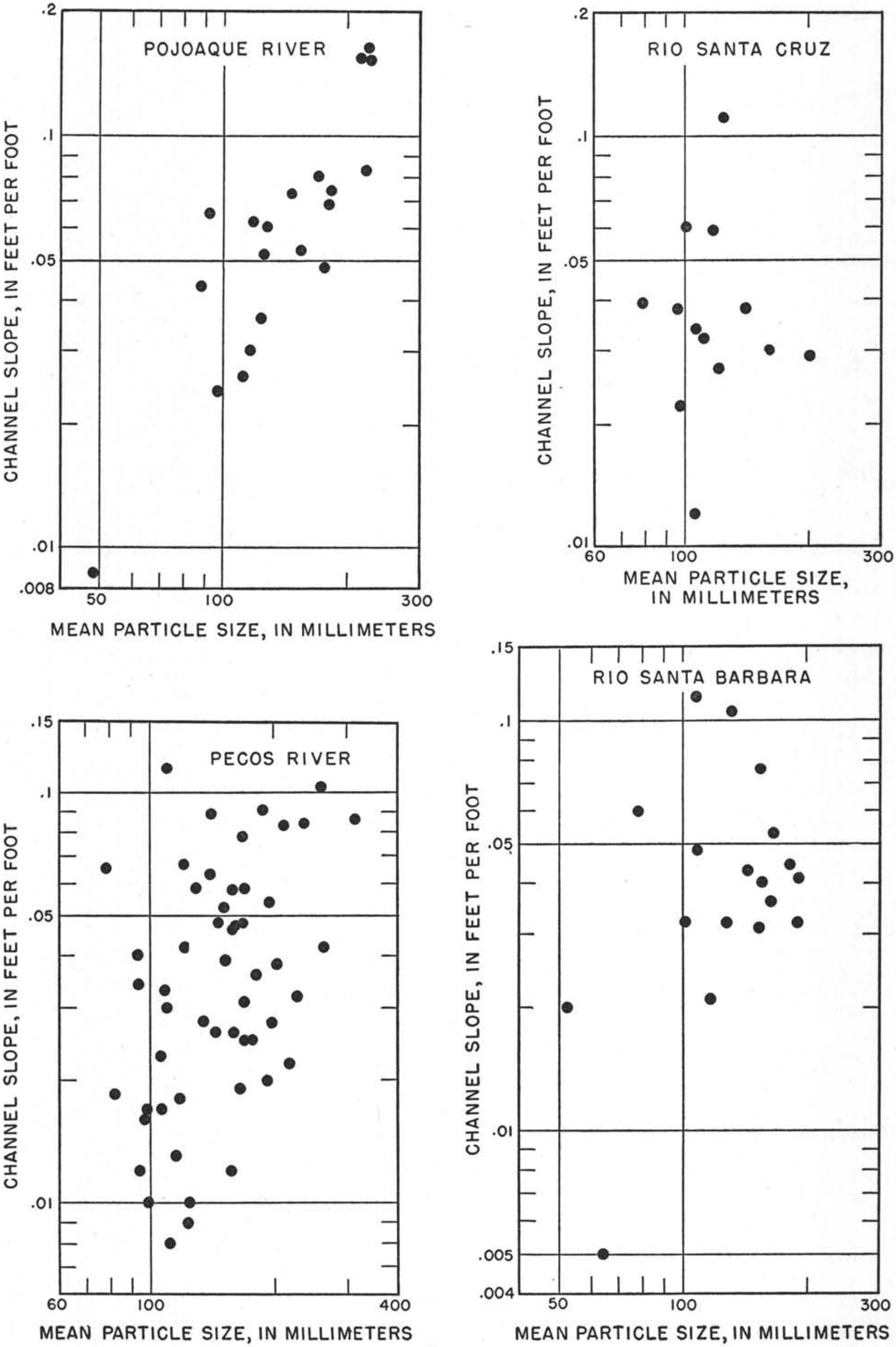


Figure 16

Relation of channel slope to mean particle size of bed material for all stations in each of the four drainage basins studied varying degrees to other factors, especially areal distribution of lithologic types and relief-producing events during the geologic past.

DOWNSTREAM CHANGES IN CHANNEL ROUGHNESS

Wolman (1955) expressed downstream change of channel roughness as a function of discharge,

$$n' = r Q^y,$$

in which n' is the roughness factor in a slightly modified version of the Manning equation.[6]

$$v = \frac{1.5\, d^{2/3}\, s^{1/2}}{n'}. \qquad (11)$$

Substituting the function of each variable,

$$kQ^m = \frac{1.5\, (cQ)^{2/3f}\, (tQ)^{1/2}}{r\, Q^y}.$$

It follows that

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

and

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

Because values of exponents *f*, *z*, and *m* have been determined for Sangre de Cristo streams, *y* can be computed. The results are as follows:

	y
Rio Santa Barbara	- 0.14
Pecos River	- 0.29
Rio Santa Cruz	- 0.58
Pojoaque River	- 0.91

By way of comparison, Wolman (1955) obtained values of z ranging from - 0.28 to - 0.32 for Brandywine Creek. Leopold and Miller (1956) concluded that z = - 0.3 for arroyos in New Mexico.

There appears to be no really satisfactory way to determine the degree of relationship between downstream rates of change of roughness and size of bed material. Ranges in size of bed material are roughly the same for all streams considered here. In the case of the Rio Santa Barbara (fig. 11), mean particle size increases somewhat irregularly downstream and is associated with a small rate of change of roughness. For the Pecos River (fig. 13), where particle size decreases in steps but otherwise remains fairly constant through long distances, the rate of change of roughness is somewhat greater. Mean size of bed material in the Rio Santa Cruz (fig. 12) decreases, though not as much as for the Pojoaque River (not plotted). For the latter two streams, the decrease of mean particle size cannot be sufficient to account for the extremely high rate of change of roughness required by the computations based on hydraulic considerations. As was discussed previously, the hydraulic argument that velocity in these two basins increases downstream at an exceptionally high rate may not be entirely valid. The fact remains, however, that the channel dimensions are anomalously small. Apparently they cannot be explained in terms of a special relationship of channel roughness to particle size of bed material.

6. In the Manning equation, hydraulic radius replaces depth, and slope refers to slope of the energy gradient rather than the water surface.

PROBLEM OF COMPETENCE

During the last 15 years, the writer has had many opportunities to observe Sangre de Cristo streams under conditions ranging from essentially no flow to slightly greater than bankfull stage. Only rarely is a particle seen to move during ordinary flows. Even at high-water stages, when the water is murky with suspended load, movement of bed material is not readily detected by "feel" during wading operations nor indicated by the sound of boulders clanking together. These observations, together with other kinds of field evidence and inferences based on velocities required to move particles of various sizes, lead to the conclusion that a large fraction of the bed material in high mountain streams is immobile at discharges up to the bankfull stage.

Many experimental investigations of competence have been made, but unfortunately, very few of these have involved gravels. Procedures used have varied considerably, especially with regard to the definition of competent velocity; namely, the critical velocity required to move a particle of some specific size. Some workers have referred to mean velocity of the current, whereas others have attempted to measure or compute "bed velocity." Unfortunately, there is no simple basis for conversion between these two measures of velocity.

The most recent review of sediment movement in relation to current velocity is that of Menard (195o). Using data derived from his own experiments and also from the literature, he concluded that

$$D = 1.0\, v^2, \qquad (12)$$

where particle size (D) is expressed in millimeters and mean velocity (*v*) in feet per second. Menard's values of competent velocity for normally distributed sediments are slightly smaller than those of Hjulstrom (1939), who used sediment of uniform size. The conclusion that stream competence varies as the square of the velocity has had several other supporters; however, larger values of the exponent have also been reported.

The relation expressed in equation (12) will be used here in an attempt to estimate the fraction of movable bed material in Sangre de Cristo streams. It should be clear at the outset that the results obtained must be considered tentative, because the degree of applicability of this relation is unknown. The fact that the experiments on which it is based involve smaller materials and slopes less steep than those of most mountain streams provides grounds for considerable uncertainty.

Comparison of estimated bankfull velocity and measured size distribution of bed material at each station with the relation for competent velocity [equation (12)] yields an estimate of the fraction of bed material which can be moved by the stream at bankfull stage. Data from such a comparison at 1o6 stations are summarized in Table 13.

Part A of Table 13 shows that all the bed material at two-thirds of the stations is too coarse for transport under the assumed competency conditions. There are only 7 stations at which more than half the bed material is movable. The 5 stations at which less than 10 percent of the bed material is immobile are near the mountain front, in the basins of the Rio Santa Cruz and the Pojoaque River. It will be recalled

TABLE 13. NUMBER OF STATIONS AT WHICH GIVEN PERCENTAGE OF BED MATERIAL IS TOO COARSE FOR TRANSPORT

Particle diameter and mean velocity are assumed to be related according to the equation $D = 1.0\ v^2$.

A. Estimated bankfull velocities

	PERCENTAGE							
	0	1–9	10–29	30–49	50–69	70–89	90–99	100
Pojoaque River	1	2	0	1	1	3	3	9
Rio Santa Barbara	0	0	0	0	0	1	2	15
Rio Santa Cruz	0	2	1	1	1	2	3	3
Pecos River	0	0	1	0	0	3	12	39
Totals	1	4	2	2	2	9	20	66

B. Estimated bankfull velocities × 2

	PERCENTAGE							
	0	1–9	10–29	30–49	50–69	70–89	90–99	100
Pojoaque River	5	3	2	1	1	2	0	6
Rio Santa Barbara	0	0	1	1	4	3	4	5
Rio Santa Cruz	3	3	2	2	1	1	1	0
Pecos River	1	2	3	8	6	11	13	11
Totals	9	8	8	12	12	17	18	22

that the values of estimated velocities for both these streams were anomalously high (fig. 9).

