Clastic Sedimentation in Deep Springs Valley California

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By LAWRENCE K. LUSTIG

EROSION AND SEDIMENTATION IN A SEMIARID ENVIRONMENT

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A study of the bolson environment and the formation of alluvial fans

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CLASTIC SEDIMENTATION IN DEEP SPRINGS VALLEY, CALIFORNIA

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ABSTRACT

This report treats the size distribution of sedimentary deposits in Deep Spring Valley, Calif., and the formation of alluvial fans.

The distribution of clastic sediments and the parameters of the size distribution have been mapped when possible. Maps of the mean size, mean roundness, and lithology of pebbles, and the weight percentage of granules and silt: clay ratios of the granule-to-clay size fraction are presented. Maps of the mean size, standard deviation, skewness, and kurtosis of the granule-to-clay size fraction are also included. The largest particles exhibit fluctuation in size in a downfan direction and cannot be mapped on a reasonable contour interval; the relation of maximum particle size to local slope is shown on scatter diagrams. These data indicate that size fluctuation and poor sorting of sediments are characteristic of the environment.

Granules range in abundance from about 3 to 5 percent in the basin center to about 15 to 20 percent near bedrock outcrops. Because the abundance of granules is greatest in coarse polymodal sediments—whereas their abundance decreases with decreasing grain size and the tendency toward unimodality—it is suggested that granules consist of aggregates of sand-sized particles that are rapidly reduced to these components by mechanical weathering rather than by the action of water. The general absence of granules in ancient sediments is thought to be related to the scarcity of discrete constituents in source rocks in the 2- to 4-mm size range.

Clay is scarce in both the surface sediments and modern mudflows. The weight percentage of clay in the granule-to-clay fraction ranges from about 0 to 6 percent and averages about 2 percent in the 90 samples studied. Silt is much more abundant than clay and is the primary mode in most of the polymodal samples. The mapped distribution of silt-clay ratios shows that these ratios are probably governed in part by wind transport and deposition. The path of infinite silt: clay ratios is interpreted as a reflection of the present wind track in the basin.

The maps of mean size, standard deviation, skewness, and kurtosis of the granule-to-clay size fraction are interrelated, and proper interpretation of both the distribution and mixing of sediments within the basin is greatly aided by considering this interrelation. The tails of the distribution, namely granules and silt-clay, are shown to be of primary importance because they exert a marked effect upon the several parameters. Skewness and kurtosis are shown to be mappable and useful parameters in such studies. The field zero skewness, for example, defines a zone of sediment mixing in the basin because it reflects a balance between the tails of the size distribution. Kurtosis isopleths, however, show a high that trends across the zero skewness field, indicating that, although balance of the tails is attained, the magnitude of the spread of the tails increases. It is thought that skewness and kurtosis, which are commonly neglected in basin studies, provide information of a nature not attainable by other means. Moreover, absolute values of kurtosis may be of importance in sediment transport problems because these values probably reflect the competence of flow to some extent.

A semiquantitative method, based upon measurement of the integrated intensity of the (001) reflection, was used to determine clay-mineral ratios. The data show that a considerable range of montmorillonite: illite values occurs despite the rather uniform climatic conditions that prevail, thus suggesting a relation between clay minerals and source-rock composition. Kaolinite and (or) chlorite comprise about 20 percent of the average sample and are relatively constant in abundance.

The competence of transport is approximated by using field data to estimate tractive force in the expression $\tau = \gamma\, dS$. Maximum particle size and local slope values are substituted for d, the depth of flow, and S, the slope of the energy gradient, respectively; the specific weight γ is omitted. The dS products, which are directly proportional to tractive force, are given for about 500 sampling stations, and the distribution of values is shown by maps.

A set of orthogonals, drawn through isopleth trends, is thought to represent paths of sediment transport. The predictive possibilities of the method are tested, and the results show that this particular set of orthogonals does represent sediment transport paths because the orthogonals coincide with active channels on alluvial fans. The results also show that tractive force within the channels is generally greater than that associated with surface sediments. It is suggested that this implies a change in process and that the greater tractive force associated with the modern process results from an increase in the average density of flow.

The data and observations of alluvial fans acquired during this investigation suggest that the following characteristics of fans are common to those in the western part of the Great Basin and perhaps to other fans as well; any hypothesis of the formation of the fans must therefore account for these features:

1. The loci of deposition on alluvial fans have shifted from areas well within their catchment areas to the middle reaches of fans, far below the mountain front.

2. Trenches in the apex regions of fans are misfit relative to present flow conditions by reason of their excessive depth, which ranges to about 200 feet.
3. Paired terraces that are continuous with fan surfaces occur within catchment areas and may extend to the divide.
4. Abandoned braided channels occur on fan surfaces at higher elevations than the floors of the modern active channels.
5. Fan surfaces and abandoned channels exhibit desert varnish and (or) weathering stain, whereas in the active channels and on modern deposits these are absent.
6. Hanging fans occur in tributary canyons within the catchment areas, and their surfaces are continuous with both terraces and fan surfaces below the mountain front.
7. The estimated tractive force is greater within the active channels than on fan surfaces.
8. The percentages of clay and organic material in both surface sediments and modern mudflows are extremely small.

The applicability of three general hypotheses of the formation of alluvial fans, namely the evolutionary or Davisian, dynamic equilibrium, and climatic, is discussed. It is concluded that although certain elements of truth are contained within each of these three hypotheses, the climatic explanation can best account for the characteristics cited above. It is thought that the conditions and consequences of one climatic cycle, of the several that have occurred during the formation of the fans, are as follows.

During a period of fan building, aggradation occurred within catchment areas and on fan surfaces, below the mountain front, at correlative levels. Precipitation was more frequent and more widespread than today. In response to this climatic regime, the abundance of both clay and vegetation was greater, and the fan surfaces were more resistant to erosion. Water: sediment ratios were higher, and the medium of transport had a lesser tractive force. Floods of more frequent occurrence spilled over numerous shallow channels onto the fan surfaces, spreading laterally as they emerged from the mountain front. Sediment was deposited over wide areas of the fans as a consequence of these conditions.

Climatic change was manifested by a change from widespread to local precipitation and by a lesser frequency of precipitation. This resulted in a reduction of vegetation and a decrease in the abundance of clay. Both catchment slopes and previous deposits became more easily erodible, thus contributing toward a reduction in the water: sediment ratio of a given flood and the more frequent occurrence of mudflows. These flows of greater density, viscosity, and tractive force deepened the trunk channels in catchment areas and trenched the upper reaches of the fans. Terraces in catchment areas and hanging fans in tributary canyons were thus produced, and the sediment removed in the process was deposited far downfan.

The shift of the loci of deposition downfan, in response to climatic conditions, produces growth of fans at their lower boundaries and a general decrease of slope in the upper reaches. The sediment added in these lower reaches is redistributed and incorporated into the basin fill during the subsequent climatic episode. At this time, lake levels will rise, the trenches will gradually fill with sediment, and aggradation in catchment areas and on fan surfaces will again occur.

The morphology of fans thus tends toward equilibrium with successive climatic fluctuations. The attainment of such equilibrium, however, is a function of the duration and intensity of each climatic episode, and it may or may not occur during these periods. Fan growth is both upward and outward, during alternate intervals, and will continue for some finite period, at the end of which time a given basin will be completely filled with sediment.

INTRODUCTION

Two of the most common generalizations that emerge from the literature on alluvial fans are (1) local floods in arid regions transport sediment to the mountain front and deposit it in a fanlike form as a consequence of a decrease in gradient and the absence of lateral restriction, and (2) alluvial fans are the product of a balance between erosion of the mountains and deposition in adjacent basins.

Although each of the foregoing statements appears to be quite logical, their widespread acceptance tends to obscure the fact that the formation of alluvial fans is poorly understood. The problem can best be considered in terms of its component parts, namely: (1) What is the distribution of sediments on alluvial fans? (2) What is the competence of the transporting medium? (3) What is the dominant mechanism of sediment transport today? (4) Has this or another transport mechanism been dominant in the past? (5) What is the nature and significance of channels on alluvial fans? (6) Is the modern process one of fan building or of degradation? (7) Does this process differ from that of the past? and (8) What is the rate of the modern process?

This study was undertaken to answer these questions and thereby enlarge our understanding of the formation of alluvial fans. A considerable part of the paper is devoted to question (1) above, the distribution of sediments, and to such component questions as: (1) What is the size distribution of sediments on alluvial fans and on the valley floor? (2) Are all the sedimentary parameters mappable? and (3) Which parameters would best serve to characterize the clastic sediments of a bolson environment?

The field investigation covered by this report comprised 8 months during the summers of 1960 and 1961 in eastern California and western Nevada. Quantitative data were obtained in Deep Springs Valley, Calif., and observations of a qualitative nature, particularly of channels on alluvial fans, were made in Fish Lake, Eureka, Saline, Panamint, Death and Owens Valleys. All are in the general vicinity of Deep Springs Valley.

In considering the question of the distribution of sediments within the environment, the first four moments of the size-frequency distribution were studied. If conclusions are based upon study of but a single parameter, such as the mean size or median size, then it must be assumed that this parameter is the most significant. The writer would hesitate to make this assumption, because the significance of other variables, such as kurtosis, is not well understood. For this reason the mean size, standard deviation, skewness, and kurtosis of the size distribution were mapped, thus providing a sound basis for a discussion of sediment mixing and the geologic significance of these parameters.

It was thought that the best approach to the question of possible differences between the modern and former processes lay in consideration of the competence of the transporting medium. Accordingly, a field approximation for this variable was sought and the results mapped. The conclusions that are presented, regarding differences between past and present processes, stem largely from a consideration of these maps and their implications and are supported by other data and field observations.

ACKNOWLEDGMENTS

The project reported on in this paper served as the basis for a doctoral thesis submitted to Harvard University in 1962. The project was conducted largely under the guidance of the late John P. Miller. The many stimulating discussions of the problems involved and the technical assistance which he provided, in both Cambridge and the field, more than satisfied the requirements of an adviser. Of still greater importance, however, was his interest in the study and evident enthusiasm; these provided a source of encouragement that cannot be too strongly emphasized.

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PHYSICAL SETTING

GEOGRAPHY

LOCATION AND TOPOGRAPHY

Deep Springs Valley is located in the western part of the Great Basin, one mountain range east of Owens Valley, Calif., and approximately midway between Mono Lake to the northwest and Death Valley to the southeast. Considering the Great Basin in its entirety, Deep Springs Valley is an atypical basin only by reason of its small size and perfect closure (pl. 8).

The White Mountains border the valley on the west, as far south as Westgard Pass. South of the pass, as well as on the east side of the valley, the name Inyo Mountains is applied. The maximum relief is about 3,100 feet in the vicinity of Deep Springs playa and about 5,600 feet in the north end of the valley. From the northwest corner of the drainage basin (pl. 8) to a point about 10 miles south along the divide, elevations generally range from 10,000 to 11,000 feet. Elevations decrease gradually to 7,300 feet at Westgard Pass, then rise again and average about 8,500 feet around the south rim of the basin. Elevations elsewhere are generally 6,000–7,000 feet at the divide, excepting breaks at Soldier Pass and Piper Mountain which are at 5,400 and 7,700 feet, respectively.

DRAINAGE

The interior drainage of Deep Springs Valley is typical of closed basins. About 40 ephemeral streams enter the valley proper, an area of about 47 square miles. Total drainage area of the basin (pl. 8) is about 144 square miles, exclusive of the valley proper.

The relation between fan area and catchment area noted by Denny (1965) in Death Valley does not hold in Deep Springs Valley. The Wyman-Crooked Creeks drainage area, for example, is nearly equal to the entire area of the valley (pl. 8) and is about three times as large as the Westgard Pass catchment area. Planimetric measurement of the respective fans from either topographic sheets or aerial photographs, however, would show that they are roughly equal in area.

Several reasons exist for this discrepancy between fan area and catchment area. A relation between the drainage area of a watershed and discharge has been demonstrated by many workers, but correlation of fan area with drainage area implies that the planimetric measurements reflect volumes. Clearly, the basic relation involved is that the volume of sediment removed from a given drainage basin should equal that which is deposited below the mountain front. Because differences in relief can occur despite equality of watershed areas, the planimetric measurement of drainage areas may not accurately reflect the volumes of sediment that have been removed. Likewise, such measurements may also fail to reflect the volumes of sediment in alluvial fans, because planimetric areas are a function of the surface slopes, which will vary.

A second consideration involves basin tectonics. It is not reasonable to assume that the downfaulted basin floors are horizontal in the light of much seismic evidence to the contrary. The volumes of sediment in alluvial fans may therefore differ because of basement tilting and subsequent burial of sediment, despite present equivalent surface expression.

ALLUVIAL FANS

Each of the ephemeral streams in Deep Springs Valley traverses an alluvial fan. The fans are larger on the west and northwest sides of the basin than on the southeast. The larger fans generally extend 2–3 miles from their apexes into the valley, whereas fans in the vicinity of Paiute Chute (pl. 8) can be as small as a few hundred feet in extent. The depth of alluvium is unknown, but it exceeds 700 feet in the vicinity of the bench mark in the north end of the valley (pl. 8), where a waterwell of this depth did not reach bedrock.

The general aspect of alluvial fans in Deep Springs Valley is similar to that of fans elsewhere in the Basin and Range province and need not be considered in detail. The most important features of fans, the active and abandoned channels, will be treated below. The approximate limits of the two fans discussed extensively in this report, Antelope Springs and Wyman-Crooked, are shown on plate 8.

Although some fans are clearly defined, merging of adjacent fans is not uncommon. These produce a continuous apron of alluvium and render difficult the task of drawing individual fan boundaries on maps. The merger of the Antelope Springs fan (pl. 8) with those adjacent to it can be seen in figure 85, and the apron of debris that occurs southeast of the playa is shown in figure 86. The darker areas visible owe their color partly to a preponderance of dark quartzite, in contrast to much white marble on the Antelope Springs fan, and partly to the presence of varnish and desert pavement.

An occurrence of color contrast on a single fan that does not result from the presence of a desert pavement is found on the Wyman-Crooked Creeks fan (fig. 87) in the north end of the valley. Although the dark area seen in the background contains a greater abundance of basalt than other parts of the fan surface, much of the color contrast results from the presence of weathering stain or varnish. No closely packed mosaic of stones occurs, however. Both the spacing of particles and range in size is comparable to that which occurs elsewhere on the fan. The presence of varnish in Deep Springs Valley is therefore independent of desert pavements, although both may occur together.

The alluvial fans considered in detail in this report are too few in number to provide valid correlations among such variables as slope, drainage area, and lithology. The data obtained suggest, however, that fans containing coarse debris are generally of steeper slope than, for example, the Antelope Springs fan. The slope of the latter ranges from 3 to 8 percent over much of its surface, and the largest particles present do not exceed 4 feet in intermediate diameter. Slopes on the Wyman-Crooked Creeks fan (fig. 87) range from about 2 to 11 percent and boulders 6–8 feet in diameter are not uncommon. Still greater slopes and larger particles occur on the fans rimming the south end of the basin (fig. 86) and on Paiute Chute (fig. 88), just north of the playa. On the other hand, the reentrant (fig. 89) just north of Crystal Peak (pl. 8) is as much as 16 percent in slope and consists mainly of granitic sand with few large particles. Particle size is, of course, largely a function of the lithology of the source area and the presence or absence of joints, cleavage, and other structural features. Given a source rock that produces no particles larger than cobbles, for example, neither discharge, slope, nor any other factor can enlarge the resulting particle size. Fans consisting of the finer debris contributed by such a source rock, however, are usually of gentle slope.

VEGETATION

The vegetation in the Deep Springs Valley drainage basin reflects orographic control. Sage and other shrubs floor the valley, attaining a maximum height of about 3–4 feet. The active channels on alluvial fans are generally bare of vegetation, but scattered-to-dense sage occurs on the bordering low terraces. In the active channel of the Wyman-Crooked Creeks fan (fig. 87), however, phreatophytes are abundant. This may reflect subsurface seepage beneath the channel, which is derived from the perennial flow in Wyman Canyon.

FIGURE 85.—Aerial view of the Antelope Springs fan showing its merger with adjacent fans. Note the contrast between the color of the sediments within the channels and that of the fan surface, which is darker. The transition from a relatively straight to a braided channel pattern on the fan coincides with the first depositional zone below the mountain front. The former lake level at about 5,200 feet and the linear ridges that mark the most recent expansion of Deep Springs Lake are designated as *A* and *B*, respectively.

Piñon and juniper occur between elevations of about 6,000 and 9,000 feet, together with sage that becomes less abundant with increasing altitude. Above 9,000 feet, the famous bristlecone pine is dominant in the White Mountains. These trees are restricted to some extent by lithology as well as altitude. They are most prevalent in areas where the white Reed Dolomite of Precambrian(?) age crops out; this can readily be seen either in the field or on aerial photographs. The writer noticed that the rather gnarled bristlecones all appeared to twist in the same direction along their axes. This was true whether they occurred on north- or south-facing slopes. From a chance encounter with a botanist in the mountains, who was engaged in a study of the trees, the writer learned that "as everyone knows, the twisting is in response to the Coriolis Force."

CLIMATE

PRECIPITATION

Climate is also a function of elevation, as attested to by records from the two weather stations in the area. One of these stations is in the White Mountains at an elevation of 10,150 feet, in the northwest corner of the drainage basin (pl. 8). The other station is in the valley at an elevation of 5,225 feet. The average annual precipitation in the mountains is about 13 inches, whereas it ranges from 3 to 5 inches in the valley. The maximum 24-hour precipitation for these two stations was 3.10 and 1.50 inches, respectively. It should be pointed out, however, that one cannot state with certainty that the latter values were maxima in the area; they are maxima only for their respective locations and for the period of record.

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FIGURE 86.—Aerial view of the southeastern part of Deep Springs Valley. Note the apron of alluvium, produced by the merger of fans, beyond the playa and to the right. The channel at *A* is entrenched, and lighter colored sediment, indicative of modern transport, can be seen along its course and at the lower boundary of the fan.

During the winter months precipitation in the White Mountains is derived mainly from the residue of cyclonic storms that migrate across the Sierra Nevada and Owens Valley from west to east. The sporadic occurrence and distribution of thunderstorms and cloudbursts in desert areas in summer, however, is well known and Deep Springs Valley is no exception. Although the period of observation in the area was short, comprising only two summers, the writer gained the impression that certain tracks of thunderstorms across the valley were followed with greater frequency than others. The track of six thunderstorms in 1961 was from the southeast, over the playa, and directly across the Antelope Springs fan to the mountains (pl. 8). A second group of storms also arrived from the southeast, but these entered the valley between Paiute Chute and the general vicinity of Soldier Pass. These storms produced some minor precipitation at the valley weather station and proceeded to the Wyman-Crooked Creeks canyon area.

Storms arriving from the north or northeast produced rainfall on Piper Mountain (7,700 ft) and continued in a southerly direction, again passing over the valley weather station.

The remaining storms entered the valley from the southwest, west, or northwest. In these instances, precipitation usually ceased somewhere within the drainage area of the White Mountains and little, if any, rain fell in the valley proper.

Despite the paucity of data on the frequency of track of these storms, two facts are indicated. First, the frequency of precipitation is not everywhere equal within the basin. It will later be shown that this accords with the occurrence of certain hanging fans in tributary valleys. Second, frequency of precipitation in the valley proper is also not everywhere the same, and the total amount of precipitation on basin floors is not great. This may account for the preservation of scattered lake terraces in desert basins; in

FIGURE 87.—Aerial view of the Wyman-Crooked Creeks fan. The darker color of the fan surface, above the dashed line, results from numerous basalt boulders and varnished or stained boulders. The basalt boulders are derived from basalt outcrops within the catchment area and from the dark mass in the background. Note the vegetation in the active channel, which is entrenched to a depth of about 40 feet, and the abandoned distributary channels at higher elevations than the active channel.

addition, it suggests that local runoff on alluvial fans is small compared with that in the mountains.

TEMPERATURE

The difference between the average annual temperature at the White Mountain weather station (pl. 8) and that in the valley is 18°F. Maximum temperatures greater than 105° have not been officially recorded in the valley, but in the absence of an official shade tree, temperatures from 115° to 120° were common. During two rather warm periods sun temperatures of 130° were recorded; ground temperatures are probably even greater.

WINDS

Information on wind intensities and dominant tracks in the valley is not available. The usual diurnal fluctuation of track, up and down canyons, was observed. It is unknown, however, whether any net transport of fine sediment results from this process.

Only three major duststorms occurred during the period of observation. The direction of sediment transport during each storm was from south to north, again prompting the question of frequency of track. Data on silt: clay ratios, to be discussed below in the section entitled "Silt and clay," suggest that the observed track may in fact be preferred in the valley.

SUMMARY OF GEOLOGY

The geology in the Deep Springs Valley area has been treated by Knopf (1918), Miller (1928), Nelson (1962, 1963), and others. Nelson conducted a detailed stratigraphic study and mapped the geology of the area. Much of the information to follow stems from his work complemented by personal observations. An outline of the geology, pertinent to this report, is shown on plate 8.

In the White Mountains the stratigraphic section consists of 21,000 feet of Precambrian(?) and Lower

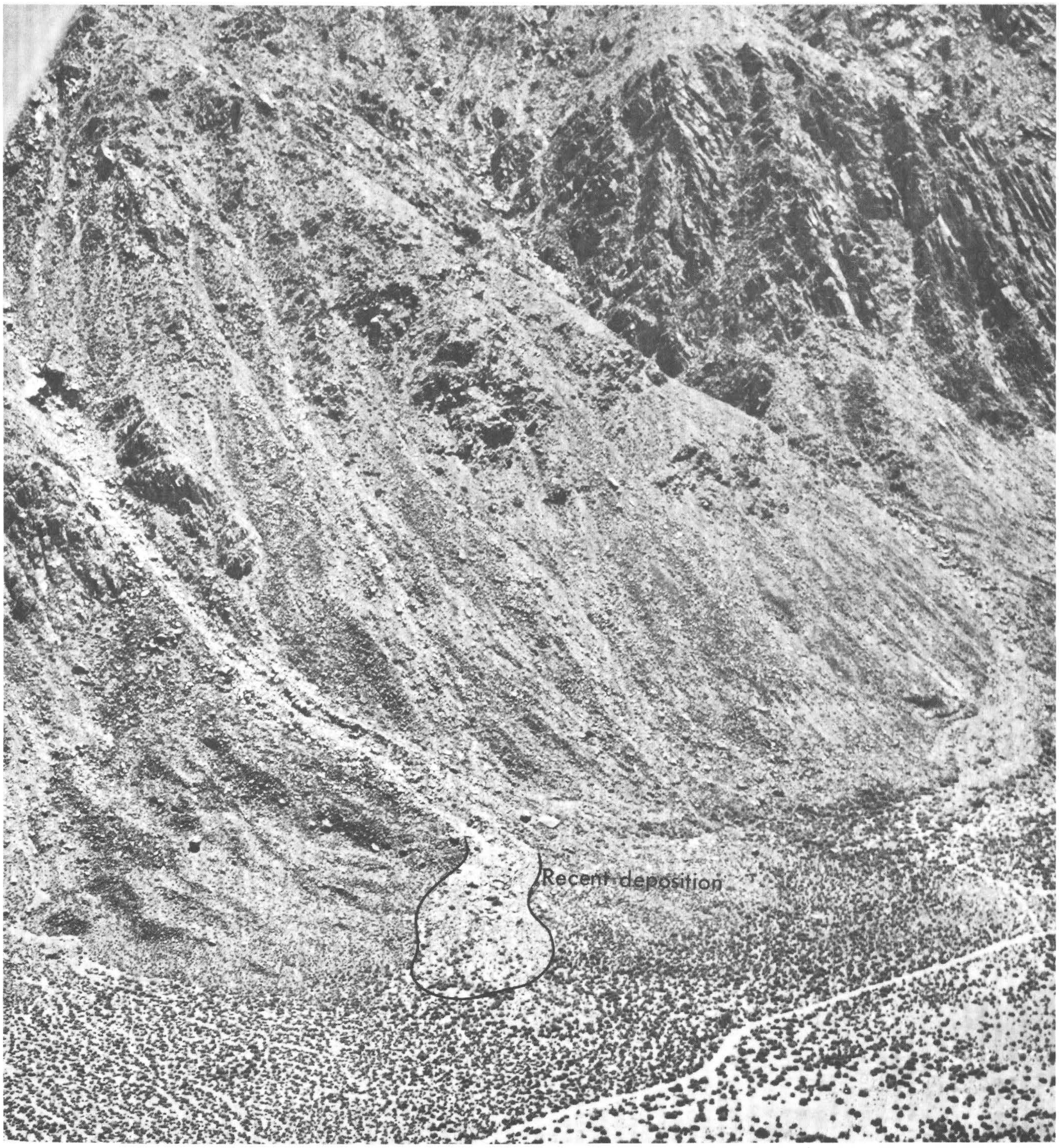

FIGURE 88.—Aerial view of Paiute Chute. Note the prominent active channel and the bordering levee walls. Several similar chutes occur in this general area, on the east side of the valley. The area of recent deposition on this young fan is below the termination of the active channel and at the lower margin of the fan, bounded as shown in the photograph.

FIGURE 89.—Aerial view of the Crystal Peak reentrant. Crystal Peak is the small plug to the left of the dashed line that delineates the steeply sloping reentrant. Surface slopes are as much as 16 percent.

Cambrian strata and about 1,600 feet of Middle Cambrian sedimentary rocks; the latter are relatively restricted in outcrop area. Of the 10 formations described by Nelson (1962), only those pertinent to a discussion of the distribution of clastic sediments in the valley are mentioned here.

The sedimentary sequence has been folded, intruded by a granitic batholith and one or more stocks, and faulted. Structurally, the White Mountains are a southerly plunging anticline within the drainage basin considered. Because the fold axis trends nearly due south, whereas the topographic trend is southeast, the adjacent syncline to the east is exposed farther south in the Inyo Mountains. Erosion of the fold in the White Mountains has therefore revealed the oldest rocks in the section. The structural relationships become apparent in Wyman Canyon (pl. 8) where, despite considerable local complexity, repetition of strata to either side of the Precambrian(?) heart of the fold can be seen.

The formations of Precambrian(?) age encountered in Wyman Canyon, which cuts across the fold axis, are the Wyman, Reed, and Deep Spring. The Wyman Formation is greater than 9,000 feet thick (base unexposed), and the Reed Dolomite and Deep Spring Formation are 2,000 and 1,500 feet thick, respectively. These units crop out over a distance of about 6 miles along the canyon. The Wyman Formation consists of crumpled and contorted schists, phyllites, shales, and quartzites which are mottled gray, brown, or black and are generally banded. The overlying Reed Dolomite is a white to light-bluish-gray dolomite, and the younger Deep Spring Formation consists mainly of reddish-brown or yellowish-brown sandstones. South of Wyman Canyon the Wyman Formation gradually pinches out because the fold plunges in this direction. It comprises only a minor percentage of the rock types within the Antelope Springs catchment area, whereas the Reed Dolomite and Deep Spring Formation crop out in abundance.

A Cretaceous(?) granitic intrusive, of roughly 5 square miles in outcrop area, has pierced the sedimentary rocks between Antelope Springs and Wyman canyons and a contact metamorphic sequence occurs. South of the Antelope Springs drainage area the younger, Lower Cambrian sedimentary rocks, crop out. Dark sandstones and shales with varying amounts of limestone predominate; this assemblage, in varying proportions, continues around the south end of the basin. The contact between the sedimentary rocks and the granite occurs east of the playa, and the Inyo Range on the east side of the valley is granitic from this point northward.

Scattered remnants of Tertiary basalt and an underlying white friable tuff crop out along Crooked Creek, to the north of the Wyman-Crooked Creeks fan (fig. 87), and in the northeast corner of the basin.

The foregoing summary of the geology of the area indicates that certain rock types are sufficiently distinctive and restricted in area to serve as indices of probable source area at sampling stations in the valley. An abundance of sedimentary rocks derived from the Wyman Formation, for example, unquestionably indicates transport from the Wyman-Crooked Creeks system (pl. 8) in the northwestern part of the basin even if it occurs near the playa. Knowledge of the distribution of basalt remnants and color and other distinctions among specific basalt sources also proved useful during this study, particularly in treating the problem of sediment mixing within the basin and attendant effects upon the sedimentary parameters. Although several fine distinctions were drawn during the study, for convenience the lithologic data on the valley sediments have been summarized under the gross headings of granitic rocks, sedimentary rocks and basalt (table 2).

ASPECTS OF PLEISTOCENE GEOLOGY

No evidence of extensive glaciation in the White Mountains, within the drainage basin considered, was noted by the writer. This may not be true for the part of the range farther north, however. During the Pleistocene the snowline was undoubtedly much lower than it is today, but even at the headwaters of Wyman and Crooked Creeks (pl. 8) at elevations of 10,000–11,000 feet, evidence of cirques or moraines is lacking. A buried soil occurs in the headwaters of Crooked Creek, and signs of frost action were noted at several localities; these features represent the only evidence of Pleistocene climates in the mountains. An extensive snow cover and local ice caps probably existed in the higher parts of the range, however. If so, then combined with greater precipitation and lower temperatures, these local caps and the snow cover must have contributed to a much greater runoff and discharge than obtain today.

Miller (1928) was the first to discuss the Pleistocene lake in the valley and its probable extent. He estimated that its maximum depth was on the order of 400–500 feet. Given present elevations, a lake 500 feet deep would nearly cover the entire valley floor. The highest known stand of the lake was at about 5,200 feet, which accords with gravel deposits around the south end of the valley. The 5,200-foot contour follows the base of the low bedrock hill to the right of the Antelope Springs fan (fig. 85). Color contrasts also suggest a former high stand at this level. If the lake did not rise above this level, however, then according to present elevations, it could not have reached the northern end of the valley. The bench mark at 5,220 feet (pl. 8) would represent the northern limit of the former lake. A 5,200 foot-level would appear to insure that the lake did not overflow the basin; but present topography is not sufficient proof of this, and a former connection should be sought.

The former connection of many lakes in the Mohave region with the Pleistocene Owens Lake, as a result of overflow from one basin to another, is well known; the subject has been reviewed by Blackwelder (1954), among others. Westgard Pass, however, is at an elevation of 7,300 feet (pl. 8), sufficient to ensure that Deep Springs Lake was not part of this chain by way of Owens Lake. If a former connection in some other direction is sought, then the "new map" showing the distribution of pluvial lakes by Feth (1961) is of no greater assistance than the original compilation by Flint (1957, p. 227). It is probable that every basin in this region contained a lake, although not all overflowed their basin. A field reconnaissance of the Great Basin could provide much of the missing data.

Deep Springs Valley is adjacent to Eureka Valley, to the east, and Fish Lake Valley, to the northeast. A direct connection between Fish Lake Valley and Deep Springs Valley is possible, but field evidence suggests that the two may also have been connected by way of Eureka Valley. Horse Thief Canyon, which connects Fish Lake and Eureka Valleys, was traversed, and many solution pockets were noted by the writer along the base of the canyon walls. The divide between these two valleys is within Fish Lake Valley at an elevation of about 5,280 feet, and Horse Thief Canyon runs downgrade over its entire length into Eureka Valley.

Soldier Pass (pl. 8), which joins Deep Springs and Eureka Valleys, was also traversed. A carbonate-cemented pebble conglomerate probably formed in Pleistocene time and now largely stripped away, crops out along the floor and walls of the pass. The gravel fraction consists entirely of angular granitic fragments. The beds dip to the west, suggesting postlake uplift and tilting. Other evidence of uplift during Recent time can be seen at several localities on the east side of Deep Springs Valley. In the vicinity of the present playa another carbonate-cemented conglomerate crops out above the valley floor; scarps trend abruptly across alluvial deposits, and springs and bogs occur in this zone. Nelson (oral commun., 1961) has suggested that a major fault follows the zone east of the playa to the north and then turns east through Soldier Pass. Differences in elevation within the Inyo Mountains on either side of the pass are in accord with this suggestion.

Uplift of about 200 feet in the vicinity of Soldier Pass appears reasonable. If this is true, then the pluvial Deep Springs Lake overflowed through the pass into Eureka Valley. A small isolated playa occurs in a reentrant of the Inyo Mountains, on the Eureka Valley side of the divide, at the same latitude as the Deep Springs playa. Overflow from the pluvial Deep Springs Lake may have spilled into this sink, which has subsequently been uplifted.

Another conglomerate locality in the north end of Deep Springs Valley also indicates postlake uplift and tilting. These carbonate-cemented conglomerates occur near the most southern of the group of small drainage systems just southwest of Piper Mountain (pl. 8). The beds dip to the west, and in a few places they are as much as 200 feet above the present valley floor. They represent an absolute criterion for uplift, because the gravels are of lithologic types derived solely from the Wyman-Crooked Creeks drainage area across the valley. The gravels are well rounded and range in size from pebbles to large cobbles. This occurrence suggests that sediment from the Wyman-Crooked Creeks system must formerly have been transported directly across the valley floor in a more easterly direction than today. Former channels on the Wyman-Crooked Creeks fan also trend in this direction. Uplift and tilting, following withdrawal of the lake, produced the present southerly trend leading to the playa.

In summary, it is considered probably that Deep Springs Lake did extend over the entire valley floor, but the absolute depth or extent of this lake cannot be judged from present topography. This example of postlake uplift probably holds true for many of the western basins. Traverses of likely interbasin canyons provide a better field test of possible former connections between lakes than does examination of present elevations of accordant levels.

BASIC DATA

The quantitative data obtained in Deep Springs Valley are listed in tables 1 through 6 and were used to construct the various maps presented in this report. Sampling stations in the north end of the valley and on the Antelope Springs fan are shown in figures 90 and 91, respectively. The data include (1) measurements of local slope and maximum particle size at 496 stations; (2) 12,400 determinations of pebble size, roundness, and lithology; (3) sedimentary parameters of 90 samples of the granule-to-clay size fraction; and (4) approximate clay-mineral ratios for 43 samples. A planetable map of the active channel on the Antelope Springs fan (pl. 9) was constructed from data from about 600 stations on a scale of 1 inch = 100 feet.

Sampling stations in the north end of the valley are 1,000 feet apart, based upon an arbitrary grid system. The bench mark at 5,220 feet (pl. 8) was used as point 0.0 (fig. 90). Stations were located by means of pace and compass, or by odometer and compass if the point could be reached by jeep. Although the odometer was checked over measured miles for accuracy and tire pressures were not overlooked, some lack of precision must inevitably result because of the roughness of terrain. The same is true of paced distances in the area. The resultant inaccuracies of the maps, however, do not detract from the general trends shown.