Kalinske (1942, p. 641) suggests that maximum instantaneous velocities attained in streams are rarely more than twice the mean-point velocity. The data presented in part B of Table 13 are based on the same competency conditions as in part A, with the additional assumption that maximum velocities are twice the estimated bankfull velocities. Under these conditions, more than half of the bed material is movable at only 36 of the 106 stations. All but 3 of the 17 stations at which less than Io percent of the bed material is immobile are in the basins of the Rio Santa Cruz and the Pojoaque River. The 22 stations where 100 percent of the bed material is too coarse for transport are in headwaters reaches, all less than 8 miles, and most less than 4 miles, from the divide.

Another approach is to consider the velocities required to transport particles of a particular lithology which have moved downstream from specific outcrop areas. Quartzite is the ideal lithology for this purpose because, as discussed previously, it occurs in the bed material downstream from restricted outcrops in the basins of the Rio Santa Cruz, the Rio Mora, the Rito de las Chimoyosos, and the main stem of the Pecos River. Assumptions regarding competency and instantaneous velocity were the same as in Table 13. The data, plotted in Figure 17, indicate that the estimated bankfull velocity at 17 of the 25 stations considered is not great enough to transport any of the quartzite particles which have actually moved several miles downstream. Furthermore, if the estimated bankfull velocity is doubled in accordance with the assumption about variations in instantaneous velocity, there are only 6 stations at which l00 percent of the quartzite particles present could be transported.

Several other features of the graphs in Figure 17 should be noted. The percentage of quartzite fine enough for transport increases downstream in the four basins considered. An exception is station 1, at the mouth of the Rio Mora, where channel width is exceedingly large (which in turn causes the computed velocity to be low). Stations where little or none of the quartzite is movable by doubling the estimated bankfull velocity are, with a few exceptions, fairly close to headwater divides.

The results presented in Table 13 and Figure 17 can be interpreted in several ways. They may indicate that the experimental data and assumptions used in this method of evaluating competence are not applicable to high mountain streams. Alternatively, it could be argued that the computed estimates of bankfull velocity are too small. Yet another possibility is that the general conclusions reached are valid; namely, velocities attained at the bankfull stage are at many places inadequate to move bed material. If this is true, the observed cases in which transport downstream from specific outcrop areas has occurred may be either the result of even larger floods, which still occur occasionally, or the residual effect of some former period during which considerably larger discharges prevailed.

Some evidence that bed material may be moved during floods which exceed the bankfull stage was found along the Rio Santa Cruz and the Pojoaque River in the lower portions of their courses near the mountain front. Piles of fresh, bouldery debris extend over sizable areas adjacent to the channels and in places bury the basal parts of large trees. No cases of this kind were observed anywhere in the high mountains nor along the lower reaches of the Rio Santa Barbara and the Pecos River.

Competence can be defined also in terms of the shear on the streambed; specifically, by the expression for tractive force (τ),

$$\tau = \rho\, d\, s, \qquad (13)$$

where ρ is the specific weight of water, d is depth, and s is slope. Obviously, if slope decreases downstream faster than depth increases, tractive force decreases. Exponents in equations relating slope and depth to downstream increase in bankfull discharge of Sangre de Cristo streams are summarized as follows:

RIVER	SLOPE	BANKFULL DEPTH
Rio Santa Barbara	—0.36	0.49
Pecos River	—0.63	0.30
Rio Santa Cruz	—0.30	0.22
Pojoaque River	—0.48	0.09

These relations indicate that tractive force at the bankfull stage decreases appreciably downstream for the Pecos and Pojoaque Rivers, decreases slightly in the Rio Santa Cruz,

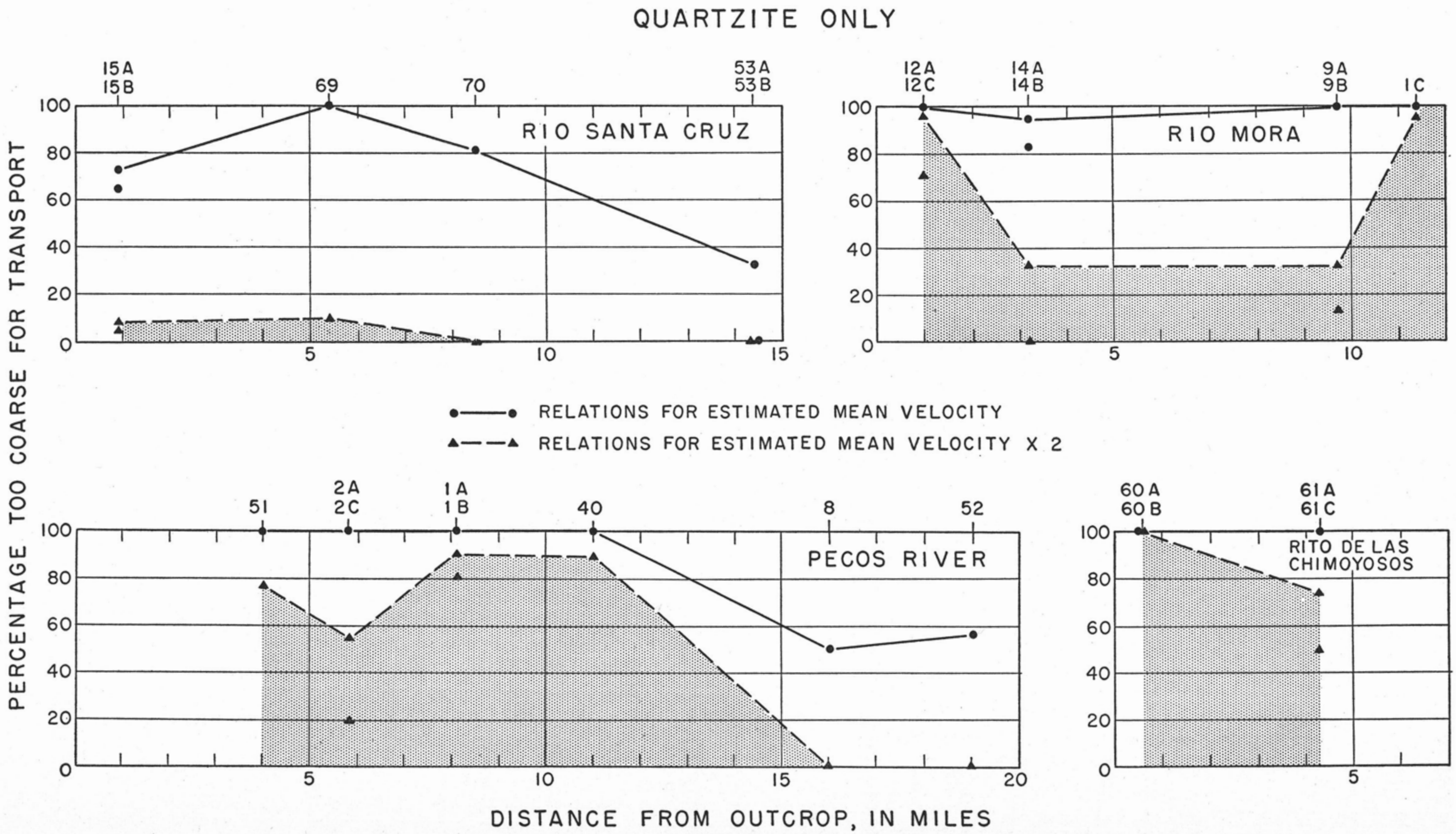


Figure 17

ESTIMATES OF COMPETENCE BASED ON THE ASSUMPTION THAT PARTICLE SIZE AND MEAN VELOCITY ARE RELATED ACCORDING TO THE EQUATION $D = 1.0\ v^2$

The data plotted refer only to quartzite particles which have moved downstream from the limit of outcrop. Stippled areas show percentages of quartzite particles too large for transport even if the estimated bankfull velocity were doubled. Where stations at tributaries give different results, the line is drawn arbitrarily through the higher value. Numbers of the stations referred to are plotted along the top edge of each graph.

and increases in the Rio Santa Barbara. These trends coincide with the general changes in mean size of bed material in the various streams. Size decreases along the Pecos and Pojoaque Rivers, remains approximately constant for the Rio Santa Cruz, and increases slightly in the Rio Santa Barbara (fig. 11-13). These differences in trend are due partly to bedrock and tributary influences, and probably are to some extent residual effects of Pleistocene glaciation and frost action on stream regime. The fact that downstream changes in tractive force and particle size are roughly parallel may be significant, but this cannot be ascertained until more definite relations between particle size and critical tractive force are established.

In summary, considerations based on the assumption that competent size is proportional to velocity squared, indicate that bed material is immobile even at bankfull discharges, except at stations near the downstream limits of this investigation. Velocity increases downstream for all cases considered here, but tractive force, which is another measure of stream competence, may increase, decrease, or stay nearly constant in the various drainage basins. Particles of quartzite and other lithologies have moved many miles from the source outcrops, but the possibility that they were transported during some previous period of greater discharge cannot be eliminated.

As mentioned previously, available field evidence suggests that bed material rarely moves even during bankfull flows. One additional piece of information supports this observation. Log dams have been installed at several places along some of the more heavily fished streams, especially in the vicinity of Cowles. The fact that these dams have been in place for a decade or longer without becoming filled with sediment is an indication that little transport of bed material occurs. Deposits behind beaver dams lead to the same conclusion, for most of them contain some sand and silt but very little gravel.