Sampling stations on the Antelope Springs fan are located along contours that differ in elevation by 50 feet, determined by altimeter, and are 200 feet apart on each contour line.

Stations on Paiute Chute (fig. 88) and on the miniature fans in Westgard Pass, treated below, are 100 and 5 feet apart, respectively. Bearings taken from each station to the fan apex permitted the plotting of sampling points. This procedure was also followed at each station on the Antelope Springs fan (fig. 91).

The largest particle within a 50-foot radius of the sampling stations was located by observation, and the length of the intermediate axis was measured to the nearest tenth of a foot. Both the search for the largest particle and its measurement required some digging in many places. Slopes were measured to the nearest one-half percent with an Abney level over a 100-foot reach between the particle and the apex region of the fan. Because paths of sediment transport do not necessarily coincide with a straight line from the fan apex to a given particle, the 100-foot reach chosen was that which appeared to be the most probable transport path from field observation. If a particle occurred within a channel or boulder train, for example, then slope was measured along these trends regardless of deviation in direction from the apex. The rock types of the largest

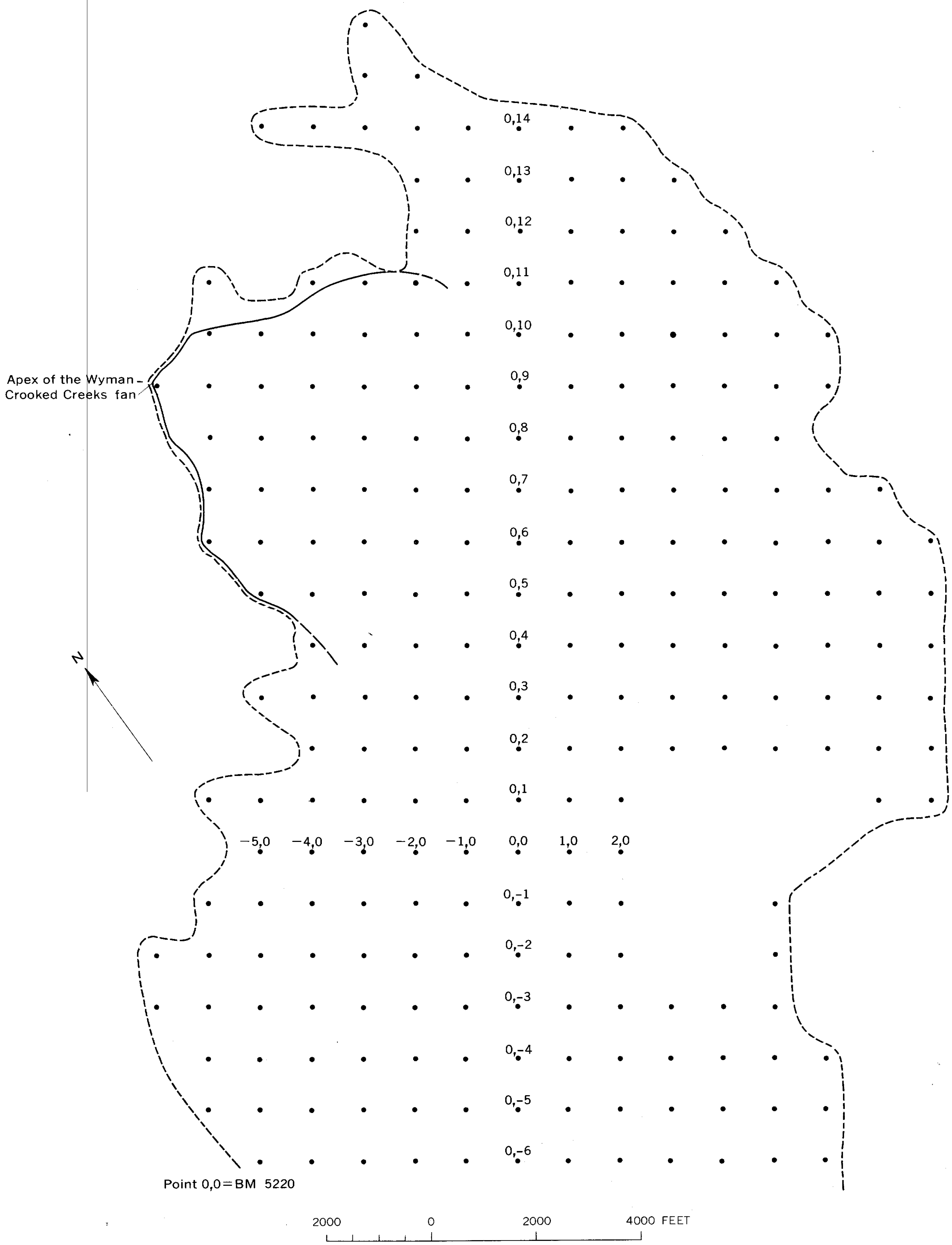

FIGURE 90.—Map showing sampling stations in the north end of Deep Springs Valley.

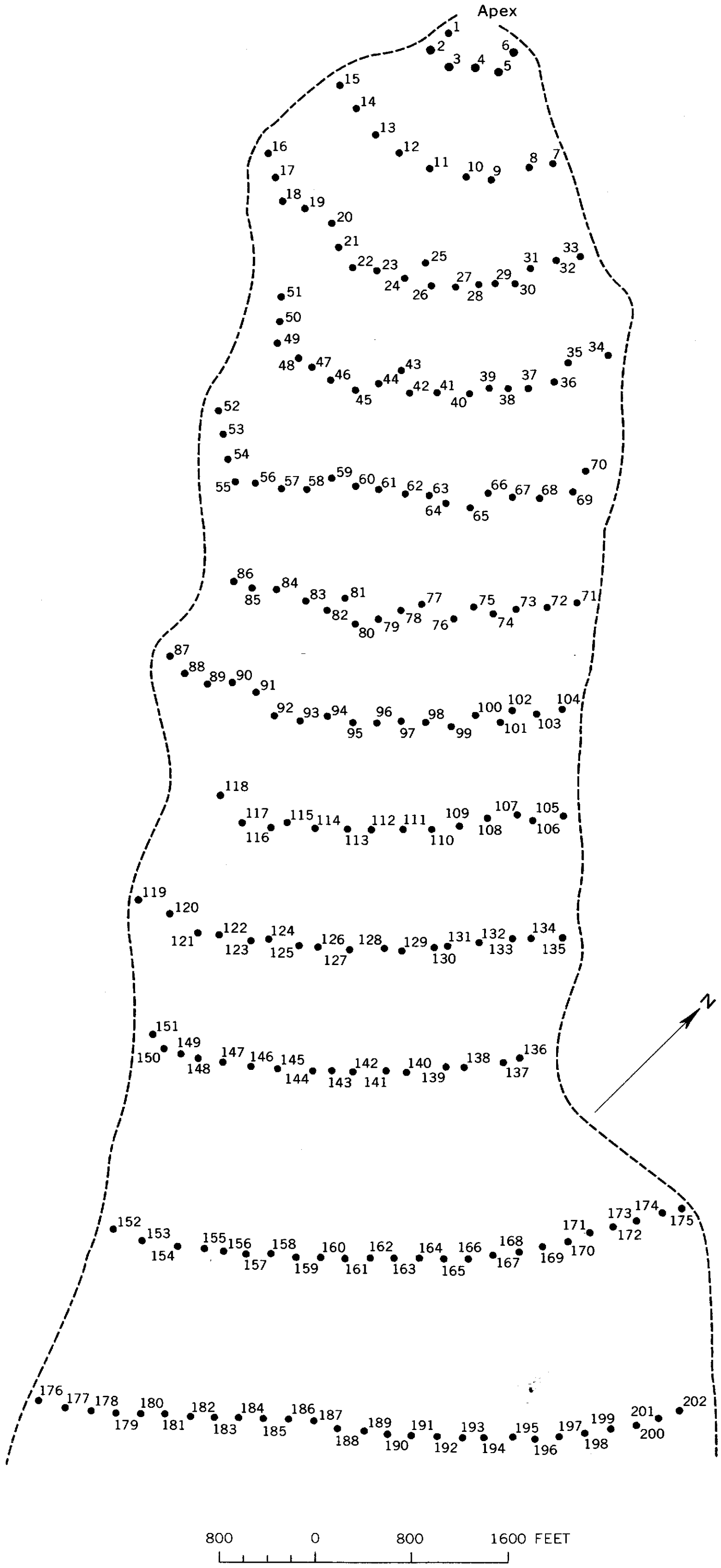

FIGURE 91.—Map showing sampling stations on the Antelope Springs fan.

particles were also obtained, but they are not included in the tables.

Data on pebbles were obtained only in the north end of the valley. At each sampling station (fig. 90) 50 pebbles were selected. These were obtained by pacing off a rough square, of about 12 paces to a side, and picking up the pebble nearest to the toe of the boot with each pace. Unlike the method described by Miller (1958) for sampling the Sangre de Cristo streams, the pebbles were not selected with eyes averted because of possible objection to the entire procedure by reptiles and other desert denizens.

For each pebble, the smallest, intermediate, and largest axis was measured with a vernier caliper to the nearest tenth of a millimeter, and the lithology was determined with a hand lens. The roundness of each pebble was determined by visual means and recorded as angular, subangular, subrounded, rounded, or well rounded. These categories were later assigned values of 0, 1, 2, 3, and 4, respectively, and the mean roundness for each rock type and for the entire sample was calculated.

In order that these procedures not appear to be open to the influence of preconceptions, it might be well to point out that the time requirements and other factors insured objectivity. The number of stations sampled per day was understandably small, because of the amount of data gathered. This prevented consecutive sampling from the valley floor to the mountain front along sampling lines. Also, as a matter of convenience, sampling lines were often left uncompleted and were returned to several days later when sampling along perpendicular lines. These factors alone would be sufficient to insure that one would not select larger or smaller particles of a given degree of roundness according to some concept of what the data "should be" at a given point. In addition, no mean value was computed until all stations in the north end had been sampled. The maps of the several variables presented indicate that the arbitrary sampling grid cuts across most trends at odd angles; this would not result from subjective sampling.

Samples of the granule-to-clay fraction were obtained from a total of 290 stations. Excepting grab samples from two mudflows, the upper 4 inches of sediment within a 5-foot radius of each point was obtained with a can or small shovel and successively quartered until the remainder would fill a 4- by 6-inch sample bag. Much care was expended to insure that the finer fraction not be lost during this procedure. Ultimately, 90 samples were chosen for analysis. These include 66 from the north end of the valley, 19 from the active channel on the Antelope Springs fan (fig. 9), 2 from Paiute Chute, 2 from a mudflow in Owens Valley, and 1 from a thin mudflow in Wyman Canyon.

These samples were split in order to obtain about 50 grams of each and then wet sieved through a 62-micron mesh. After oven drying, the granules were separated from the sands, and the latter were sieved with U.S. Standard Sieves; this provided a separation at half phi intervals. The less than 62-micron fraction was washed into settling tubes, and any additional sediment of this size that was obtained as a pan fraction during sieving of the sands was added. Silt and clay were separated at the 4-micron boundary by successive decantations. Time intervals for decantation were determined by settling velocities at the temperatures that prevailed. Because of the height of fall, which was 25 cm, the average time interval was about 4.25 hours. Most samples required about 10 decantations, but this varied, of course, with the amount and proportions of the sediment. The <4-micron fraction was drawn off, and after a few days were allowed for settling, a concentrated slurry was obtained and clay-mineral mounts were made by sedimentation on glass slides. The silt fraction was oven dried and weighed, and the weight of clay was determined by difference.

The mudflow samples were boiled in chlorox several times before separation of the size fractions, and the weight of organic matter was later determined by difference.

The weight percent of each class interval, from granules to clay, was calculated and the results cumulated. The cumulative percents were then plotted on arithmetic probability paper against phi diameter, and the values necessary for determination of the statistics used were read from each curve. The statistical measures used in this report to characterize the size distribution of the granule-to-clay fraction are those suggested by Folk and Ward (1957). They are as follows:

$$\text{Mean size } (Mz)=\frac{\phi_{16}+\phi_{50}+\phi_{84}}{3},$$

$$\text{Inclusive graphic standard deviation } (\sigma_I)=\frac{\phi_{84}-\phi_{16}}{4}+\frac{\phi_{95}-\phi_5}{6.6},$$

$$\text{Inclusive graphic skewness } (Sk_I)=\frac{\phi_{16}+\phi_{84}-2(\phi_{50})}{2(\phi_{84}-\phi_{16})}+\frac{\phi_5+\phi_{95}-2(\phi_{50})}{2(\phi_{95}-\phi_5)},$$

and

$$\text{Graphic kurtosis } (Kg)=\frac{\phi_{95}-\phi_5}{2.44(\phi_{75}-\phi_{25})}.$$

These measures represent a reasonable compromise between including a sufficiently large percentage of the size distribution to obtain the information desired and attempting to minimize the effort expended in attaining it. McCammon (1962) has recently reviewed the efficiency of various statistics of mean size and sorting. He shows that the measure of mean size (Mz) used in this report has an efficiency of 88 percent, whereas $\frac{\phi_{16}+\phi_{84}}{2}$, for example, has an efficiency of 74 percent. The respective efficiencies of inclusive graphic standard deviation (σ_I) and $\frac{(\phi_{75}-\phi_{25})}{1.35}$ are 79 and 37 percent. To further increase the efficiencies of the statistics chosen, it would be necessary either to include additional percentiles within the range considered or to increase the range, or both.

CLASTIC SEDIMENTS

The clastic sediments of Deep Springs Valley range in size from boulders to clay. At a given point, therefore, a volume sufficient to fill a small truck might be required in order to investigate the size distribution of a single sample. For this reason, and also because the implications of the size distribution are the primary concern, it was thought best to treat three discrete fractions of the total sediment. These are the largest particles, the pebble fraction, and the granule-to-clay fraction. Although this treatment is not entirely satisfactory, several considerations show it to be useful. First, separate treatment of the largest particles permits a discussion of competence; second, the weight percentages determined for such significant size classes as silt and clay are maxima; finally, the alternatives are in some respects still less satisfactory. If particles that occur at a given tape interval are measured, for example, there is an inherent tendency to skew the distribution toward the larger particles because pebbles will invariably be selected in preference to granules when both are present.

LARGEST PARTICLES

The 496 particles that represent the maximum size present at each station are listed in table 1 together with the slopes on which they occurred. Although large particles occur more frequently near apex regions of fans, they are not restricted to these areas. Figure 92 shows the relation between maximum size and slope for stations in the north end of the valley (fig. 90). The only station omitted from figure 90 is −7, 9 where a 16.5-foot particle occurred on a slope of 11 percent. The next largest size present at these stations was slightly <8 feet in intermediate diameter, and for reasons of convenience the ordinate was not extended beyond 10 feet.

The scatter of points on figure 92 is apparent. This can partly be explained by considering that several fans are represented and that variations of lithology, and especially joint spacings, will produce blocks of unequal size at the source. A plot of slope and maximum particle size for 202 stations on the Antelope Springs fan (fig. 93) shows a rather good trend, except for the particles on gentle slopes.

Blissenbach (1952, 1954) claimed that the relation between maximum particle size and slope was arithmetic for fans in Arizona and that the decrease in size downfan was uniform. His data were scanty, however; even the data for Antelope Springs (fig. 93), consisting of a limited size range, would not plot well on an arithmetic scale. In this report, the scatter of points is thought to be of greater significance than the general trend because understanding of process is the goal.

Particles larger than 1 foot in diameter occur in areas of zero slope, and particles larger than 5 feet in diameter occur on slopes of 2 percent in the north end of the valley (fig. 92). On the other hand, particles as small as 0.2 foot in diameter are the maximum size present on slopes as steep as 14 percent. For the Antelope Springs fan, where the range of both size and slope is more restricted (fig. 93), particles half the size of the largest present occur on slopes as gentle as 2 percent. Conversely, particles as small as 0.1 foot in diameter can represent the maximum size on slopes as great as 5 percent, although this fan exhibits maximum slopes of only 8 percent. It might be argued that the examples cited are the extremes of the distribution and therefore of little consequence. Regardless of the trend of the means, however, the extreme values require explanation.

The fact that boulders have been reported on playas (McAllister and Agnew, 1948; Kirk, 1952; Clements, 1952; and others) clearly indicates that the phenomenon of large particles in areas of zero slope is not restricted to Deep Springs Valley. Guilcher and Cailleux (1950) observed pebbles moving along an icy road in Denmark under the impetus of the wind. This observation led to a wind tunnel experiment by Grove and Sparks (1952) who duplicated the phenomenon. Extrapolation of their data by Schumm (1956) indicated that a velocity of 122 miles per hour would be required to move a rock weighing 1 pound across ice. Although the movement of boulders on playas is usually considered apart from any question of transport on fans, it is obvious that any boulder must first have been moved down a fan if it occurs on a playa. An explanation for the fluctuation of maximum particle size in a downfan direction will be offered below (p. 167).

THE PEBBLE FRACTION

The distribution of the mean pebble size in the north end of Deep Springs Valley is shown on figure 94. Although several trends are apparent, the most striking feature of the map is fluctuation in size, which is represented by closures around high and low values.