The opinion that little movement of bed material occurs in high mountain streams derives support from the work of Brush (in press), who estimated that 10 to 20 percent or more of the bed material in Appalachian streams is immobile at bankfull flows.

Any attempt to explain the apparently modest competence of high mountain streams must include several factors. Such streams are characterized by small discharges; also by small velocities, despite their steep slopes. The material which is delivered to them from steep valley walls is generally very coarse. Also, Pleistocene glaciation and frost action must have provided large quantities of coarse debris, as indicated by remnants of moraines, terraces, and other depositional features in most valleys. An additional consideration, which may be of great importance, is the fact that the water is essentially clear. If mountain streams carried greater suspended loads, their competency would be greatly increased. This viewpoint finds support in the character of materials carried onto alluvial fans or transported by arroyos.

The only definite conclusion which can be drawn from the foregoing discussion of competence is that additional experimental data are required before the "modern" versus "historical" aspect of streambed material can be interpreted.

Effect of Geology on Stream Characteristics

In the discussion presented thus far, the data have been organized on a hydrologic basis; that is, the relationships discussed apply to drainage basins as a whole or to specific segments of channel. The purpose of this section is to explore the possibility that differences in lithology and other geologic factors exert controls on stream characteristics which are not only independent of, but perhaps also greater in magnitude than, effects due to hydrologic factors.

DRAINAGE BASINS

Such easily measured properties of drainage basins as organization of the drainage net and basin form may differ on the various bedrock lithologies, but cannot be compared satisfactorily because of the varying degrees of quality of maps covering this area.

There seems to be no obvious structural or lithologic control of drainage-basin shape. However, basins of low-order streams on the Santa Fe formation appear to be considerably more elongate than basins of the same order on resistant rocks.

The relationship of drainage area to stream length apparently is not related to geology. For arroyos on the unconsolidated Santa Fe formation, $L = 1.7\ A_d^{0.54}$, with exceedingly little scatter for drainage areas ranging from 0.01 to almost 1,000 square miles (data from Leopold and Miller, 1956). As was shown in Figure 4, the various perennial streams in the mountains, which flow on several different lithologies, are characterized by essentially the same relation of length to drainage area. The similar results obtained by Hack (1957) and Brush (in press) for Appalachian streams indicate that the drainage area-length relation is essentially constant for the entire range of climatic and geological conditions. This means that such a plot probably is not sufficiently sensitive to show actual differences in basin shape.

DRAINAGE DENSITY

Fifteen sample areas, each a square mile in size, were chosen at random from the total outcrop area of each principal lithology. Drainage-density measurements were made on Soil Conservation Service maps prepared from air photos at a scale of 2 inches to 1 mile. The following results were obtained:

DRAINAGE DENSITY
(Miles per square mile)

	RANGE	MEAN	STD. DEV.
Sandstone	2.50- 5.70	3.93	0.97
Quartzite	2.00- 8.10	5.97	1.16
Granite	4.80-10.60	7.57	1.55
Santa Fe formation	9.20-13.75	11.55	1.46

As might be expected from its unconsolidated character and relative lack of vegetation, drainage density on the Santa Fe formation is considerably greater than for rocks of the high mountains. Also, it should be noted that the mean value of drainage density for sandstone is somewhat smaller than for quartzite and much smaller than for granite.

The significance of differences in drainage density between the various lithologic types of "hard" rocks cannot be evaluated until the effect of altitude on drainage density of a particular lithology is determined. Because topographic maps of only half the area are available, this relation cannot be investigated quantitatively at present. From inspection of planimetric maps, it seems likely that drainage density decreases slightly with altitude. Such an effect may be partly due to differences in the cover of vegetation which tends to obscure small channels.

Quartzite is confined to the highest altitudes in the area and is mostly above timberline. Sandstone and granite cover roughly the same altitude range, and both have considerable extent above timberline. Considering these distributions of the various lithologies, the effect of altitude may not greatly change the relative magnitudes of drainage density reported above. If this is true, then it seems reasonable that the smaller drainage density for sandstone is related to its greater infiltration capacity.

CHANNEL DIMENSIONS

It was mentioned previously that bankfull width and mean depth increase less rapidly with increasing drainage area for the Rio Santa Cruz and the Pojoaque River, which flow on granite, than for the Pecos River and the Rio Santa Barbara, which flow mostly on sandstone (fig. 7). Any major effect of an individual bedrock lithology should be apparent if the data in a plot of channel dimensions versus drainage area are grouped according to the bedrock type at each station. Conclusions drawn from such a plot are summarized below, but the diagram itself is not reproduced here.

The rate of change of width and depth with increasing drainage area is roughly the same for stations on quartzite, limestone, and sandstone, but stations on granite show a smaller rate of increase than the others. This effect, however, is not due to lithology alone, as can be shown by plotting the Pecos River stations located on granite by themselves. Such a plot indicates that width and depth increase at the same rate as for stations on other lithologies, and at a considerably greater rate than for stations on granite in the basins of the Rio Santa Cruz and the Pojoaque River. Thus, if observed differences in channel dimensions are the result of some property of granite, such a relationship applies only to basins underlain almost entirely by granite, and not to basins like the Pecos River where granite is only one of several lithologies.

CHANGES AT THE MOUNTAIN FRONT

Figure 18 is a plot of channel width versus drainage area. The category labeled "perennial" streams includes all the data for mountain stations; the data for width of ephemeral streams were taken from Leopold and Miller (1956). Such a representation is also a separation according to geology; namely, between "hard" rocks of the mountains and unconsolidated deposits of the Rio Grande Depression. Although the scatter for each category is great, there can be little doubt that the

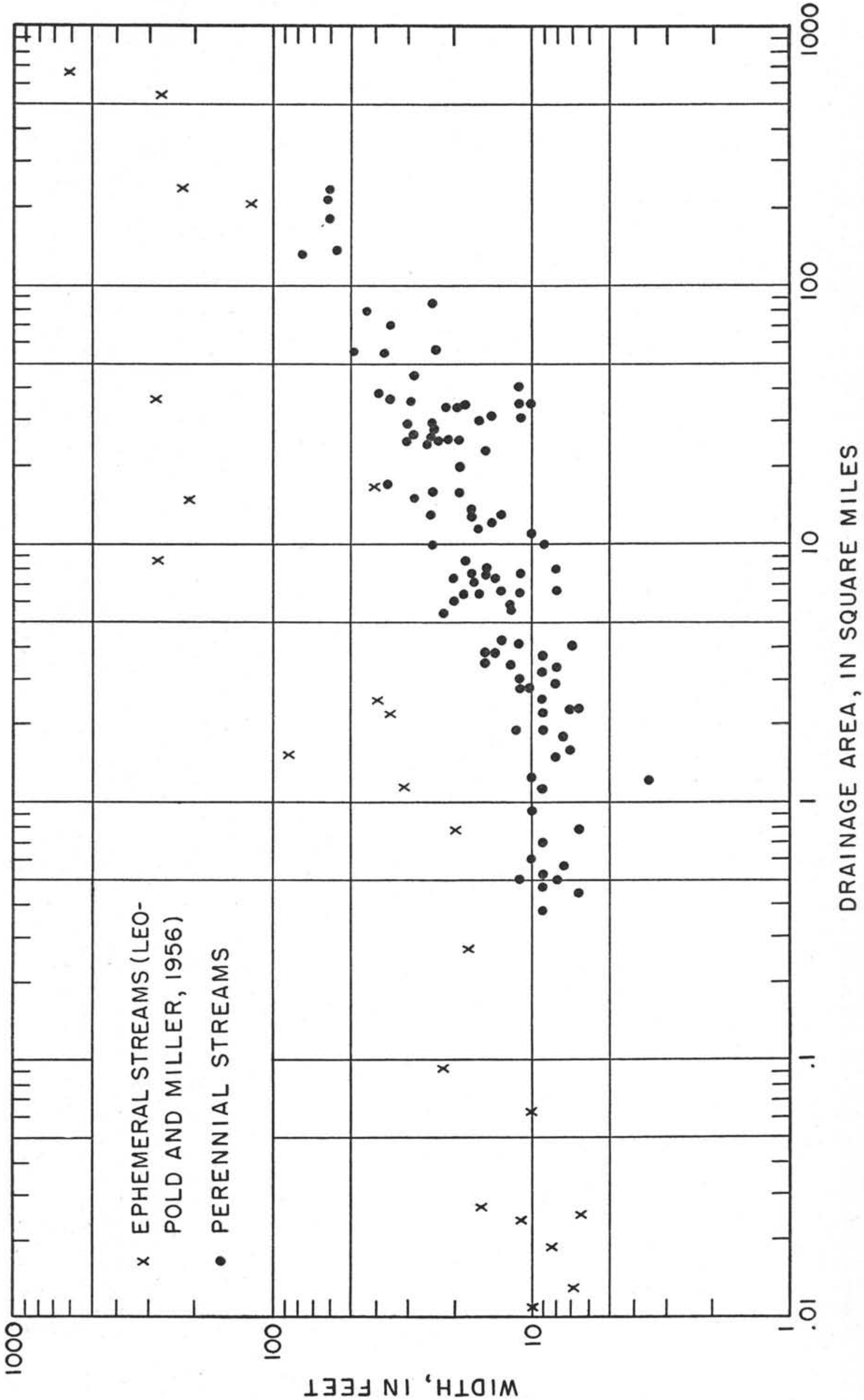


Figure 18

RELATION OF CHANNEL WIDTH TO DRAINAGE AREA, SHOWING DIFFERENCES BETWEEN PERENNIAL STREAMS OF THE MOUNTAINS AND ARROYOS OF THE LOWLANDS rate of change of width with increasing drainage area is similar in both, and furthermore, that channels of ephemeral streams are 3 to 4 times wider than nearby perennial streams with the same drainage area. Along any particular stream, this large change in width occurs within a few hundred feet downstream from the mountain front, as can be seen in Figure 19 and Plate 12. The small peaks in the curves of Figure 19 are of special interest because they apparently indicate an initial slight overadjustment of width in the unconsolidated deposits.

It seems fairly obvious that the abrupt change in channel width is due primarily to geological causes, specifically to inherent differences in erodibility of granite and the unconsolidated Santa Fe formation. However, climatic factors are involved also, because they are partly responsible for the differences in hydrologic regime of perennial and ephemeral streams. Furthermore, the effect of climate on vegetation is an important factor in determining the erodibility of channel banks. Because of these considerations, the general impression that geologic differences greatly outweigh climatic factors in causing the observed changes in channel width cannot be proved conclusively at present.

CHANNEL SLOPE

In the plot of channel slope against drainage area (fig. 20), bedrock lithology at each station is indicated. The scatter among points for mountain stations is very large and shows no significant separation according to lithology. As might be expected, channel slopes for any particular drainage area are less on the Santa Fe formation than in the mountains. Likewise, the rate of decrease of slope with increasing drainage is greater for mountain streams than for the arroyos. Locally, several different relationships of particle size of bed material to channel slope are represented by the data plotted in Figure zo. For the Rio Santa Barbara (fig. 11), both particle size and slope increase downstream with increasing drainage area. In the case of the Pecos River (fig. 13), both particle size and slope decrease; for the Rio Santa Cruz (fig. 12), particle size remains approximately constant, but slope decreases at about the same rate as for the Pecos River. Particle size is approximately constant for arroyos, and slope decreases, though at a much smaller rate than for most mountain streams.

As is shown in Figure 21, channel slope of mountain streams decreases with increasing stream width (meaning also increasing drainage area and discharge), but scatter among the plotted points is extremely large. In general, narrow, steep channels are restricted to headwaters reaches, but there are several exceptions to this rule. Glacial features, changes in lithology, and tributary influences provide explanations for some of the exceptions. It should be noted in Figure 21 that differences between lithologies of "hard" rocks appear not to affect the change in channel slope with increasing width. Furthermore, the data do not indicate any consistent relation of slope to width at constant drainage area (or discharge); that is, isopleths of drainage area cannot be drawn through these scattered points. Decrease of slope with increasing width is only about half as great for arroyos as the overall rate for mountain streams.

In summary, the relations of slope to channel width and drainage area show a lithologic separation into only two categories, hard rocks of the mountains and unconsolidated deposits of the Rio Grande Depression. This fact leads to the rather obvious conclusion that combined effects of geologic factors are more important determinants of both absolute values of channel slope and downstream rates of change of slope than are hydraulic factors. In particular, the differences in available relief between the two areas of contrasting lithology were determined by previous geologic history. This means that slope is a partially independent variable, determined by inherited conditions which the stream can gradually modify, but only within certain limits.

BED MATERIAL

The most striking effect of geology on bed material in stream channels occurs at the boundary between the hard rocks of the mountains and unconsolidated deposits which comprise the Santa Fe formation. Over short distances downstream from the mountain front, channel width increases markedly, and bed material changes from particles of cobble and boulder size to dominantly coarse sand, with scattered pebbles which cover less than 1 percent of the channel surface (pl. 12).

Perennial streams draining southward and eastward from the Sangre de Cristo Range undergo more gradual transitions in size of bed material. As was shown in Plate 1o, marked changes in bed material occur along the Pecos River; this change, however, is not accompanied by a great increase in channel width. In contrast to the Santa Fe formation, the Paleozoic and Mesozoic sedimentary rocks exposed in the Pecos River basin downstream from the main mountain mass are well consolidated; nonetheless, they apparently yield very large amounts of sandy sediment. Sidwell (1941), who studied heavy minerals in the Pecos River between Dalia and Carlsbad, noted that the bed material consists almost entirely of poorly sorted sand (mostly < 1 mm). In another paper, Sid-well (1939) mentions that gravels disappear from the bed of the Canadian River, which drains the eastern slope of the Sangre de Cristo, in roughly the same distance as for the Pecos.

Within the mountains, local bedrock of most, but not all, lithologies greatly affects the composition and size of bed material in stream channels (fig. 3). Data obtained in this investigation but not plotted in Figures 11-13 can also be used to show these effects by considering the fraction of bed material with the same composition as the bedrock at the station. Obviously, such an analysis requires that stations which drain monolithologic areas be eliminated. If this procedure is followed and all four drainage basins are considered together, it is found that sandstone makes up more than 50 percent of the bed material at stations where it crops out, whereas granite, amphibolite, and quartzite comprise 15 to 20 percent of the bed material at stations where they crop out. These differences suggest that sandstone weathers somewhat more readily than the other lithologies and therefore yields more debris to the stream channels.

Actually, the situation is not this simple, because of practical problems in accurately defining the lithology at a particular station. Topographic and geologic relations along the Pecos River serve to demonstrate this point. The Pecos Valley is a deep gorge, at most places cut entirely through the Paleozoic sediments into the underlying Precambrian metamorphic and igneous rocks (pl. 4).

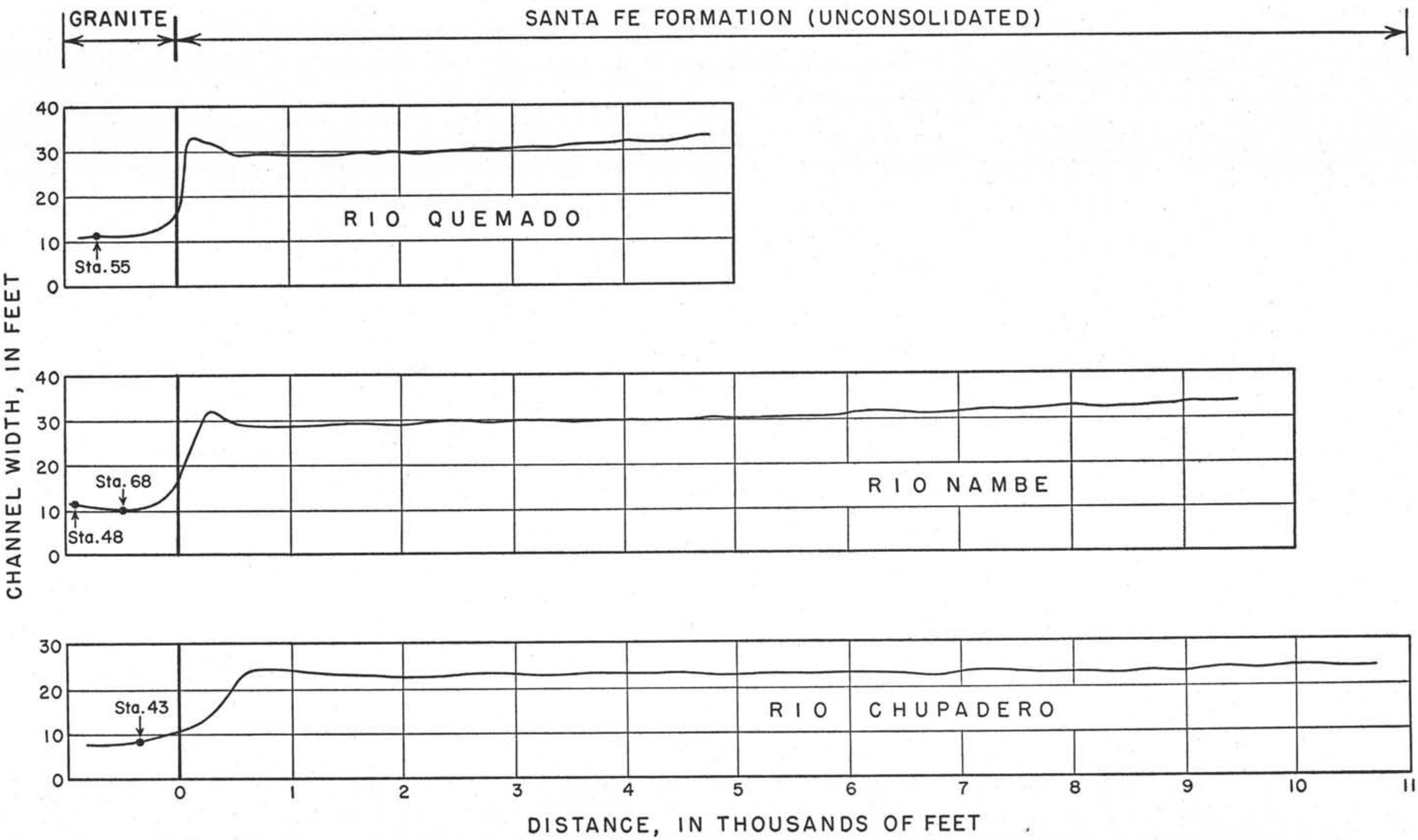


Figure 19

EXAMPLES SHOWING CHANGE OF CHANNEL WIDTH AT THE MOUNTAIN FRONT

Width measurements were taken at intervals of 100 to 200 feet in the first quarter mile downstream from the mountain front, and subsequently at intervals of 500 to 1,000 feet.