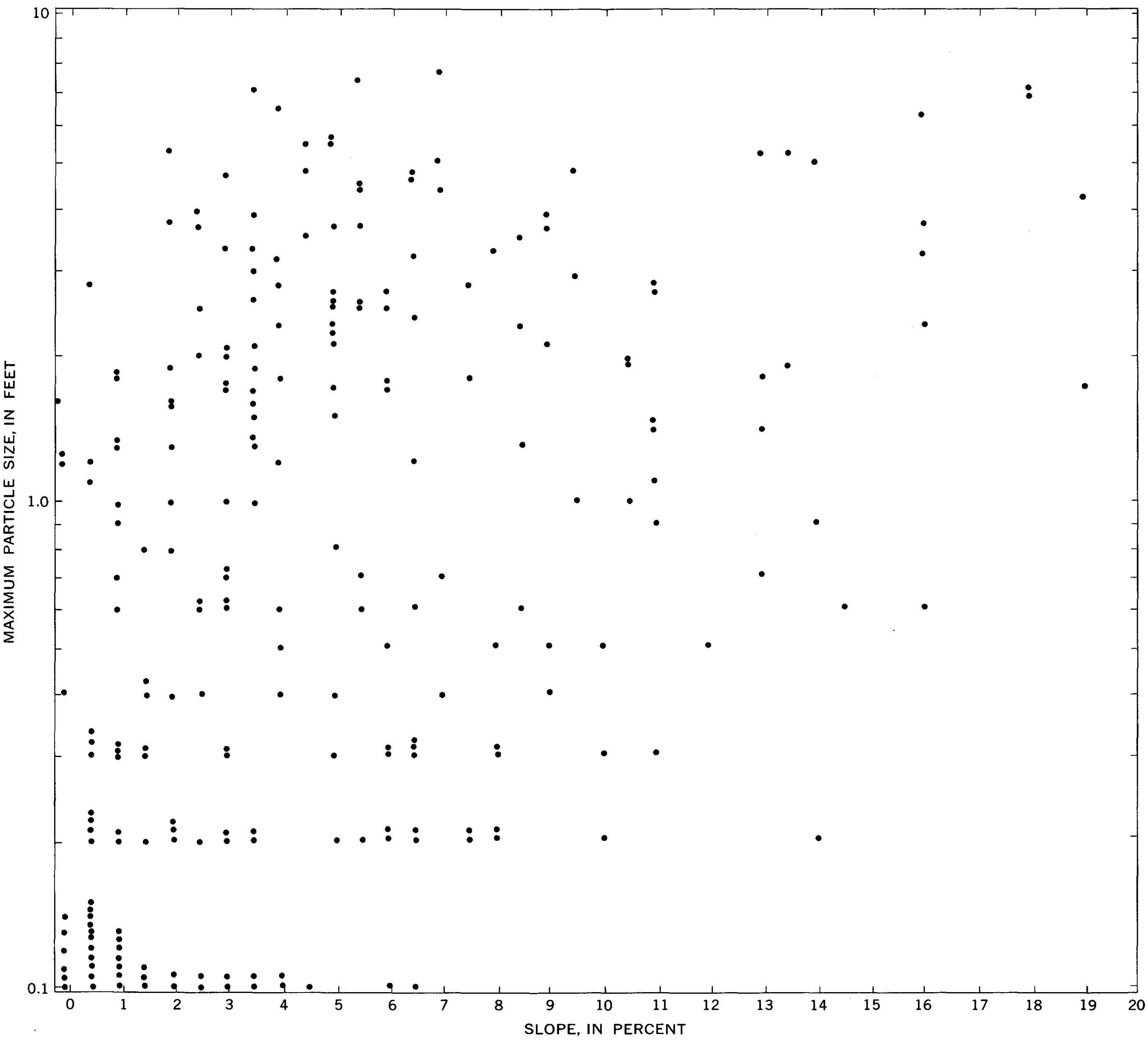

FIGURE 92.—Graph showing the relation of maximum particle size to slope at stations in the north end of Deep Springs Valley.

Some closures around high values are the result of sampling in a wash. This is true, for example, of the 34-mm high in the southwest corner of the map and the 28-mm high farther north. This 28-mm high defines the shallow wash shown in figure 95. Other high and low values result from extremes at only one or two points; this deserves further consideration.

First, it should be acknowledged that two individuals will seldom, if ever, produce identical isopleth maps from the same data. Nevertheless, there are limits that will restrict the possible differences. If the pebble data (table 2) were in accord with a uniform decrease in size downfan, then isopleths would be forced to conform with the basin boundary to some extent, or trends on individual fans would at least emerge more clearly. Neither is true of the map under consideration. Second, the question can be raised whether the mean or the extremes of the distribution is to be emphasized. If stress is to be placed upon average or mean trends, then the variations can be reduced by constructing a moving average map. Choice of a larger isopleth interval would also achieve this end. As with the largest particles, however, the fluctuations in size probably are significant and for this reason no attempt should be made to obscure the fact that they occur by smoothing the data. Each point on the map represents the average of 50 measurements; the map is therefore thought to depict accurately the distribution of pebble size.

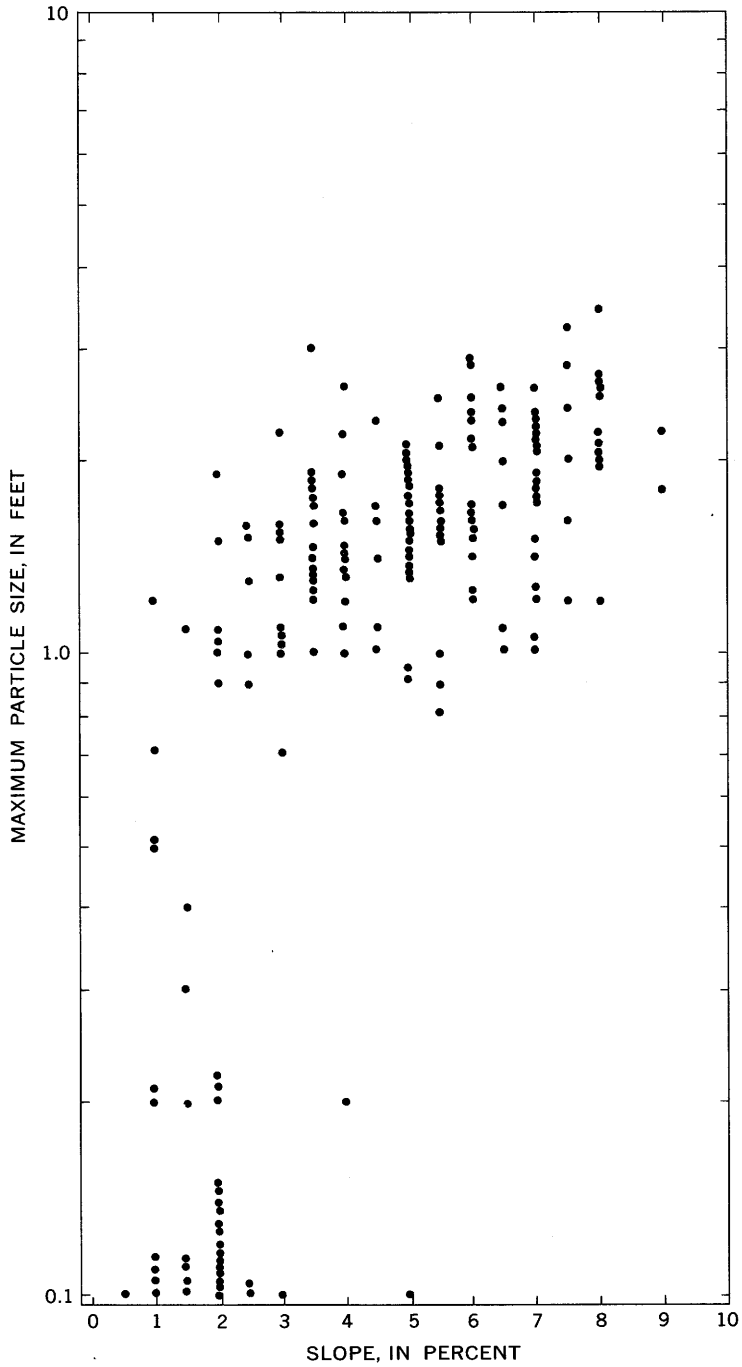

FIGURE 93.—Graph showing the relation of maximum particle size to slope at stations on the Antelope Springs fan.

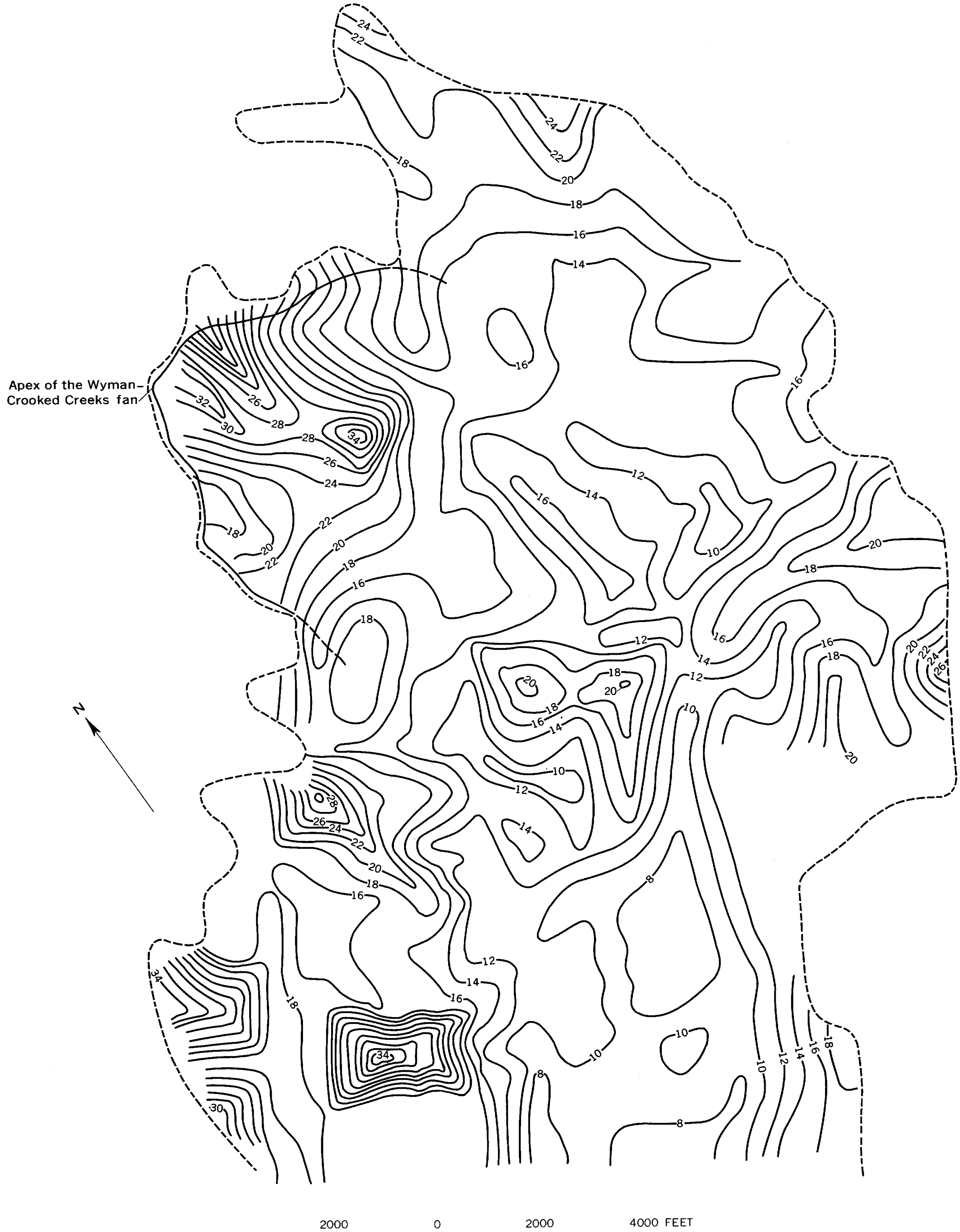

FIGURE 94.—Map showing mean pebble size (in millimeters) in the north end of Deep Springs Valley. Dashed line represents approximate bedrock basin boundary.

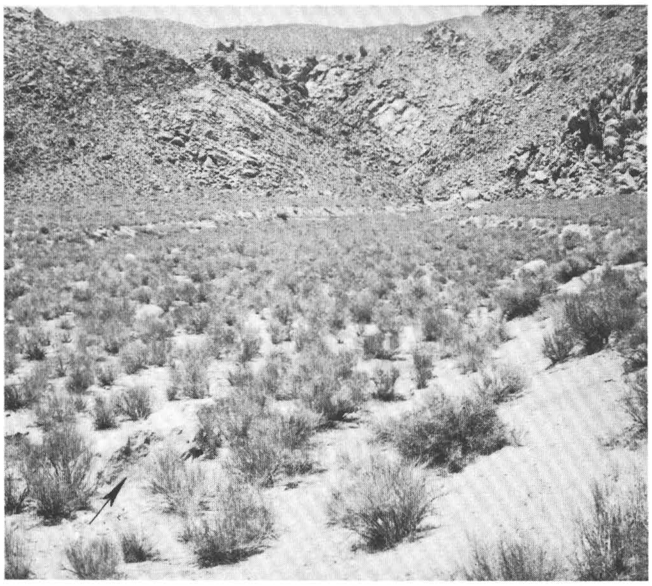

FIGURE 95.—View of a shallow wash in the Crystal Peak area. The wash at this point is approximately 193 feet wide and 2 feet deep. The sediments are predominantly of sand size, consisting of granitic weathering products. The boulder in the foreground (arrow), however, is about 4 feet in diameter.

The roundness (fig. 96) and the lithology (fig. 97) of pebbles in the north end of Deep Springs Valley are correlative. Fields of high mean roundness outline two main lithologic groups. These are the metamorphic and sedimentary types that issue from the Wyman-Crooked Creeks canyons (pl. 8) and the basalts derived from the cap on Piper Mountain and the low tilted masses at its base. The latter can be seen in figure 98. The color contrast at the base of the slope roughly coincides with the 50-percent contour in this area on figure 97 and with the roundness field of 1 on figure 96. Quartz and feldspar pebbles, derived from coarse grained granites and a few pegmatites, and scattered pebbles from aplite dikes are included in the plutonic rock percentage. These pebbles, in addition to the angular granite pebbles that occur close to their source areas, are more angular than the sedimentary, metamorphic, and volcanic rock types. For this reason the mean roundness reflects the lithologic distribution. The occurrence of closures on both the roundness and lithology maps is worthy of note. These reflect fluctuations in the distribution similiar to those found for mean size (fig. 94).

It is apparent from the roundness and lithology maps that a zone of mixing of sediments occurs along the central and east-central parts of the basin. Sediments from the Wyman-Crooked Creeks system, the northern and northeastern parts of the basin, and from the granitic Inyo Mountains on the east all merge in this area. The sedimentary parameters of the granule-to-clay fraction, to be discussed below, also reflect this mixing of sediments.

THE GRANULE-TO-CLAY FRACTION

GRANULES

In a review of the various size grades, Pettijohn (1957, p. 47) notes that the results of many studies indicate that granules (2–4 mm) are less abundant in nature than are other size groups. He also states that there appears to be a deficiency of very coarse sand as well. The basis for these statements is the fact that a sedimentary deposit consisting of both gravel and sand generally exhibits a distinct mode for each of these two fractions. The break in the frequency curve, between these 2 modes, falls within the 1- to 4-mm size range.

Several attempts have been made to explain the frequency minimum. Sundborg (1956, p. 191–194), for example, argues on hydraulic grounds that particles between 1 and 6 mm are those most readily moved when bed transport commences and are the last to come to rest when it ceases. According to this argument, granules are scarce in nature because they are in nearly constant motion, relative to other size classes, and therefore simply wear out. Kagani (1961) also advances a hydraulic explanation and, in addition, advocates placing size-class boundaries at all conspicuous frequency minima. This would, in effect, tend to remove granules entirely from consideration as a distinct size class.

An investigation of the clastic sediments in Deep Springs Valley, Calif., provided a good opportunity to consider the problem of the scarcity of granules. Because sediment transport is typically both brief and intermittent in a bolson environment, the conditions postulated by Sundborg (1956) do not pertain. A simple test of the hydraulic theory can therefore be applied; granules should occur in no lesser abundance than other size classes if the theory is correct. The basic data for the following discussion are listed in table 5; the weight percentages of granules are listed in table 3.