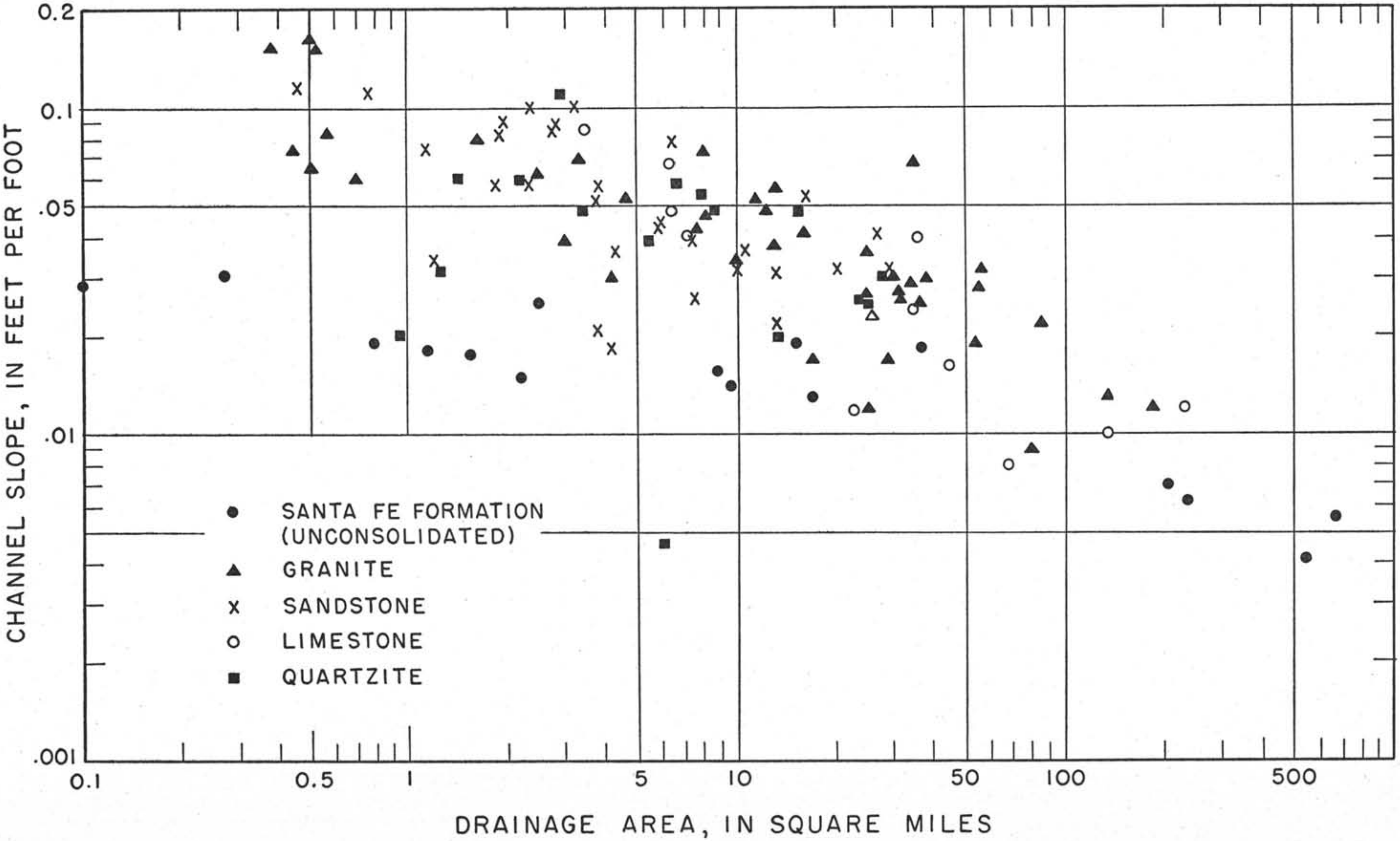


Figure 20

Relation of channel slope to drainage area for mountain stations

Lithology of bedrock at each station is designated. The data for arroyos on the Santa Fe formation are taken from Leopold and Miller (1956).

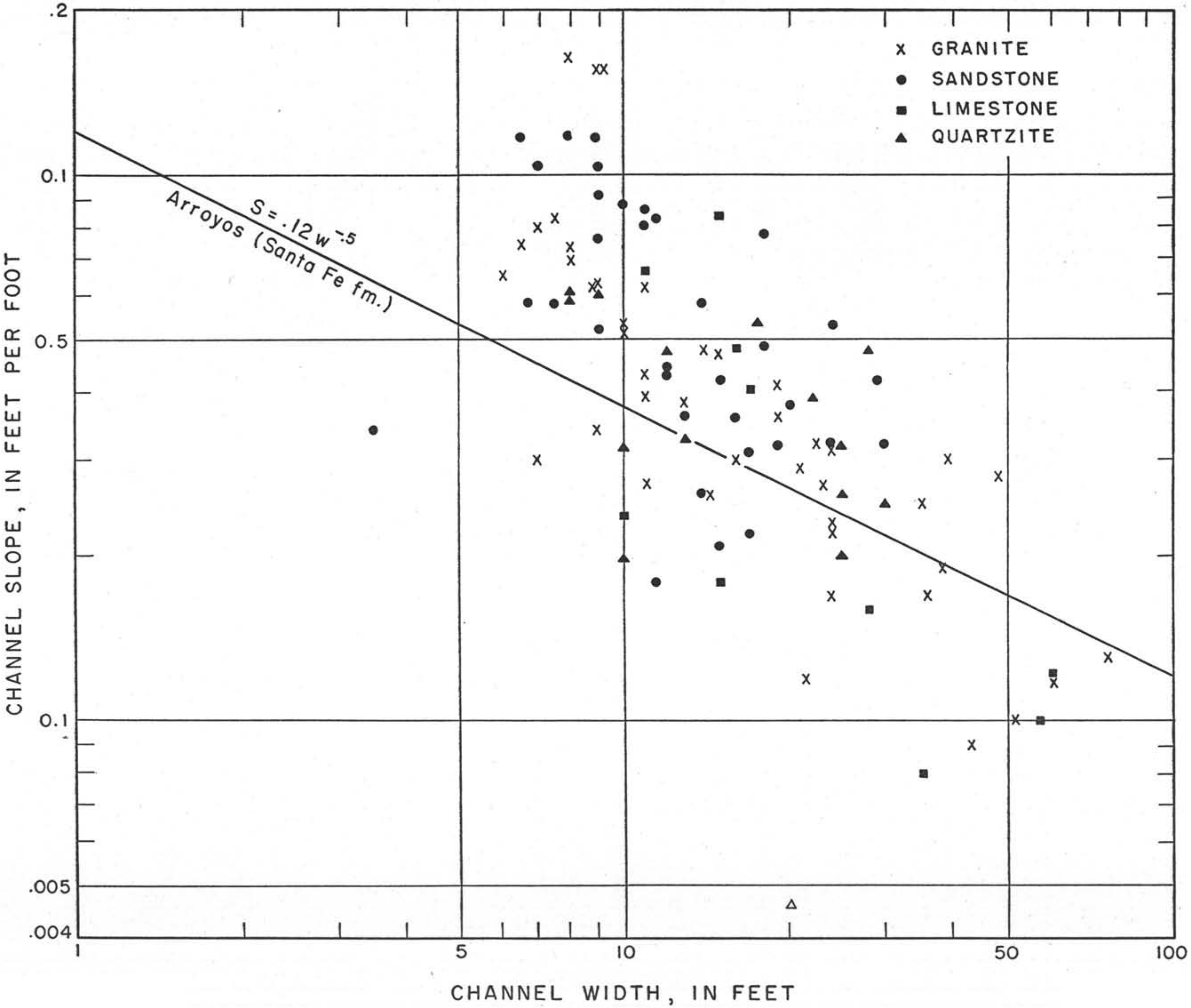


Figure 21

RELATION OF CHANNEL SLOPE TO CHANNEL WIDTH FOR MOUNTAIN STATIONS

Lithology at each station is designated, and the curve fitted through plotted points for arroyos is shown for comparison.

Thus, the bedrock lithology in the immediate vicinity of stations in the gorge is one of the Precambrian rock types. At most places, however, the overlying Paleozoic sedimentary rocks form high cliffs which contribute debris to the stream. The importance of this factor can be seen by considering the case of limestone bed material in the Pecos River. Where limestone occurs as relatively restricted outcrops near the bottom of the valley (for example, miles 13-24, fig. 13; see also pl. 4), it does not contribute more than 30 percent of the bed material, and at most places contributes less than r o percent. Downstream from station 52 (mile 26.5, fig. 13), where limestone becomes much more abundant in the sedimentary rocks of the valley walls, it is the dominant lithology of the bed material. Thus, the abundance of a particular lithologic type in the bed material of such streams is related to the ease of weathering, size of the source area, and resistance to abrasion in the channel.

It was noted at several places that limestone fragments contributed to the bed material from relatively small outcrops in the channel persist only short distances downstream (generally a few hundred feet). Conditions for solution are exceedingly favorable in these streams and probably play a major part in this relation. Hack (1957) also noted a rapid decrease in amount of limestone downstream from outcrops in Appalachian streams.

Only small percentages of shale and schist-phyllite occur in the bed material of Sangre de Cristo streams even where large outcrops occur. Also, particles of these lithologies break up and disappear within very short distances downstream from outcrops.

The effect of local lithology on particle size of bed material is seen somewhat more readily by examining maximum rather than mean particle size. As shown by the data summarized in Table 14, sizes of the largest particles derived from granite, sandstone, quartzite, and amphibolite exceed by two or more times the size of the largest limestone fragments. The observed differences in maximum size between lithologies are not related to stream length. At the majority of stations where granite, sandstone, quartzite, or amphibolite is exposed, the local lithology contributes the largest particles present in the bed material. Except for station 7 (farthest downstream on the Pecos River), there is no station at which the largest fragment of limestone exceeds the maximum size of other lithologies represented.

Differences between lithologies in the maximum sizes of fragments contributed to streambed material are due primarily to jointing and to weathering characteristics. The additional effect of faulting on particle size is obvious at several stations, especially along the West Fork of the Rio Santa Barbara. The headwaters portion of this valley coincides with a major fault for a distance of several miles (pl. 1), and at many places in this reach, quartzite fragments in the bed material are relatively small. The marked increase in size about 7 miles from the divide occurs at a high quartzite cliff which lies west of the fault. Another area where size of streambed material is affected by faulting is near the head of Horsethief Creek. There are several places, especially in the drainage basin of the Rio Santa Barbara, where terrace deposits, glacial debris, or frost-rubble accumulations seem to influence the character of streambed material more than the local bedrock.

POOLS AND RIFFLES

The preceding description of geological factors which affect the size of bed material in high mountain streams paves the way for discussion of the previously mentioned fact that Sangre de Cristo streams at most places lack well-defined pools and associated riffles. The typical condition is shown in Figure 22, which is a map and profile of a 5oo-foot segment of Holy Ghost Creek at an altitude of about 8,600 feet. Deeps and shoals typical of alternating pools and riffles do not appear in the profile. Scattered large boulders in the channel cause the principal breaks in the profile.

It should be pointed out that pools are not completely absent in Sangre de Cristo streams. In high mountain meadows (e.g., pl. 5A), the relations of bed-material size and channel slope to stream velocity apparently are suitable for development of pools and riffles. Also, near the downstream limits of all the basins considered (within the mountain mass), pools occur locally. In all cases, however, they are restricted to short segments of channel and, hence, do not exhibit a characteristic length or spacing.

Leopold and Wolman (1957, p. 59) state that pools and riffles are fundamental features of nearly all natural streams, and that their spacing is proportional to bankfull discharge. The only apparent explanation for the general lack of pools and riffles in Sangre de Cristo streams is that discharges (and hence velocities) are too small to handle the coarse material present in the channels. Most of the literature on pools and riffles refers to rivers with greater discharges and smaller bed material than is typical in high mountain streams. This point is demonstrated by Figure 23, which compares the ranges in size of bed material, channel slope, and bankfull discharge of Sangre de Cristo streams with ranges of the same properties for streams described as having pools and riffles. The comparative data used are for Brandywine Creek, Pennsylvania (Wolman, 1955), and streams in the Appalachians of central

TABLE 14. SIZE OF LARGEST PARTICLES CONTRIBUTED TO THE STREAMBED MATERIAL BY VARIOUS BEDROCK LITHOLOGIES

For each station, the size of the largest particle with the same lithology as the local bedrock was tabulated. Data for all stations having the same bedrock lithology were combined in the size distribution.

LITHOLOGY	PERCENTAGE OF STATIONS WHERE LARGEST PARTICLE IS FINER THAN GIVEN SIZE (in mm)									SIZE RANGE (in mm) OF LARGEST PARTICLES
	100	200	300	400	500	600	700	800	900	
Sandstone	3	3	20	40	57	70	90	90	97	97-1,230
Limestone	9	54	82	82	100					72-495
Quartzite	0	17	40	52	64	76	88	94	100	141-890
Granite	0	6	17	34	60	71	82	88	91	115-1,160
Amphibolite		0	38	50	62	100				225-570

Pennsylvania (Brush, in press). Brush's data partially overlap those of the present study, but some of the headwaters streams which he studied also lack pools and riffles. Rivers mentioned specifically by Leopold and Wolman (1957) as having pools and riffles contain bed material even smaller than the examples shown in Figure 23.

In summary, the available evidence seems to indicate that the bankfull flow of a high mountain stream is essentially powerless to cope with the bouldery debris in its channel. This coarse material is derived in part from present-day weathering of steep cliffs, but commonly a portion of it consists of deposits formed as a result of Pleistocene glaciation and frost action. Pools and associated riffles can develop only where some transport of bed material is possible. Bankfull discharges of high mountain streams are insufficient to provide the velocity required for transport except at a few places where bed material is relatively small, generally for some special geological reason.

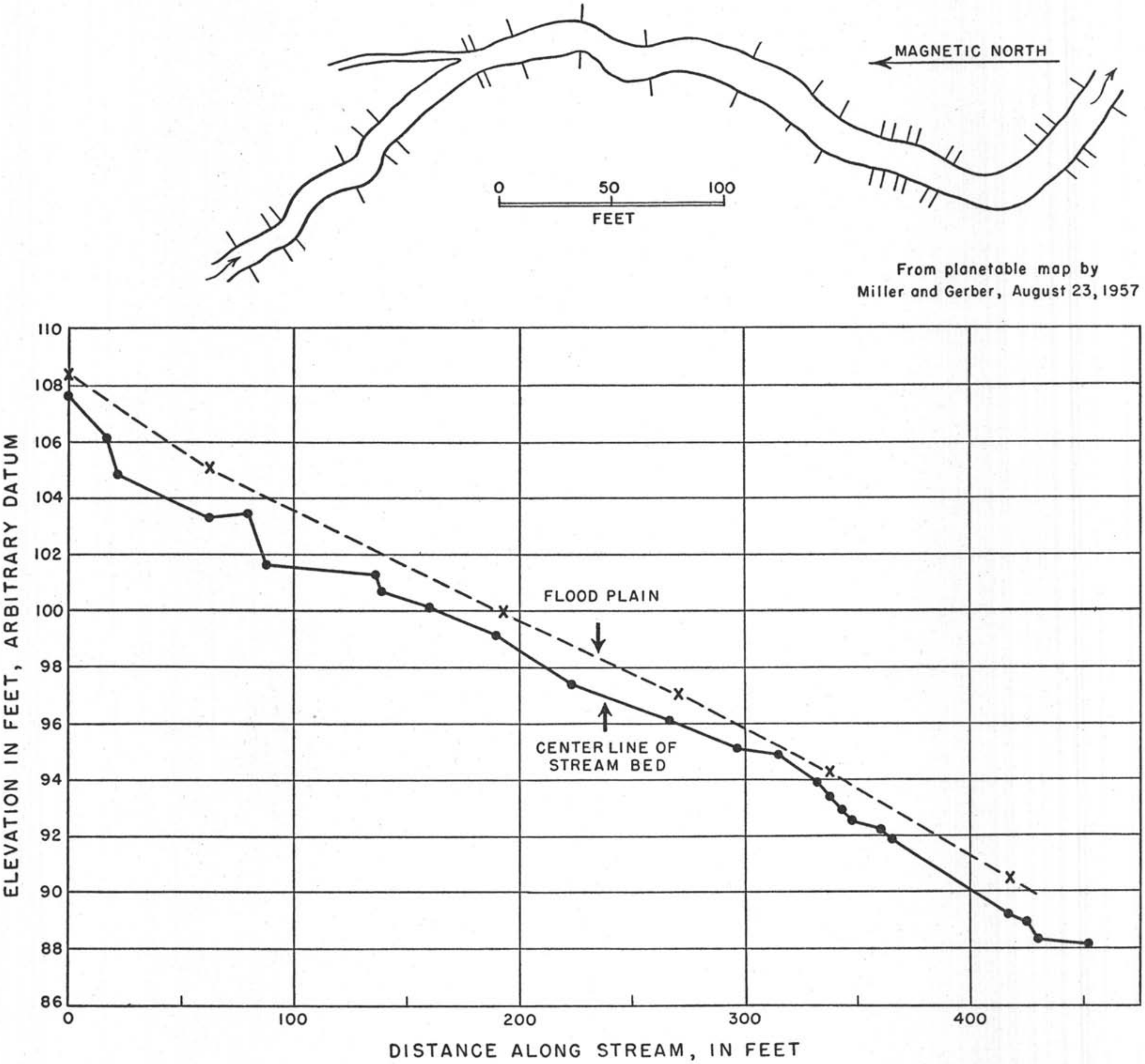


Figure 22

MAP AND PROFILE OF HOLY GHOST CREEK AT END OF CAMPGROUND ROAD (NEAR STATION 38)

Pools and riffles are absent; irregularities in the profile are caused by boulders in the channel.