The weight percentage of granules in the granule-to-clay size fraction was determined for 90 samples. In addition to 19 samples from the Antelope Springs active channel (pl. 9), 71 samples were studied. These include 66 from the north end of the valley, 2 from Paiute Chute, and 3 from mudflows. Of these samples, 24 were unimodal and 47 were bimodal. In the unimodal group no mode coincided with the granule, very coarse sand, fine-sand, and clay sizes. Granules, however, occurred as the primary mode of 4 and the secondary mode of 5 of the bimodal samples. Among the

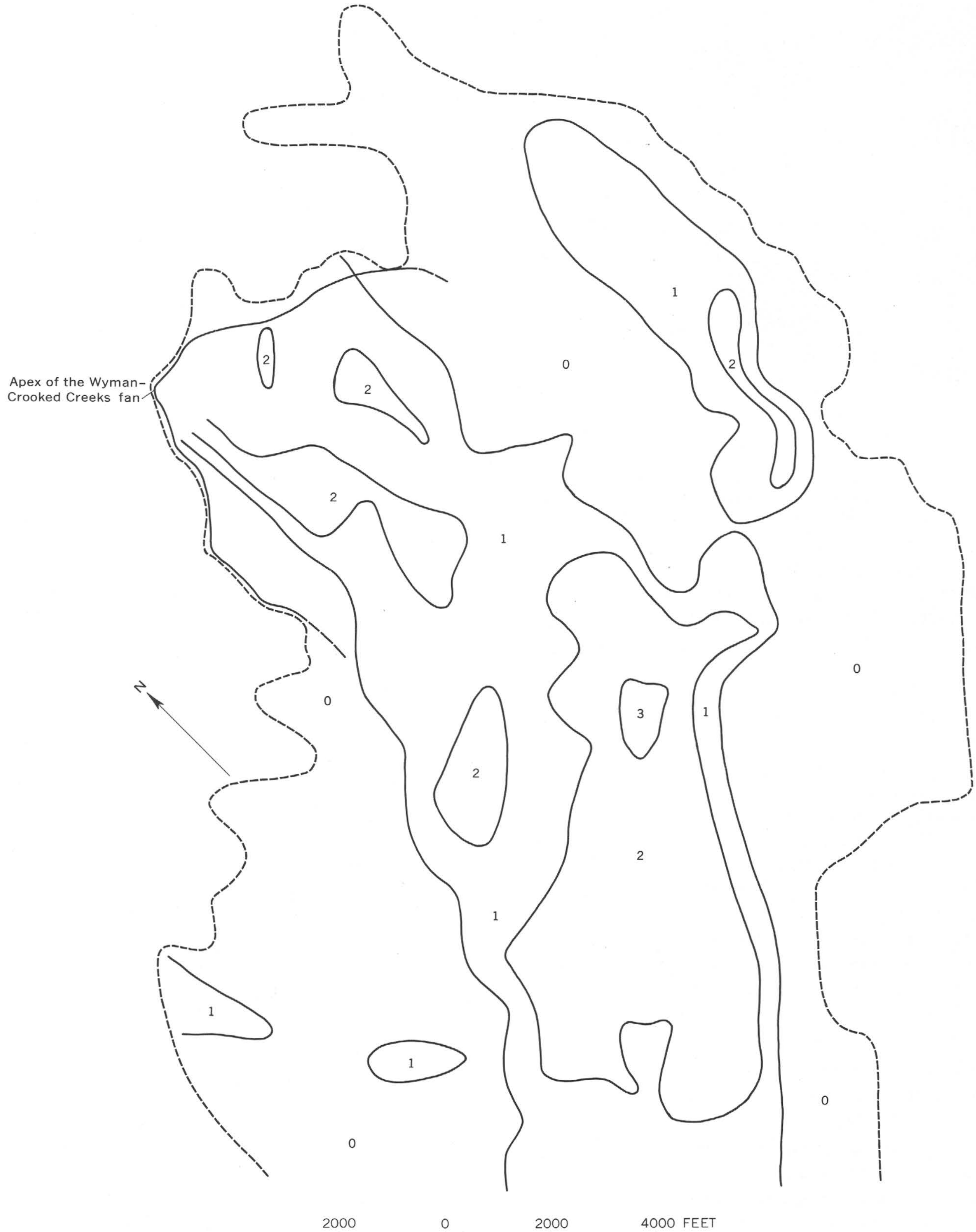

FIGURE 96.—Map showing mean pebble roundness in the north end of Deep Springs Valley. Dashed line represents approximate bedrock basin boundary. Isopleth values: 0=angular; 1=subangular; 2=subrounded; 3=rounded; 4=well rounded.

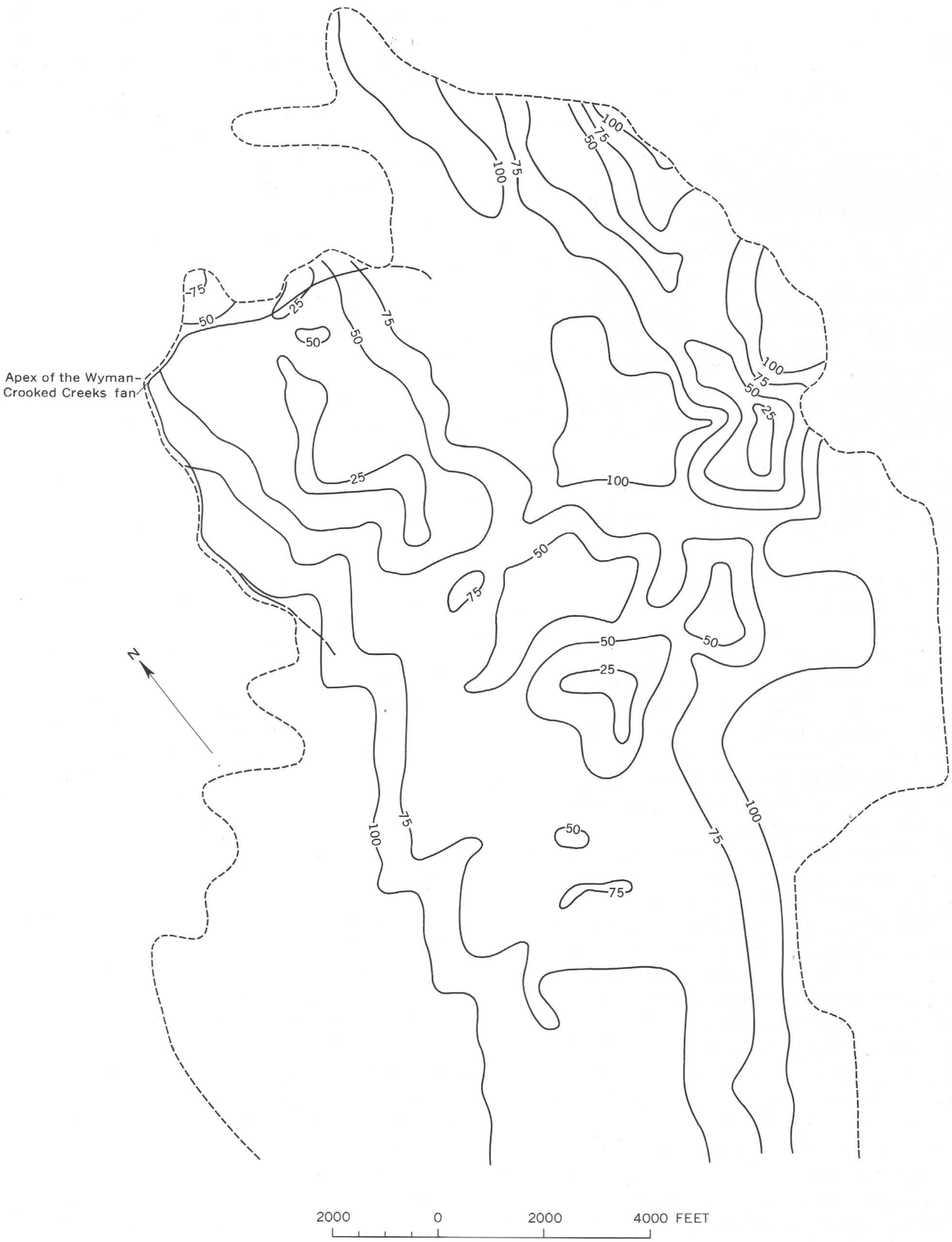

FIGURE 97.—Map showing pebble lithology in the north end of Deep Springs Valley. Dashed line represents approximate bedrock basin boundary. Values represent the percentage of plutonic rocks, including quartz and feldspar. Isopleth interval 25 percent.

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FIGURE 98.—View of the basalt mass and small fan at the base of Piper Mountain. The roundness and lithology of pebbles in the north end of the valley are greatly affected by the basalt particles derived from this area. The lighter colored debris in the foreground is mainly granitic and is derived from the northernmost point in the basin, to the left of the photograph.

19 Antelope Springs channel samples, 13 were unimodal and 6 were bimodal. Granules again failed to occur as the modal class in the unimodal sediments. The secondary mode of 5 of the 6 bimodal samples, however, did coincide with the granule size class. If the frequency of occurrence of the 8 size classes as both primary and secondary modes is calculated, then one finds that granules occur more frequently than 3 and less frequently than 4 of the size classes. On this basis, therefore, granules are very nearly the median mode of the bolson sediments.

The basis for this calculation is arbitrary and results in an exaggeration of the frequency of modal occurrence of granules. Although the fact is frequently overlooked, modality is also arbitrary, however. Only a single sample among the 90 studied was unimodal when class intervals were chosen at half phi units; the remaining 89 samples were polymodal. Upon grouping weights to produce intervals in accord with the full phi units of the Wentworth size classification, however, the 37 unimodal samples discussed previously were produced. Because any classification of size must be considered arbitrary rather than natural, the obvious conclusion to be drawn is that the unimodal samples result from averaging of the data. Many of the bimodal samples could be transformed into unimodal distributions by setting the class intervals equal to 2 phi units. In the limit, of course, any sediment could be shown to be not only unimodal but of the same size class as any other.

A more pertinent factor than frequency of occurrence as a modal class, however, is the absolute abundance of granules. Their abundance in the granule-to-clay size fraction (fig. 99) ranges from 15 to 20 percent near the basin margin to the north and east and is slightly less on the western side, in the Wyman-Crooked Creeks fan area. This fact may reflect greater absolute transport distances for these sediments. At least 3–5 percent of this fraction consists of granules at any point in the center of the basin. Although this percentage is a maximum for the total sediment, because gravel has been omitted, it is pertinent to the observations of Yatsu (1959), who argued that the break in slope along the intersection of an alluvial fan with the valley floor coincides with the modal frequency break in the gravel-to-sand distribution referred to above. Granules should, therefore, be absent below this break in slope or in the central part of the basin. The evidence from Deep Springs Valley, however, suggests a decrease in abundance of granules toward the basin center but not their disappearance.

The granule high in the southwestern part of the area shown is difficult to account for, except as an actual "island" of granules. It will be shown below that the area between this high and the 25-percent granule high to the east is, in effect, a crossroad of sediment mixing in the basin. The seemingly anomalous high may therefore represent the sum of granule contributions from the northeast and eastern trends which has subsequently been isolated by a flood of finer sediments from the Wyman-Crooked Creeks system.

The decrease in abundance of granules within relatively short distances from source areas not only requires explanation but serves to refute the hydraulic answer provided by Sundborg (1956). Because transport of the surface sediments considered here is relatively infrequent, constant motion of the granule size class cannot be appealed to in the search for the cause. Selective sorting in a downfan direction might appear to be a likely alternative. The studies referred to by Pettijohn (1957, p. 47), however, treated ancient as well as recent sediments. If selective sorting of sediment is the general explanation for the reported scarcity of granules, then it is difficult to understand why granule concentrations are absent in ancient sediments; lag deposits would necessarily have to accumulate in some areas but, to the writer's knowledge, none have been reported in the literature. A mechanical theory, suggested by the data from Deep Springs Valley, is thought to be the most reasonable explanation.

About 80 percent of the granules examined was found to consist of polymineralic rock fragments; quartz and feldspar grains constituted the remainder.

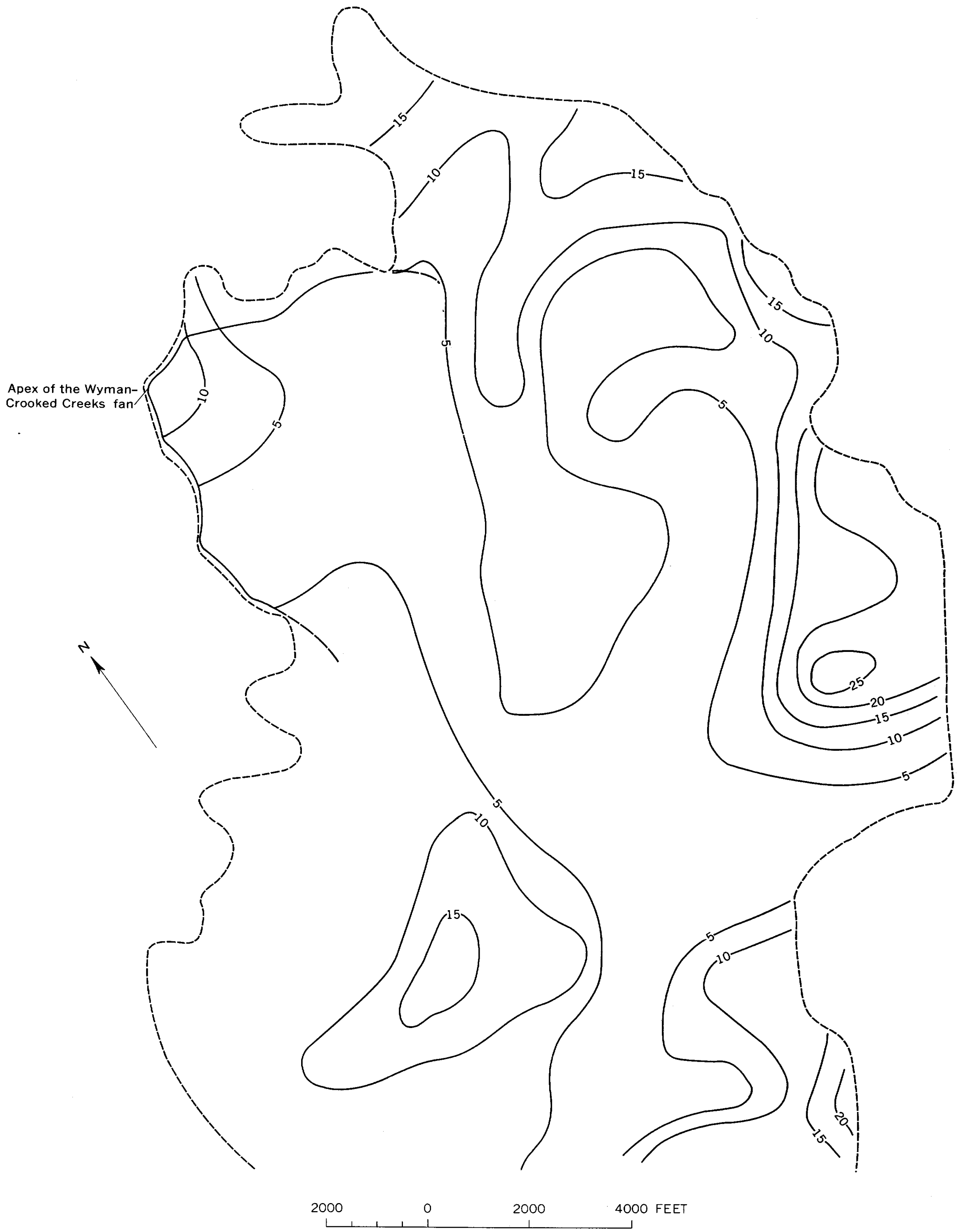

FIGURE 99.—Map showing the distribution of granules in the north end of Deep Springs Valley. Isopleth interval 5 percent by weight of the granule-to-clay size fraction

This implies that the plutonic rocks, which are the primary source of the clastic sediments considered, must consist of mineral grains that are predominantly less than 2 mm in size. The results of Dake (1921), who found that less than 10 percent of the quartz grains in plutonic rocks exceeded 1 mm in size, tend to confirm this implication. The fact that medium or coarse sand is the dominant size class in all of the unimodal samples studied, whereas granules are more abundant in the bimodal samples, suggests that the polymineralic granules are unstable aggregates of sand-size grains. These aggregates tend to disappear in nature by reason of rapid reduction to their components. The reduction is achieved by mechanical disintegration through weathering and is aided by periods of transport, however brief. The transition from polymineralic granules to sand-size components is accompanied by a consequent change from predominantly bimodal to unimodal frequency distributions. The deficiency of discrete mineral grains in the 2- to 4-mm size range in plutonic source rocks should be regarded as the fundamental cause of the relative scarcity of granules in nature.

SILT AND CLAY

The most striking single feature shared by all but 2 of the 90 samples studied is the scarcity of clay. The 2 exceptions will be explained below. The weight percentage of clay (table 3) in samples from the north end of the valley ranged from 0 to 6.37 and averaged about 2.0 percent. Samples from the Antelope Springs channel (pl. 9) range from 0 to 2.76 percent. Samples from Paiute Chute and the Owens Valley mudflow averaged about 2.5 and 3.0 percent, respectively. Although much care was taken in both collection and analysis of the samples, it is probable that some clay was lost. Because these weight percentages apply to the granule-to-clay fraction alone, however, the percentage of clay would be still lower if the total sediment provided the basis for calculation.

In contrast to the scarcity of clay, silt is very abundant. Silt occurred as either the primary or secondary mode in 31 of the 47 bimodal samples, referred to previously, and had the greatest modal frequency on this basis. The relative abundance of silt to clay is given in table 3. Noteworthy is the fact that even the thin atypical Wyman mudflow, which contained 17.38 percent clay, had a silt: clay ratio of more than 4 to 1. This ratio is in accord with those of the more typical samples from the Owens Valley mudflow (BP–1, BP–2) and Paiute Chute.

Because silt: clay ratios appeared to be consistent in the mudflow samples despite dissimilarity of the flows, it was thought that this variable might prove to be significant. Accordingly, the ratios for samples in the north end of the valley were plotted (fig. 100). Isopleths were not drawn for two reasons. First, it was apparent that four general areas occurred in which silt: clay ratios were markedly low in comparison to a relatively uniform and higher value elsewhere in the basin. Second, samples that contained no clay had, of course, infinite ratios. The path along which these infinite values occurred was recognized as that of the most frequent wind track observed in the field (p. 137). For these reasons the four areas of low ratio were outlined, and an average value was obtained for the samples that occur within each area. The same procedure was used to separate the remainder of the basin area into fields of average silt: clay values, as shown, and the inferred wind track was drawn along the path of 0 clay percent.