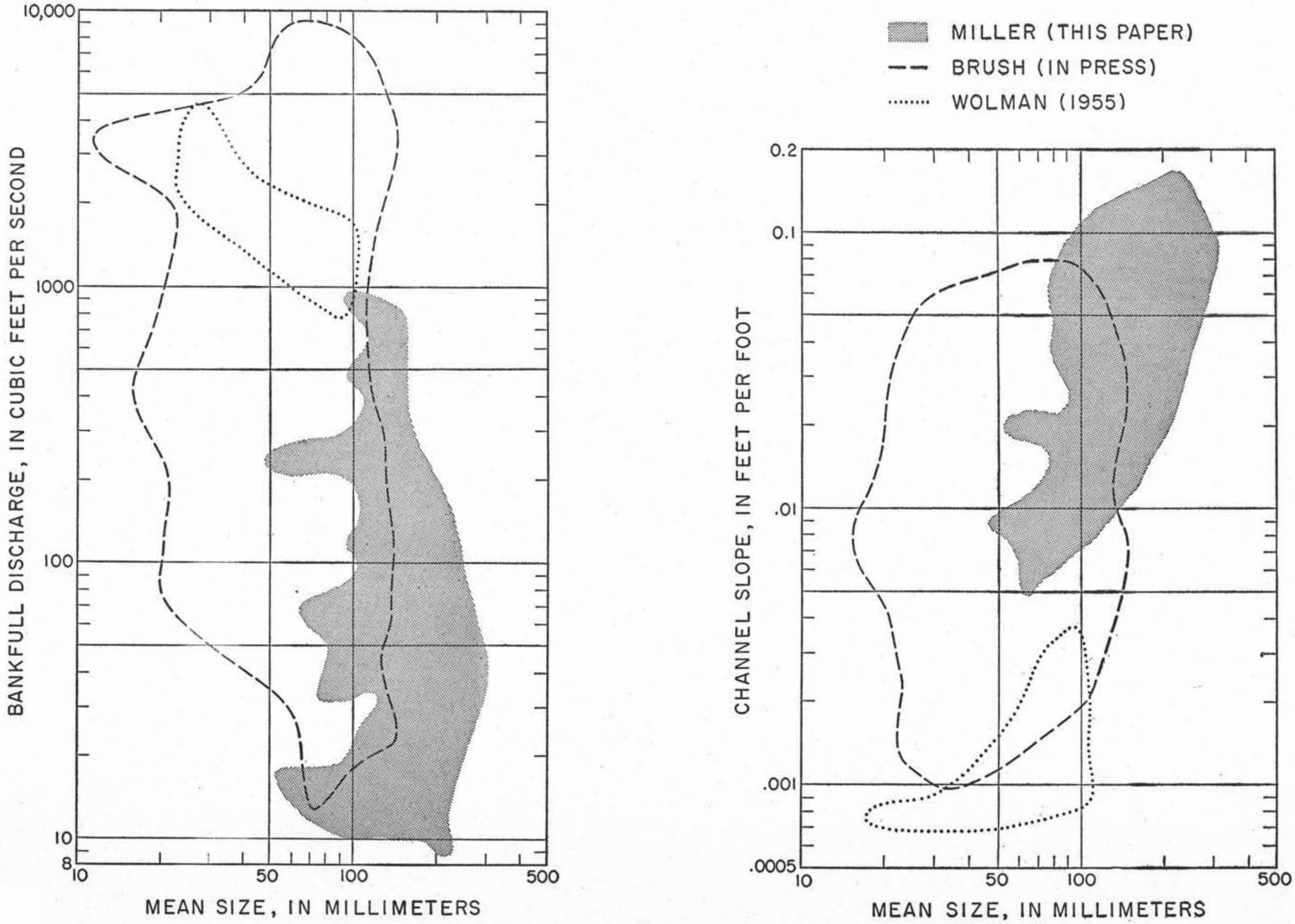


Figure 23

COMPARISON OF SANGRE DE CRISTO STREAMS WITH STREAMS IN PENNSYLVANIA

Ranges in size of bed material, channel slope, and bankfull discharge for streams in Pennsylvania which have pools and riffles, compared with Sangre de Cristo streams which lack pools and riffles.

Equilibrium in High Mountain Streams

The main reason this study of high mountain streams was undertaken is the fact that such streams are generally considered to be ungraded. This opinion has been stated, or at least implied, by many authors; for example, Davis (1902, p. 92), Mackin (1948, p. 483), Leopold and Maddock (1953, p. 51), and Culling (1957, p.266). Much has been written in recent years on the concept of grade or equilibrium, as applied to streams, but few usable criteria for recognition of such a condition have been developed. Study of ungraded streams has been largely neglected; by definition they presumably lack the diagnostic characteristics of graded streams. Consequently, it was hoped that the definition of grade might be sharpened by attempting to discover what it is not, through a study of headwaters streams thought to be ungraded.

Before comparing the properties of high mountain streams with those attributed to graded streams, it is necessary to review briefly some recent statements on the nature of the graded or equilibrium state and how it is recognized. Mackin (x948), in his comprehensive treatment of the subject, maintains that adjustment of channel slope is the major response of a graded stream to changed conditions. His principal criterion for recognition of grade is lateral cutting, which produces a surface of corrasion. Woodford's definition (1951, p. 819) is similar to that of Mackin, but he foresees serious difficulties in recognizing the condition of grade (p. 816). Holmes (1952, p. 900-901) suggests that graded valley walls and competence to move the largest particles in the channel are essential requisites for equilibrium. He also raises the question of whether the graded condition can be recognized as soon as it is attained. Except for their emphasis on adjustability of the channel cross-section rather than slope, Leopold and Maddock (1953, p. 46) agree in essence with Mackin's definition of grade. They find a close analogy between graded streams and regime canals. However, they also note certain striking similarities between graded and ungraded streams. Wolman (1955, p. 48-49) expressed views rather similar to those of Leopold and Maddock, and pointed out that the concept of equilibrium implies both stability and the ability to make adjustments. He could find no definite criteria for distinguishing the graded condition from an ungraded one. Leopold and Miller (1956, p. 28) are the latest writers to emphasize that some of the symptoms of grade also characterize supposedly ungraded streams. Culling (1957) considers the nature of the graded profile, but he does not present any new criteria for recognition of equilibrium conditions.

The foregoing review of recent literature demonstrates that there is considerable uncertainty about the diagnostic properties of graded streams. In this connection, it should be noted that there have been two rather different approaches to the problem. One involves the form of the graded longitudinal profile, whereas the other is concerned with adjustments in response to the amount and particle size of load supplied. With regard to the latter approach, there are differences of opinion about the relative adjustability of slope as opposed to channel cross-section. The dearth of reliable criteria for recognizing equilibrium conditions in modern streams is a serious shortcoming of the concept of grade.

Returning now to the consideration of high mountain streams, our objective is to determine the extent to which their properties differ from those of graded streams. The fact that the water of Sangre de Cristo streams is confined in definite channels indicates a certain degree of adjustment. It was shown (table 1) that channel width and depth are rather variable locally, but probably no more so for mountain streams than for others. Also, it was found (table 2) that these streams undergo systematic changes in channel dimensions at tributary junctions. In fact, the data presented in this paper show that downstream relations of channel dimensions to drainage area (and discharge) are very similar for both mountain streams and streams at lower altitudes (table 7). Furthermore, the downstream rates of change of channel dimensions of mountain streams compare as favorably with the same properties of regime canals as do the examples cited by Leopold and Maddock (1953) and Wolman (1955).

Mackin (1948, p. 491) emphasizes that the profile of a graded stream is a slope of transportation, not influenced directly by either corrasive power of the stream or by bedrock resistance. Holmes (1952, p. 900-901) states that a graded stream must have the competence to carry whatever sediment is delivered to it, which he interprets to mean that debris cannot be shed directly from the valley walls into the channel. Culling (1957, p. 263) makes statements which are similar to those of Holmes.

At many places, though not all, Sangre de Cristo streams do not have these characteristics of grade. Several lines of evidence indicate that most of the streambed material is immobile at bankfull flows. Furthermore, there are many places where steep cliffs contribute coarse debris directly into the channel.

It seems appropriate, however, to inquire how essential more or less continuous transport of elastic bedload is to the concept of grade. Does it mean that a stream which carries no elastic load can never achieve equilibrium? A stream flowing on an immobile bed is no different in this respect from one flowing on bare bedrock. Furthermore, even if we grant that transport of elastic bedload is an essential part of the definition of grade, what kinds of exceptions are allowed? Surely we could not change a graded reach to an ungraded condition simply by putting in one boulder too coarse for transport. If 1 isn't enough, how about 10 or 50?

High mountain streams generally lack pools and riffles, apparently because of small discharges and very coarse bed material. According to Leopold and Wolman (1957, p. 59), pools and riffles are features of nearly all natural channels, but whether this refers to graded streams only is not stated. As was mentioned previously, the channel of a mountain stream has a definite thalweg, which at most places wanders or meanders in much the same manner as for streams at lower altitudes. Except locally, high mountain channels do not have meanders or braids, nor are they perfectly straight. In any case, channel pattern probably is of little use for distinguishing between graded and ungraded streams (Leopold and Wolman, 1957, p. 73).

Mackin (1948, p. 483) notes that ungraded profiles are superficially similar to graded profiles, though completely different in origin. Leopold and Maddock (1953, p. 51) suggest that conditions of approximate equilibrium are attained as soon as a more or less smooth longitudinal profile is established. How smooth the profile must be, or the maximum extent to which the slope of a graded stream can change at lithologic contacts, is not clear. Mackin (1948, p. 491) maintains that differences in bedrock resistance have no direct influence on the slope of a graded stream, though slope may be affected if contrasting lithologies alter the amount and caliber of load supplied. If this is true, there still remains the question of how to distinguish a graded stream, in which slope changes are caused by differences in amount and caliber of load, from the ungraded condition, in which slope changes at lithologic contacts are due entirely to differences in rock resistance.

Longitudinal profiles of several streams on the western slope of the Sangre de Cristo Range (see examples in fig. 24) are fairly smooth except in the headmost portions. Also, except for the falls on the Rio Nambe, no major break in slope occurs at the contact between the unconsolidated Santa Fe formation and the more resistant older rocks. Although smooth profiles of the kind shown in Figure 24 may suggest the condition of grade, there are several other facts to be considered. As was mentioned previously, a roughly threefold increase in channel width occurs at the mountain front. However, in cases where the stream has cut gorges through several closely spaced outliers (e.g., the Rio Quemado, fig. 24), the widening occurs only below the last granite exposure, where size of the bed material decreases abruptly. Coarse debris is shed directly to the channel in each of the short gorges but does not seem to affect channel slope. Another point is that, in contrast to the situation for streams at higher altitudes, a large fraction of the bed material in reaches near the mountain front probably can be transported at the bankfull stage. Finally, there is abundant evidence that the arroyos of the Rio Grande Depression either are, or recently have been, actively downcutting (Leopold and Miller, 1956; Miller and Wendorf, 1958).

Profiles of streams elsewhere in the high mountains cannot be presented, because topographic maps are not available and no special surveys were made. However, such information as was obtained at sampling stations suggests that slope changes at most lithologic contacts range from moderate to very slight. The greatest breaks in slope occur near the heads of the glaciated valleys (fig. 24) and are apparently the result of glacial erosion.

Mackin (1948, P. 493-494) states that a graded stream will steepen its slope in response to an increase in load or a decrease in discharge. If these are typical reactions of a graded stream, we may inquire about the behavior of an ungraded stream subjected to the same conditions. As has been discussed previously, Pleistocene glaciation must have affected discharge and load of Sangre de Cristo streams, and apparently left a residue of material too coarse for modern streams to transport, except possibly at very infrequent intervals. At many places, additional coarse debris is being contributed to the channels by modern processes. Should we expect an increase of slope, so as to promote transport, or is this a property restricted to graded streams? Whatever answer one might deduce, the fact is that evidence of slope adjustment by aggradation is nonexistent; that is, alluvial deposits (mostly gravel) in and adjacent to channels are thin everywhere, though they apparently thicken slightly downstream.

It might be supposed that the presence of a flood plain is an indication of grade or equilibrium. Wolman and Leopold (1957, p. 103), however, present evidence that this may not be true. In any case, the high mountain streams considered here lack flood plains except at a few places.

In principle, one cannot object to Mackin's (1948, p. 492) statement that recognition of the differences between graded and ungraded streams is an essential prerequisite for efficiency in gathering and analyzing stream data. However, as has been brought out by the foregoing discussion, the necessary ground rules for making such distinctions have not been formalized; indeed, there is reason to believe that it may be impossible to do so. The present study was begun on the presumption that high mountain streams have not achieved an equilibrium condition. However, it is not possible to conclude unequivocally that Sangre de Cristo streams are ungraded. They definitely possess many of the characteristics generally associated with the condition of grade. On the other hand, they appear to lack certain other properties which are supposed to be typical of graded streams. It seems probable, however, that some of these shortcomings may reflect deficiencies in the present definition of equilibrium conditions. Thus, results of the present investigation seem to blur rather than sharpen the definition of grade, hence rendering this concept even more elusive than before.

As a general framework for studying ancient stream features, the concept of grade seems to serve a useful purpose. It cannot be denied that many streams have in the past achieved a condition of grade, as is shown by corrasion surfaces (rock-cut terraces). However, even this supposedly infallible criterion of grade calls for further study to determine the magnitude of imperfections in such surfaces.

As applied to modern streams, the concept of grade has not yet undergone any appreciable fraction of the testing which any valid fundamental concept must withstand. No one has yet proved conclusively that certain streams, or certain reaches, have attained a steady-state equilibrium, whereas other streams, or other reaches, are tending toward this condition but have not yet reached it. If the fact of equilibrium has not been proved and its attendant conditions described, surely it is too early to make prejudgments about which variables can provide unmistakable field evidence of equilibrium or the lack thereof. Many writers have pointed out that several of the variables involved in stream action are mutually interdependent. The greatest need at present is for additional quantitative information on the complex interrelations between factors affecting transport of sediment particles by flowing water.

In conclusion, it seems necessary to add one further remark so as to forestall misinterpretation of the argument set forth above. The author does not recommend abandonment of the terms *grade* and *equilibrium,* nor necessarily declaration of a moratorium on their use. However, it has been argued that the terms *graded* and *ungraded* as applied to modern streams are without precise meaning. For this reason, there is a real need for reexamination of the concept of equilibrium, by subjecting it to the detailed, rigorous investigation which it deserves by virtue of its prominence in present-day geomorphic thought.

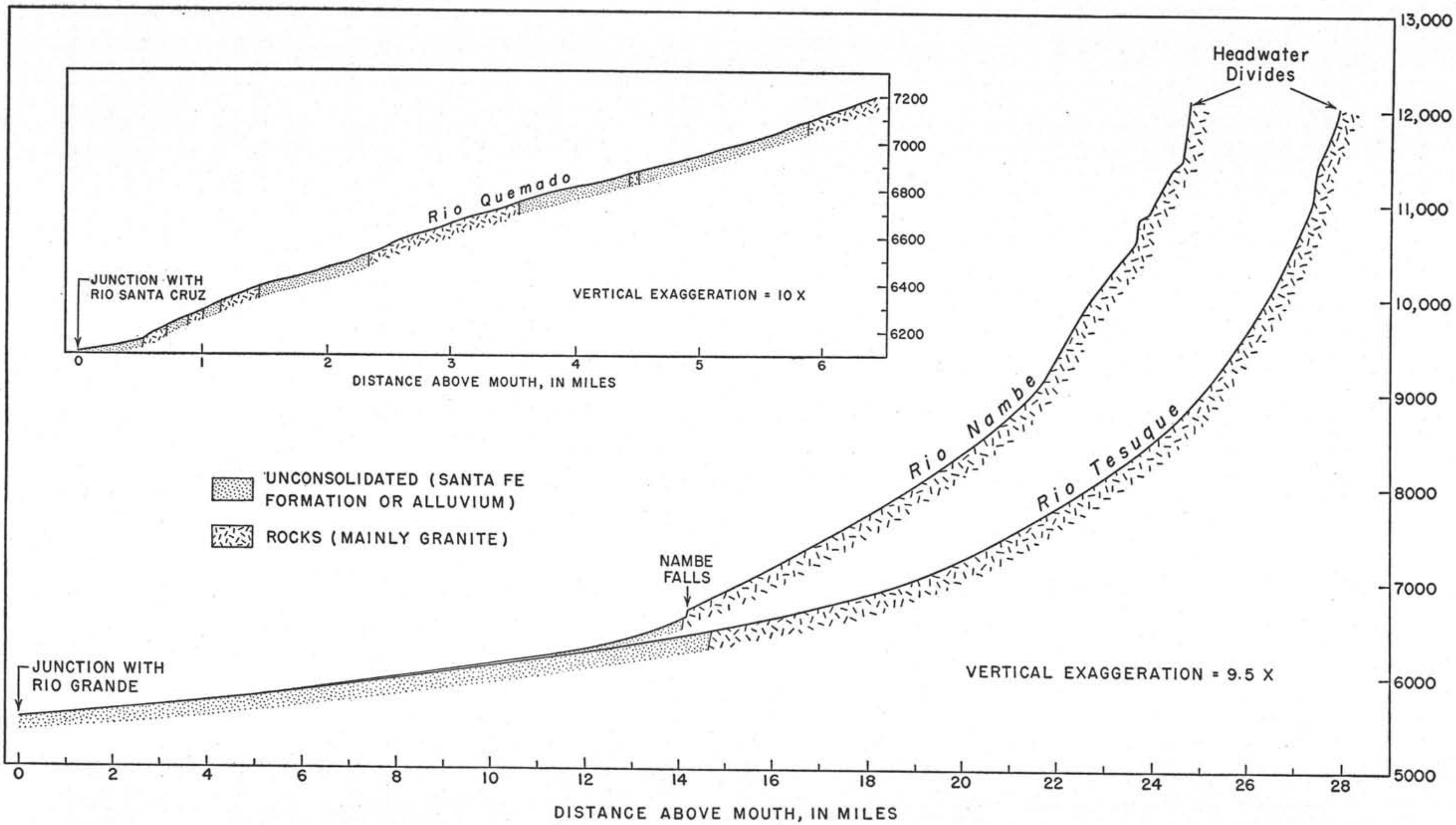


Figure 24
LONGITUDINAL PROFILES, SHOWING RELATIONS OF GEOLOGY TO CHANNEL GRADIENT

Conclusions

This investigation of high mountain streams serves to emphasize some of the numerous difficulties involved in attempting to understand the behavior of modern streams. The main problems arise from the fact that most streams which are small enough to be studied conveniently and safely lack a hydrologic record of any kind. Although there are uncertainties stemming from this and other sources, the data obtained seem to indicate several important conclusions.

Despite their generally steep slopes, high mountain streams have many properties in common with streams at lower altitudes. This is especially true for downstream changes in channel dimensions and velocity. Bed material, which is very coarse at all except a few places, may increase, decrease, or remain constant in size downstream; local changes in composition and particle size depend especially on the bedrock lithologies traversed, as well as on contributions from tributary streams. The amount of bed material transported during any long period of time (centuries or millennia) is unknown, but it seems evident that flows up to the bankfull stage do not attain sufficient velocities for appreciable transport of the coarse material present in most channels.

Except for effects resulting from the marked differences between hard rocks of the mountains and unconsolidated deposits at their western margin, geology seems to exert no recognizable influence on channel properties, although it definitely affects the size and lithologic composition of bed material. It was noted that streams of basins underlain entirely by granite have smaller channels than those on other "hard" rock lithologies, but probably this is not completely an effect of lithology.

Perhaps the most significant result of this study is the demonstration that "ungraded" streams of the high mountains have so many of the properties generally attributed to graded, or equilibrium, streams, that there are no definite criteria for distinguishing between them. This conclusion indicates that a detailed reexamination of the concept of grade is in order.

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