Several features of figure 100 are worthy of discussion. If the inferred wind track is followed as drawn, from the southeast corner northward, then the field of lowest silt: clay values in the basin (1.37) can be seen to lie in a reentrant to the right that is a natural depositional area. One of the samples within this field is 5, —1 which is one of the two exceptional samples noted above. The high clay content (26.58 percent) of this sample can best be explained by recourse to the wind deposition postulated.

The loop drawn in the inferred wind track to the northeast of the 1.37 field could not have been constructed exactly as shown from the data alone; the latter indicate only that the path must turn sharply to the left. A small incipient playa occurs precisely in this area, however. On several occasions, while in the field, gusts of wind approached this playa and circled it for a considerable length of time. The resulting windblown cloud of dust could be readily observed. The silt: clay ratios reflect this fact; high values encircle the playa and attest to the winnowing of clay, whereas much clay is present on the playa proper. The great bulk of Piper Mountain (pl. 8) apparently deflects the wind track to the west of the playa as shown (fig. 100); the remainder of the track is problematical. The field with average ratios of 4.44 may reflect both deposition of windblown clay along the elongate tongue pointing west and some contribution of clay from the northeast corner of the basin by runoff. The clays derived from the basalts in the northeast corner are well delineated by the 6.50 field.

The infinite value that occurs in the extreme north end of the basin is also interpreted as part of the wind track. Wind directions in this area were not noted in the field, however, and the track is arbitrarily shown leaving rather than entering the basin. Upon

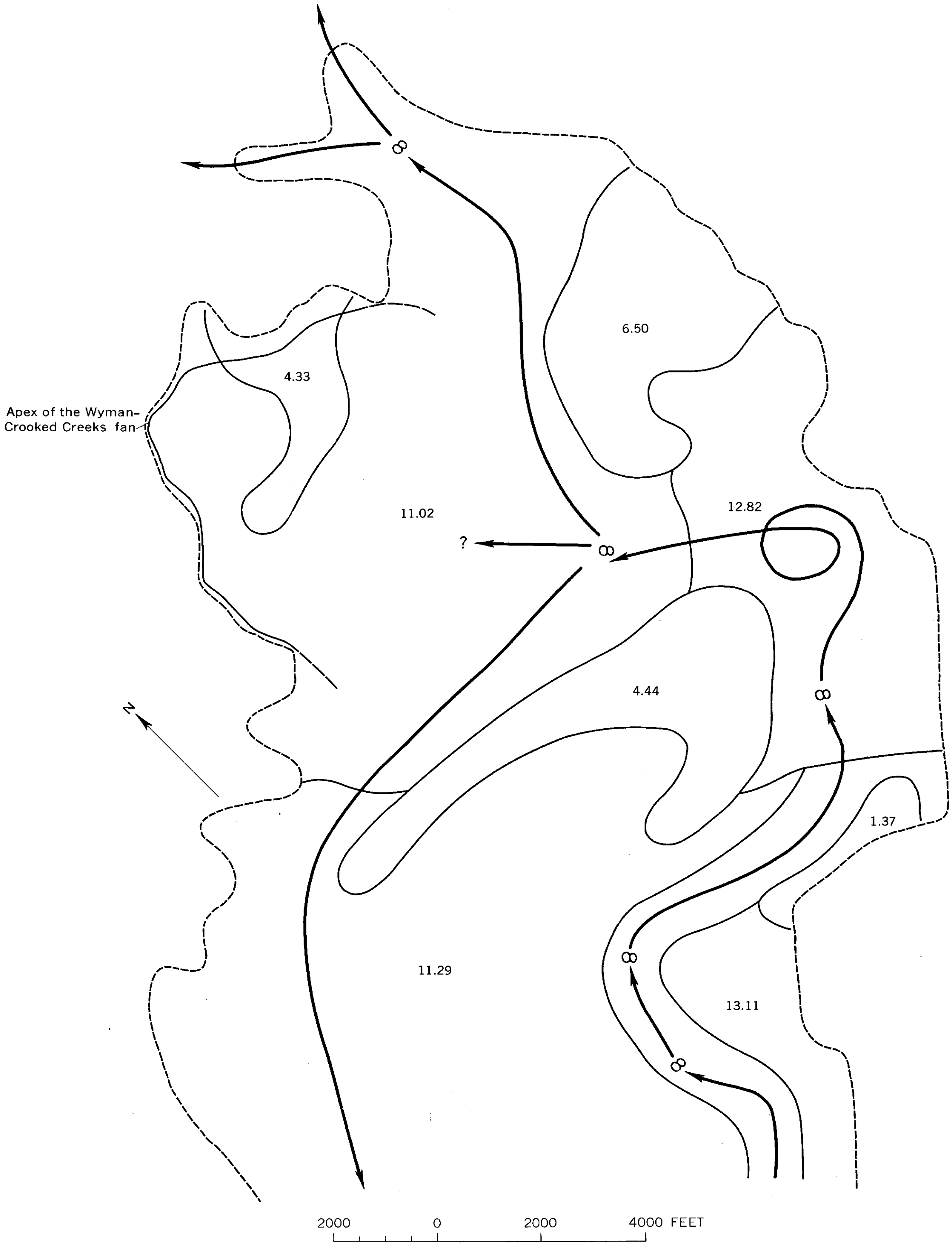

FIGURE 100.—Map showing silt: clay ratios and the inferred wind track in the north end of Deep Springs Valley. Values are means of sample ratios within areas shown. Infinite values occur along track.

occasion, the winds might equally well enter the basin along this route.

The remaining field of low silt: clay ratios (4.33) in the northwestern part of the basin is not thought to result from wind action. The basalt mass in this area (fig. 87) has probably contributed to the relatively high clay content represented by the field.

Samples from the Antelope Springs channel contain less clay, on the average, than do those from the north end of the valley or elsewhere (table 3). This may be attributed to the fact that much runoff occurs within the channel or that the rock types present do not weather as readily as basalt and granite. No consistent downchannel changes in clay abundance or in silt: clay ratios occur. This probably results from the fact that each zone of bankful deposition (pl. 9), to be discussed below, acts as a separate source area at various points in the channel system.

In summary, the study of silt and clay indicate that (1) clay is relatively scarce, whereas silt is abundant today; this is true not only for fan surface and channel samples but for mudflows as well; (2) silt: clay ratios reflect, in part, the effect of wind action, and in Deep Springs Valley the dominant wind track can be inferred from these ratios; and (3) wind action can affect the size distribution of basin sediments by reapportionment of the finer fractions.

PARAMETERS OF THE SIZE DISTRIBUTION

It should be apparent from the foregoing discussion of granules and silt: clay ratios that any attempt to characterize the granule-to-clay fraction in a bolson environment that ignores the tails of the frequency distribution is apt to be useless. The statistical measures used in this report were chosen, as noted above, because they encompass a reasonably large percentage of the total distribution of any given fraction and therefore reflect the variations in abundance of granules and silt and clay in the basin sediments.

Previous studies of the size distribution of sediments have treated beach, dune, and river sands in various combinations (Mason and Folk, 1958; Friedman, 1961). These studies have mainly sought to establish criteria for distinction between the environments cited, and they have met with varying degrees of success. The approach consisted of the determination of mean size, standard deviation, skewness, and kurtosis, followed by plotting of these variables against each other in one or more of the six possible combinations. In the present report the approach employed was to contour the areal distribution of each of the parameters. Certain of these maps have no parallel in the literature, to the writer's knowledge, and therefore cannot be compared with previous work. In addition, the granule-to-clay fraction has been treated as a distinct sedimentary unit, whereas the usual practice is to consider only the <2-mm fraction or else to treat the entire sediment if gravel and sand occur together. For this reason the absolute values (table 4) of the size-distribution parameters also cannot be compared with previous results. The treatment in this report, however, lends itself particularly well to both visual interpretation of the distribution of sediments and the geologic interpretation of skewness and kurtosis.

MEAN SIZE

The distribution of mean size of the granule-to-clay fraction in the north end of Deep Springs Valley is shown on figure 101. Because the contour interval is in phi units, the values are inversely proportional to mean size in terms of millimeters. It is instructive to compare this map with those of mean pebble size (fig. 94) and others and to note the approximate correspondence of closures or fluctuations. In figure 101, the large 2.0-phi closure on the Wyman-Crooked Creeks fan, the contour gap near the apex, and the size reversal along the northern edge of the fan all have their equivalents on the pebble map. This distribution probably reflects the fine sediment that is contributed from the basalt mass (fig. 87) at the basin margin. The silt: clay field of 4.33 (fig. 100) in this region and the location of the 5-percent granule contour (fig. 99) around the fan apex indicate that this is true.

Size fluctuation in the center of the pebble map (fig. 94), which bears resemblance to a pair of "sunny-side" eggs, is represented on figure 101 by the 1.5-phi reversal belt in the same location.

The flood of fine sediments contributed from the weathering of basalts in the northeast corner of the basin is roughly outlined by the linear 2.0-phi field on the right side of figure 101. This contribution was noted in discussion of silt: clay ratios and is shown on figure 100. There is also a general deficiency of granules within the area (fig. 99).

The zones of sediment mixing in the north end of the valley apparently result not only from the intersection of sediment transport paths, but they also reflect additional complications produced by wind action and the differing size contributions of individual rock types. The mean-size distribution (fig. 101) suggests that mixing occurs, but the map cannot reveal how or why without recourse to the additional data discussed.

Fluctuation of mean size in the Antelope Springs channel (pl. 9) is shown on figure 102. As noted above, each zone of bankfull deposition probably represents

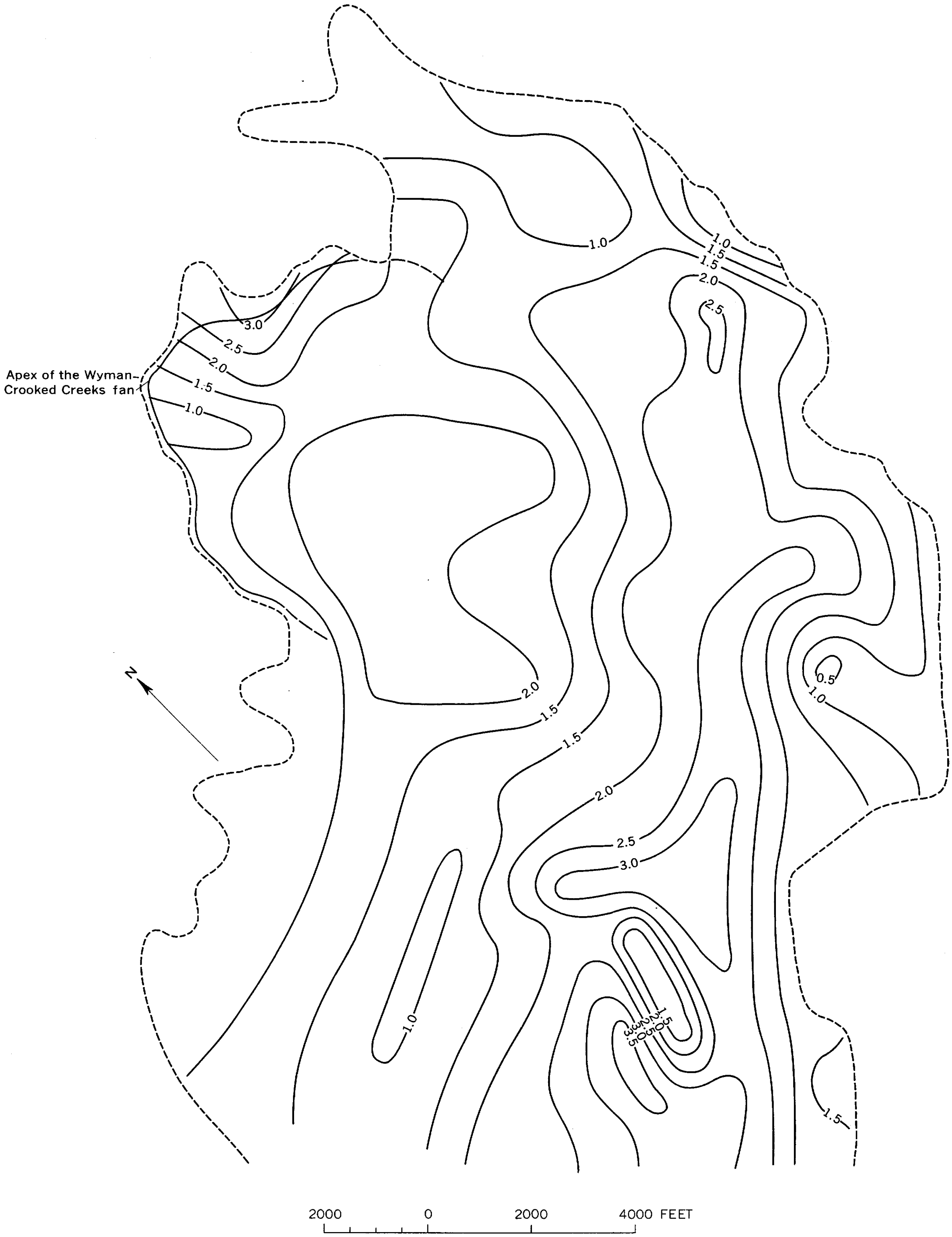

FIGURE 101.—Map showing the mean size of the granule-to-clay fraction in the north end of Deep Springs Valley. Isopleth interval 0.5 ϕ. $\phi=-\log_2 D$ (mm).

an individual source area for the respective reach below, and variation in mean size in a down-channel direction can be expected.

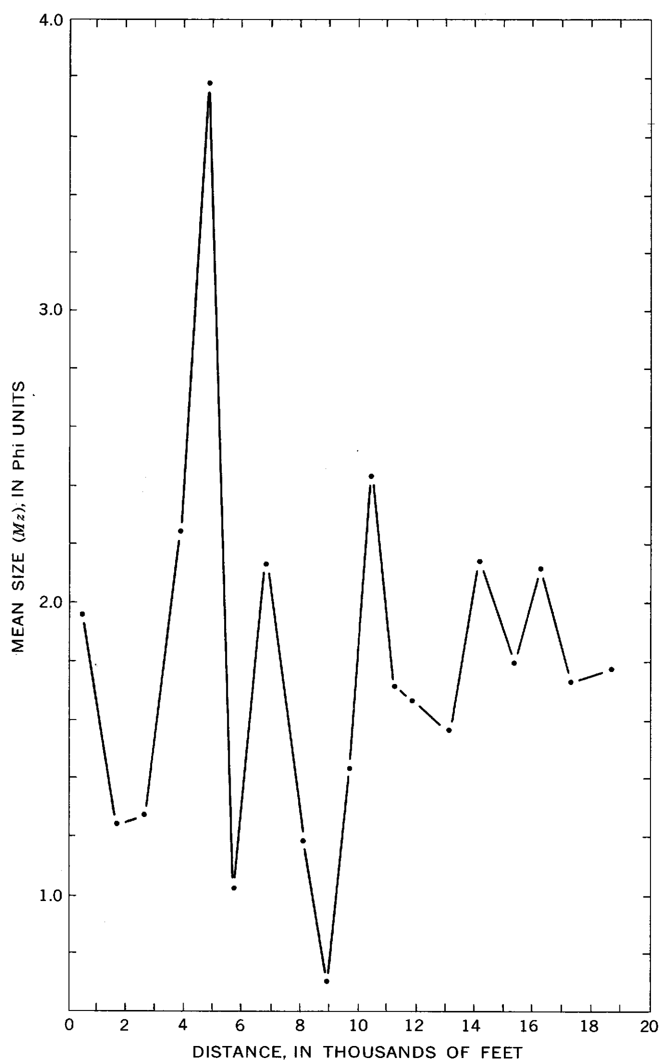

FIGURE 102.—Graph showing the relation of the mean size of the granule-to-clay fraction to distance from the fan apex in the Antelope Springs channel.

STANDARD DEVIATION

The inclusive graphic standard deviation (σ_I) of the granule-to-clay fraction in the north end of Deep Springs Valley is shown on figure 103. The degree of sorting increases down the Wyman-Crooked Creeks fan toward the basin center, from the northeastern rim of the basin. The map suggests that the more highly sorted material in the lower right corner, enclosed by the 1.9$\sigma\phi$ isopleth, may be derived from the Wyman-Crooked Creeks fan because of the accordance of trends. This was not implied by the mean-size map (fig. 101); the origin of the sediments in this area was indeterminate.

Sorting decreases to either side of the best sorted sediments, culminating in 2 belts of 2.7$\sigma\phi$, labeled *A* and *B*, but the reason is not evident from the map. If silt : clay ratios and the inferred wind track (fig. 100) are considered, however, then one explanation can be offered. The area of windblown clay deposition in the basin reentrant (1.37 field) is adjacent to one of these isopleths (A_1) on figure 103. The surface samples in the 1.37 field contain abundant fine material and are well sorted (1.5$\sigma\phi$). The belt of 2.7$\sigma\phi$ values to the left at A_2, however, lies just to the north of the inferred wind track, and the poorly sorted material in this zone might be expected because of the 3.1$\sigma\phi$ high that heads into this area from the northeast.

The other 2.7$\sigma\phi$ belt to the west (*B*) and the large adjacent 2.3$\sigma\phi$ field probably result from influx of clay (fig. 100) combined with the granule concentration (fig. 99) in this area. The distribution of mean size of both pebbles (fig. 94) and the granule-to-clay fraction (fig. 101) also shows high values in this part of the basin.

The data on granules and silt: clay ratios, which represent the tails of the granule-to-clay-size frequency distribution, show that standard deviation also reflects the complex mixing of sediments in the north end of the valley. Standard deviation, like mean size, cannot alone describe the distribution of sediment.

Standard deviation is plotted against distance from the fan apex in the Antelope Springs channel on figure 104. The data suggest a general downstream trend; standard deviation decreases from about 2.0 to 1.0$\sigma\phi$. This decrease accords with a disappearance of granules, the low abundance of clay, and the fact that most of the channel samples are unimodal. These factors may indicate that the sediments have been subjected to more frequent runoff than have those in the north end of the valley and therefore tend to become better sorted.

SKEWNESS

Inclusive graphic skewness (Sk_I) of the granule-to-clay fraction in the north end of Deep Springs Valley is shown in figure 105. The general appearance of this map, in contrast to those showing mean size (fig. 101) and standard deviation (fig. 103), is striking. The writer must confess that if he were confronted with such a map in the literature, the urge to test the validity of the contouring would be irresistible. The reader is provided with the opportunity to satisfy his own urge through access to the basic data in table 4.

The most important single feature of the distribution is the field of zero skewness. This defines more clearly than before the zone of sediment mixing in the basin because it reflects a balance between the tails of the frequency distribution. With recourse again to the granule (fig. 99) and silt: clay ratio (fig. 100) distribu-

FIGURE 103.—Map showing the standard deviation of the granule-to-clay size fraction in the north end of Deep Springs Valley. The sorting of sediments in the regions labeled *A* and *B* is discussed in the text. Isopleth interval 0.2 $\sigma\phi$.

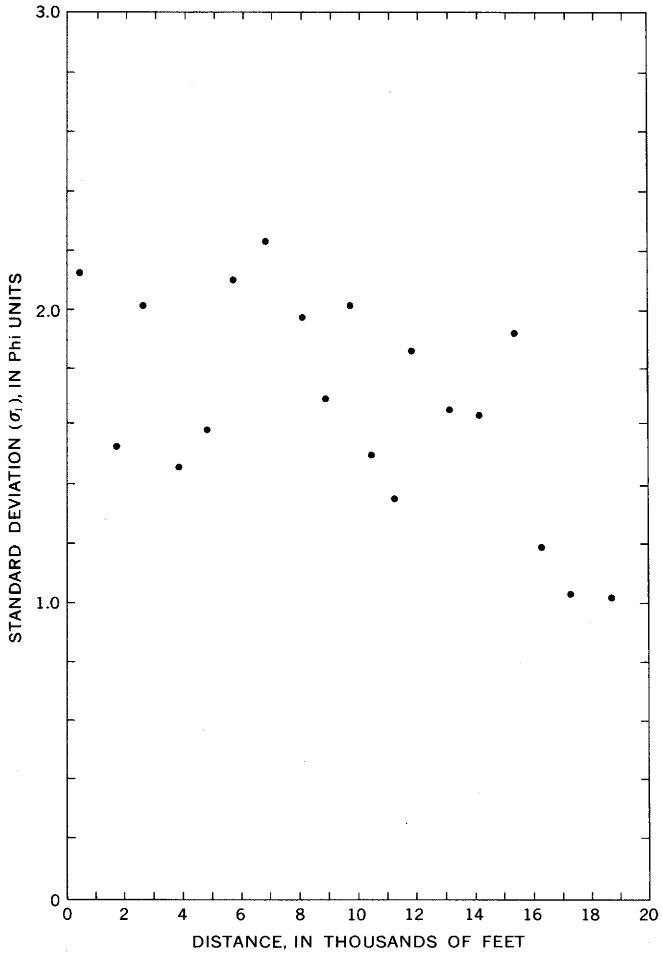

FIGURE 104.—Graph showing the relation of the standard deviation of the granule-to-clay size fraction to distance from the fan apex in the Antelope Spring channel.

tions, the skewness map can be seen to depict accurately the effects of mixing of sediments within the basin. High skewness values occur in areas where granules are abundant and decrease to zero where this size class either disappears or is balanced by an influx of fine-grained sediment.

Skewness of the granule-to-clay fraction in the samples from the Antelope Springs channel is shown on figure 106. Skewness ranges from +0.62 to —0.30. The channel samples are, on the average, only slightly positively skewed, and values fluctuate around +0.15, approximately. This might be expected in light of data previously provided; two-thirds of the samples are unimodal, and all have a high concentration of sediment in the medium-sand size class.

KURTOSIS

As defined in the discussion of basic data (p. 144), graphic kurtosis is the ratio of the spread of the tails of the distribution to that of the central portion. It is a measure of considerable sensitivity, and it was anticipated that kurtosis might prove useful in a treatment of bolson sediments, because the tails of the distribution are of primary concern. One would expect that the kurtosis distribution would complement that of skewness, because the latter reflects the relative balance of the tails whereas kurtosis reflects their relative abundance. This expectation has largely been satisfied; it will be shown by comparison of skewness (fig. 105) with kurtosis (fig. 107) distributions in the north end of Deep Springs Valley.

Kurtosis values are shown (fig. 107) in terms of departure from normality ($Kg = 1.00$) times 100. This means that the largest map value (85), for example, represents samples for which the spread from ϕ_5 to ϕ_{95} is 1.85 times as large as it would be for samples with normal kurtosis. It is evident that the belt of greatest departures on figure 107 cuts across the upper portion of the zero skewness field (fig. 105) in an east-west direction. This shows that although the tails of the distribution become balanced, the magnitude of the spread of the tails remains high and actually increases to the west. Consideration of the granule distribution (fig. 99) and silt: clay ratios (fig. 100) provides good reason for the location of the high kurtosis closure in the basin center; an abundance of both clay and granules occurs there.

The adjacent high kurtosis closure to the south, with a map value of 30, reflects the complexities of sediment mixing referred to above. These complexities include windblown clay deposition, the occurrence of granules and coarse sands contributed by the eastern reentrant, products of the weathered basalt from the northeast, and sediments derived from the Wyman-Crooked Creeks system. Kurtosis is sufficiently sensitive to the presence or absence of fine sediment in the basin samples to suggest that the inferred wind track (fig. 100) may be slightly mislocated as drawn. If the track were shifted slightly to the north along its reach between the 1.37 and 4.44 silt: clay ratio fields, then it would better coincide with the break between the two adjacent kurtosis highs of 85 and 30 (fig. 107).

The remainder of the inferred wind track need not be altered. It is interesting to note the support that each variable lends to the other. This would, of course, be predicted from theoretical considerations. The kurtosis low (—20) to the east of the two highs discussed above, roughly accords with the loop in the wind track (fig. 100) at the incipient playa. As the wind track is followed from the playa to the west and then south it coincides with a well-defined area of nearly normal kurtosis values that border the dominant (85) high. Another belt of high kurtosis values that extends from the northeast corner of the basin to its center is prob-

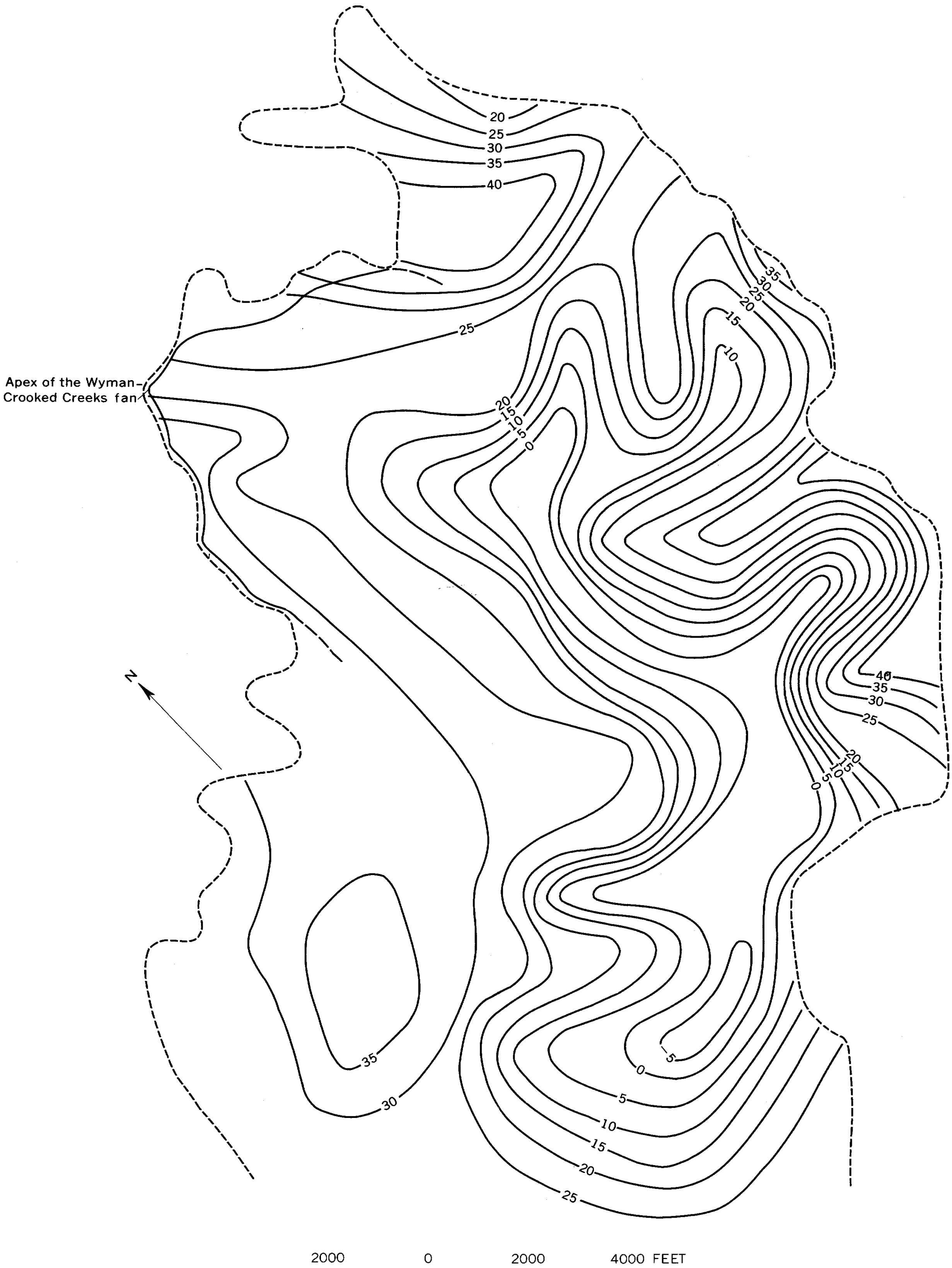

FIGURE 105.—Map showing the skewness of the granule-to-clay size fraction in the north end of Deep Springs Valley. Isopleth values=100×departure from $Sk_I=0.0$.

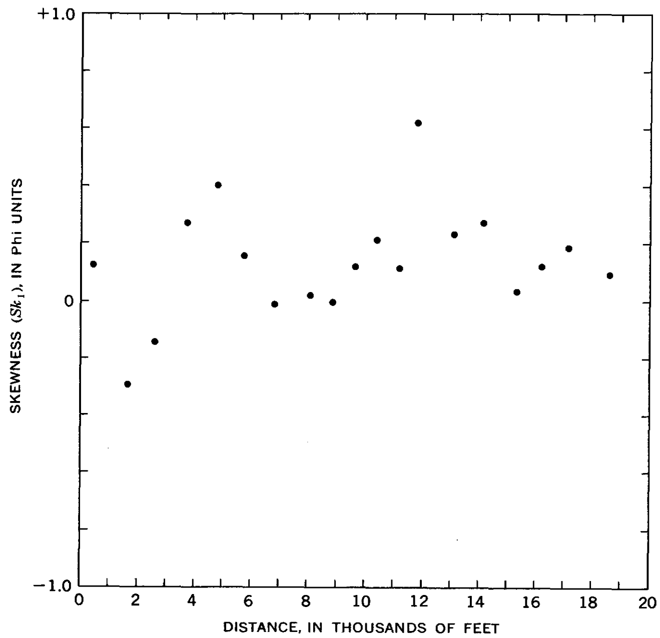

FIGURE 106.—Graph showing the relation of the skewness of the granule-to-clay size fraction to distance from the fan apex in the Antelope Springs channel.

ably a reflection of the merger between the granule tongue (fig. 99) in the upper part of the basin and the 6.50 silt: clay ratio field (fig. 100).

Finally, kurtosis values on the Wyman-Crooked Creeks fan are worthy of note. Despite the fact that the basalt on the north side of the fan (fig. 87) contributes fine sediment, as shown by both the mean size (fig. 101) and silt: clay (fig. 100) distributions, granules are relatively scarce (fig. 99) and this has produced low kurtosis values. The zero field ($Kg = 1.00$) accords with the fan limits in this region (fig. 107). In addition, kurtosis is the only parameter discussed thus far to suggest that the sedimentary trend of the fan is toward the east as well as toward the south. The probability that sediments from the Wyman-Crooked Creeks area were formerly transported to the east was mentioned earlier in connection with Pleistocene history. Sediment transport paths inferred from the tractive force data, to be discussed below, also support this contention. These facts also attest to the usefulness and sensitivity of kurtosis in the interpretation of clastic sediments in such an environment.

A scatter plot of kurtosis against distance downfan in the Antelope Springs channel (fig. 108) shows that within a wide scatter kurtosis decreases slightly, approaching normality near the end of the channel.

SUMMARY OF SIZE-DISTRIBUTION PARAMETERS

It has been shown that each of the parameters of the granule-to-clay fraction is a mappable variable and that useful information can be derived from such maps. The tails of the size distribution are of fundamental importance in the interpretation of bolson sediments because they have a marked effect upon the parameters. It is probable that significant quantities of granules and deposition or winnowing of fine sediment are common to all bolsons. The parameter maps and their implications should, therefore, have application to basins other than Deep Springs Valley and possibly to ancient sedimentary deposits as well.

The map of each of the four parameters of the size distribution required some knowledge of the distribution of the tails for satisfactory interpretation and, of course, all the maps are indeed interrelated. For this reason it is not possible to claim that certain of the parameter maps are of greater significance than others, but the reader of this report will probably agree at this point that the two orphans of basin studies, namely skewness and kurtosis, have been much neglected in the past. In addition to the merits of these two variables in investigations of sediment mixing, their absolute values may someday prove to be of significance. If competence is affected by the range of size of the sediment present (Fahnestock, 1961), then absolute kurtosis values are the best means of comparing range of size. Hence, the kurtosis of sediments should be considered in sediment-transport problems.

CLAY MINERALOGY

To determine the kinds of clay minerals present in the basin sediments and their relative abundance, 43 samples were studied. These include 18 from the Antelope Springs channel, 20 from the northern end of Deep Springs Valley, 2 from Paiute Chute, and 3 samples from mudflows. Two of the last were obtained from a mudflow that occurred in Owens Valley, west of Big Pine, Calif., on August 22, 1961. These samples are listed in table 6 together with their respective clay-mineral ratios that are discussed below.

Oriented clay aggregates were prepared on glass slides by sedimentation, and diffraction data were obtained from a Norelco X-ray diffractometer with automatic recording unit. The samples were initially run untreated; after saturation with distilled water a second pattern was obtained for each. All untreated samples exhibited one peak intensity corresponding to an interlayer spacing ranging from 14.21 A to 15.80 A. Saturation with water shifted this peak to smaller 2 θ values, indicating expansion. The expanded interlayer spacings ranged from 17 A to 20 A.

Because this paper is not devoted to an intensive quantitative study of clay mineralogy and because of the marked expansion with water noted above, glycolation was not deemed necessary. The mineral

Figure 107.—Map showing the kurtosis of the granule-to-clay size fraction in the north end of Deep Springs Valley. Isopleth values=100×departure from $Kg=1.00$.

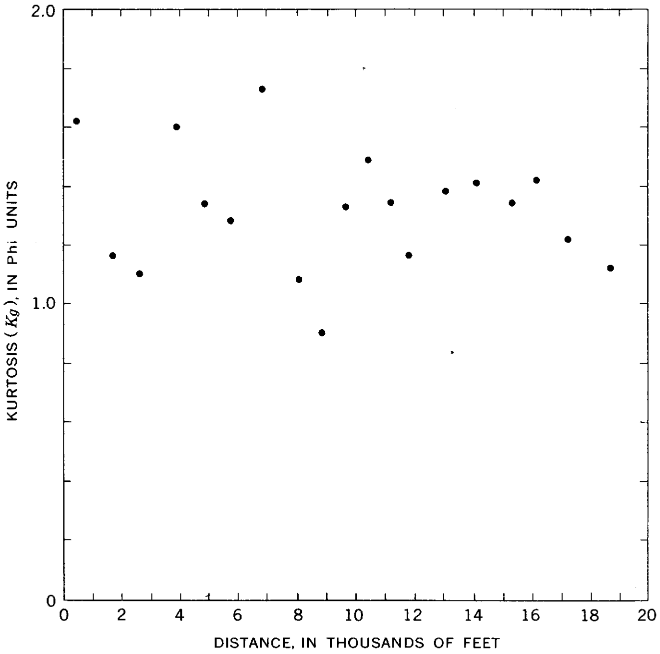

FIGURE 108.—Graph showing the relation of the kurtosis of the granule-to-clay size fraction to distance from the fan apex in the Antelope Springs channel.

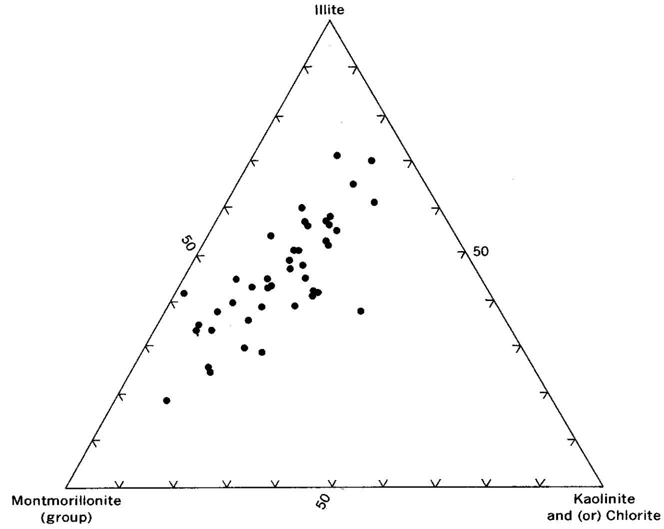

FIGURE 109.—Three-component diagram showing montmorillonite : illite : kaolinite and (or) chlorite ratios for 43 clay samples.

group that expands from 2 A to 4 A in the presence of water is designated as montmorillonite in the ratios presented. It should be stressed, however, that this designation undoubtedly includes various mixed-layer clay minerals. This is attested to by the lack of sharpness of many of the peaks and not uncommon base distance of about 4° 2 θ.

A nonexpanding clay mineral with an interlayer spacing of 10 A has been designated as illite and a second mineral, with a peak corresponding to a spacing of 7 A, as kaolinite and (or) chlorite. The separation of the (002) reflection of chlorite from the (001) reflection of kaolinite by heating or other methods was not attempted. It was noted, however, that no recognizable peak corresponding to 14 A was present after water saturation of the samples. This would suggest that chlorite is not very abundant in the samples studied; the fact that chlorite and (or) kaolinite is not abundant can be seen from the data (fig. 109).

To determine the ratio of montmorillonite: illite: kaolinite and (or) chlorite, as they are defined above, the integrated areas of the (001) reflections were determined by planimeter. These areas are linear functions of the weight percent of each mineral (Talvenheimo and White, 1952). Many factors will, of course, affect the reflection intensities. Among these are the polarization, Lorentz, and temperature factors, the degree of crystallinity, chemical composition, and others (Cullity, 1956). Because these factors have been ignored and because montmorillonite is defined on the sole criterion of expandability, the results presented must be regarded as approximations.

Weaver (1958) stated that a sample with equal amounts of montmorillonite and illite will yield an integrated intensity ratio of 3:1 in favor of montmorillonite. This result holds for 14 A montmorillonite; other ratios, varying with the montmorillonite interlayer spacing, have been published. As previously indicated, a montmorillonite group rather than pure 14 A montmorillonite is treated in this report. Because some value must be chosen, however, a factor of 3.0 was assigned for montmorillonite group spacings as wide as 14.9 A and 3.5 for spacings of 15.0 A and greater. Accordingly, planimeter measurements of the integrated intensity values for the (001) reflection of illite have been multiplied by 3.0, as appropriate, in order to arrive at the montmorillonite: illite ratios given.

Johns, Grim, and Bradley (1954) state that equal amounts of kaolinite, sedimentary chlorite, and illite will yield equal integrated intensities. The illite: kaolinite and (or) chlorite ratios can therefore be determined directly from planimeter measurements of the integrated intensities of the respective (001) reflections.

The montmorillonite (group): illite: kaolinite and (or) chlorite ratios, determined as described above, were recalculated to a base of 10 and used to plot the three-component diagram shown on figure 109. Despite the qualitative to semiquantitative approach employed, which renders the data valid only as a first approximation, two conclusions may be drawn from the diagram. First, a considerable range of montmorillonite: illite values exist despite the fact that all samples occur in a relatively uniform climatic environment. This suggests a relationship between clay-mineral type and source-rock composition, which is to some extent borne out by the data. Samples exhibiting high montmorillonite: illite values occurred mainly in basalt-rich areas, whereas all samples but one from the Antelope Springs channel contained more illite than montmorillonite. The source rocks of the Antelope Springs sediments are complex, but they include no basalt. The suggestion that a correlation exists is supported by previous work.

Droste (1961) studied the clay-mineral assemblages in several California basins that are in the neighborhood of Deep Springs Valley. He found that the clay suites strongly reflect parent rock types; samples rich in montmorillonite often contained volcanic glass. Eardley and Gvosdetsky (1960) also reported the occurrence of abundant montmorillonite in association with an ash layer in their study of a core from the Great Salt Lake.

The approximate constancy and relative scarcity of kaolinite and (or) chlorite suggested by the data on figure 109 also agrees with Droste's (1961) results. In Deep Springs Valley the sum of illite plus montmorillonite in the average sample is about 80 percent.

Because of these facts, the writer was tempted to treat clay mineralogy in the same manner as other variables already discussed and map the distribution in the basin. The data obtained, however, are both too few and too approximate. In addition, recourse to mixing of clays by means of wind action in order to explain exceptions proved necessary in an initial attempt. For these reasons it was considered best not to proceed beyond the presentation on figure 109. Clay mineralogy might well prove to be a mappable parameter in many western basins, however, if detailed studies were undertaken.

SEDIMENT TRANSPORT

An important aspect of the transport of sediments on alluvial fans is the competence of the transporting medium. Although the occurrence of large particles on alluvial fans implies a high degree of competence, it should be borne in mind during the discussion that follows that the nature of the source rock is a factor of importance. The competence of the transporting medium may have been sufficient to move still larger particles than those seen on fans, had they been available at the source. It cannot be assumed, therefore, that the competence of transport was necessarily greater on fans that exhibit the largest particles.

COMPETENCE CONSIDERATIONS

Competence is usually equated with the velocity of flow, and the relationship is stated in terms of particle diameter. Leliavsky (1955, p. 24) states that the earliest formula was provided in 1753 by Brahms who employed particle weight rather than diameter; but the report of Suchier (1883), listing velocities required to transport particles of "pea," "hazelnut," and "pigeon egg" dimensions, is typical of several early treatments. Gilbert (1914) considered the transport of particles as large as 6 mm in diameter, as did Nevin (1946) who reviewed the competence problem. The basic competence curves of Hjulström (1935, p. 298) have been extended and modified to include the transport of cobbles (Menard, 1950) and boulders (Fahnestock, 1961). Menard's work provides some information on the velocities necessary to keep a moving cobble in motion, but these are not comparable to the velocities required to initiate movement. The study by Fahnestock has produced the most pertinent data; the particles considered were as large as 1.8 feet in diameter. It was found that these particles moved more readily than would have been predicted; this was attributed to the presence of a mixture of sediment sizes. Even the data of Fahnestock (1961), however, cannot be safely extrapolated to include particles from several to many times as large, that occur on alluvial fans. The use of any velocity power law is, therefore, best avoided.

As an alternative measure of competence, tractive force can be employed. This is expressed as $\tau=\gamma dS$, where γ is the specific weight of the transporting medium, d is the depth of flow, and S is the slope of the energy gradient. The tractive force thus defined pertains to the shear exerted on the upper layer of the bed material. The formula, or some variant thereof, is widely used in engineering studies and provides one of the few available means of obtaining an approximation of the competence of transport. It is used in this report because S and d can more readily be estimated than the velocity of flow.

It should be emphasized that, at best, only an approximation to the true competence of transport can be sought in the field. For this reason several slightly more sophisticated formulae have not been considered. The shear exerted upon a boulder in a channel bed, for example, might be determined from the relationship $\frac{F/A}{\rho r^2/2}=C_D$ (Rouse, 1950, p. 122), where A is the area of the projecting surface, normal to the direction of flow, ρ is the density and r the velocity of flow, C_D is a drag coefficient, and F is the force or shear exerted. Any attempt to substitute values for the density, velocity, and drag coefficient required, however, would be a guess at best; such variables cannot be derived from field data.

Having chosen the simple expression $\tau = \gamma dS$, the main problem lies in equating field data with the required variables so that tractive force, or competence, can be determined. The key to a field approximation of tractive force is thought to lie in consideration of the largest particles. Several boulders contained pockets of pebbles, sand, silt, and clay on their uppermost side, several feet above fan surfaces. Because the finer sediment in these pockets is diverse in both lithology and size, it may represent a part of the sediment load that was transported together with the boulder and trapped upon cessation of the flow. Although later floods might have emplaced some of this sediment, this possibility seems more remote. The fan surfaces show no sign of subsequent floods; weathering stain or varnish extend down the sides of boulders to precisely the level of the fan surface on which they rest.

These observations suggest that either the depth of flow exceeded the diameter of the largest particle during transport or the finer sediment was acquired by the rolling of boulders through a shallower flow. Leopold and Miller (1956) state that cobbles moved in flash floods in arroyos even when the water was no deeper than half the diameter of the rolling object. On the other hand, the upper reaches of the Owens Valley mudflow (figs. 110, 111) clearly exceeded in depth the diameter of the largest particle. Depth of flow can therefore range from a lower limit of half the diameter of the largest particle to several times this value. Because a relationship exists, maximum particle size has been substituted for d, the depth of flow, in the tractive force expression. This value may be too low for the upper reaches of fans and too high for the lower reaches. It should be noted that this substitution cannot be made in areas that contain desert pavements, because in such areas the largest particles break down to smaller fragments and the original dimensions are therefore unknown.

The required slope in the tractive-force expression is that of the energy gradient, whereas the slopes measured in the field, by methods previously described (p. 141), are equivalent to bed gradients. Many measurements indicate that the bed slope is often a good approximation of the energy gradient, however. Moreover, efforts were made to measure a reach that corresponded with the most probable paths of sediment transport. Although local runoff may produce modification of slopes on fan surfaces, this modification might well consist of uniform lowering over a given reach. For these reasons the local slopes at each sampling point have been substituted for S, the slope of the energy gradient, in the tractive force expression.

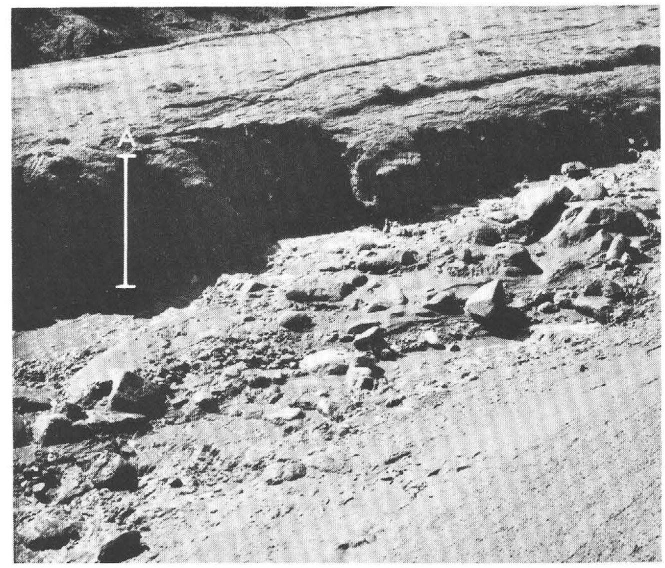

FIGURE 110.—Lateral view of the upper reach of the Owens Valley mudflow. The depth of flow at A was about 12 feet and exceeded the diameter of the largest particle that was transported. The mudflow has been dissected by a subsequent high-water stage; flowing water is visible below A and to the left of center. Note the dark color and viscous character of the mudflow deposit.

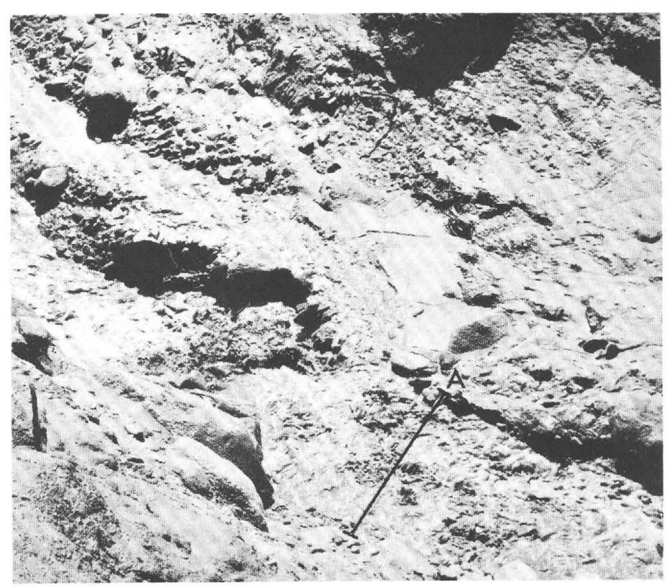

FIGURE 111.—View downstream in the upper reach of the Owens Valley mudflow. Much of the mudflow has been flushed from the channel by the turbulent water; the width of flow at A is about 5 feet. Openwork gravels may result from this process when subsequent high-water stages flush material in bankfull deposition zones.

Field values are believed to be better approximations of the slopes that prevailed during true flow conditions than values obtained from topographic maps.

The remaining variable is specific weight, or density times the acceleration of gravity. Because no single value for the density of a flow down an alluvial fan can reasonably be substituted in the tractive-force expression, this variable has been omitted. The tractive-force approximation in this report consists, therefore, of computing dS products, which are directly proportional to tractive force. These products were computed for each sampling station, their distribution mapped, and the maps interpreted as reflections of the competence of the transporting medium. The actual tractive-force values can only be determined if one assumes an arbitrary value for the specific weight of the transporting medium.

DISTRIBUTION OF ESTIMATED TRACTIVE FORCE

The distribution of slope-maximum particle-size products, or estimated tractive force, in the north end of Deep Springs Valley is shown on figure 112, and the distribution on the Antelope Springs fan is shown on figure 113. The basic data are listed in table 1, and the location of sampling stations is shown on figures 90 and 91.

The spacing of isopleths of estimated tractive force on the Antelope Springs fan (fig. 113) is rather uniform and is only slightly less so in the north end of the valley (fig. 112). The size distribution of pebbles (fig. 94) does not accord with a uniform decrease toward the basin center, however, and the relation of maximum particle size to slope (figs. 92, 93) exhibits considerable scatter. Large particles occur more frequently in apex regions (p. 145), but they are by no means restricted to these areas, as attested to by the occurrence of boulders on playas. Unless an exceedingly gross contour interval is chosen, the distribution of the largest particles cannot be mapped. If the isopleths shown on figures 112 and 113 reflect the distribution of competence, then one is confronted with the fact that, although the decrease in competence downfan is fairly uniform, maximum particle size exhibits considerable fluctuation.

This anomaly requires explanation. If one stresses the fact that a decrease in the average maximum particle size occurs on fans in order to obviate the problem, then the question of why the average size should decrease, will arise. This could be answered by recourse to selective sorting, but the bolson environment is notorious for the poor sorting that occurs therein. The floods described by Krumbein (1940) and Sharp and Nobles (1953), for example, both produced poor sorting and an irregular decrease of size. Moreover, floods of sufficient competence to transport the largest particles present probably lie in the mass transport region of the sediment transport spectrum, for which a high degree of selective sorting is not characteristic.

It is possible that floods of variable discharge could transport sediment over correspondingly varied distances downfan, but this would still fail to explain the transport of large particles across regions of gentle or zero slope. It is thought that transport by density flows, attended by buoyancy and momentum effects, provides the most reasonable answer to the observed anomaly.

DIRECTIONS OF SEDIMENT TRANSPORT

If slope-maximum particle-size isopleths are accepted as an approximation of competence isopleths, then orthogonals to these contours should represent directions of sediment transport. This assumes only that competence should decrease in the direction of flow. Although an infinite set of orthogonals can be drawn through any isopleths, those orthogonals shown (figs. 112, 113) represent a logical choice because they were drawn through contour trends. After drawing these orthogonals, it became apparent that their paths correspond closely to the location of channels on the fans. Field notes confirmed this correspondence, but to test the predictability of the method, two additional areas were studied. Paiute Chute (fig. 88) is a small fan that contains a very obvious active channel. Traverses were made up the active channel to the fan apex and down the fan surface, following boulder trains and other features suggestive of a former flow path, north of the channel. Slope and maximum particle size were determined as before at 100-foot intervals along these 2 routes. Slope-maximum particle-size products and sample locations are shown on figure 114 and the basic data are listed in table 1.

Figure 114 shows that competence, approximated as before, decreases in a relatively uniform manner downfan. The data (table 1) show that the largest particle in the active channel occurred approximately at midfan in this instance. Estimated tractive force isopleths drawn for Paiute Chute would show a trend in the contours along the path of the active channel, because the values within the channel are greater than on the fan surface.

Two miniature deposits in Westgard Pass (pl. 8) were treated in a similar manner. Despite their small size, each contains a pronounced active channel, and estimated tractive-force data were obtained. Values along the active channels of these small fans are shown on figures 115 and 116, respectively, and the basic data are listed in table 1. Sampling stations were 5 feet apart, and slopes were measured by placing a Brunton compass on a board 1 foot long on the channel floors. One of the fans (fig. 116) was too small to allow a surface traverse, and the opportunity to chase a possible flood in another area interfered with completion of a surface traverse on the other (fig. 115).

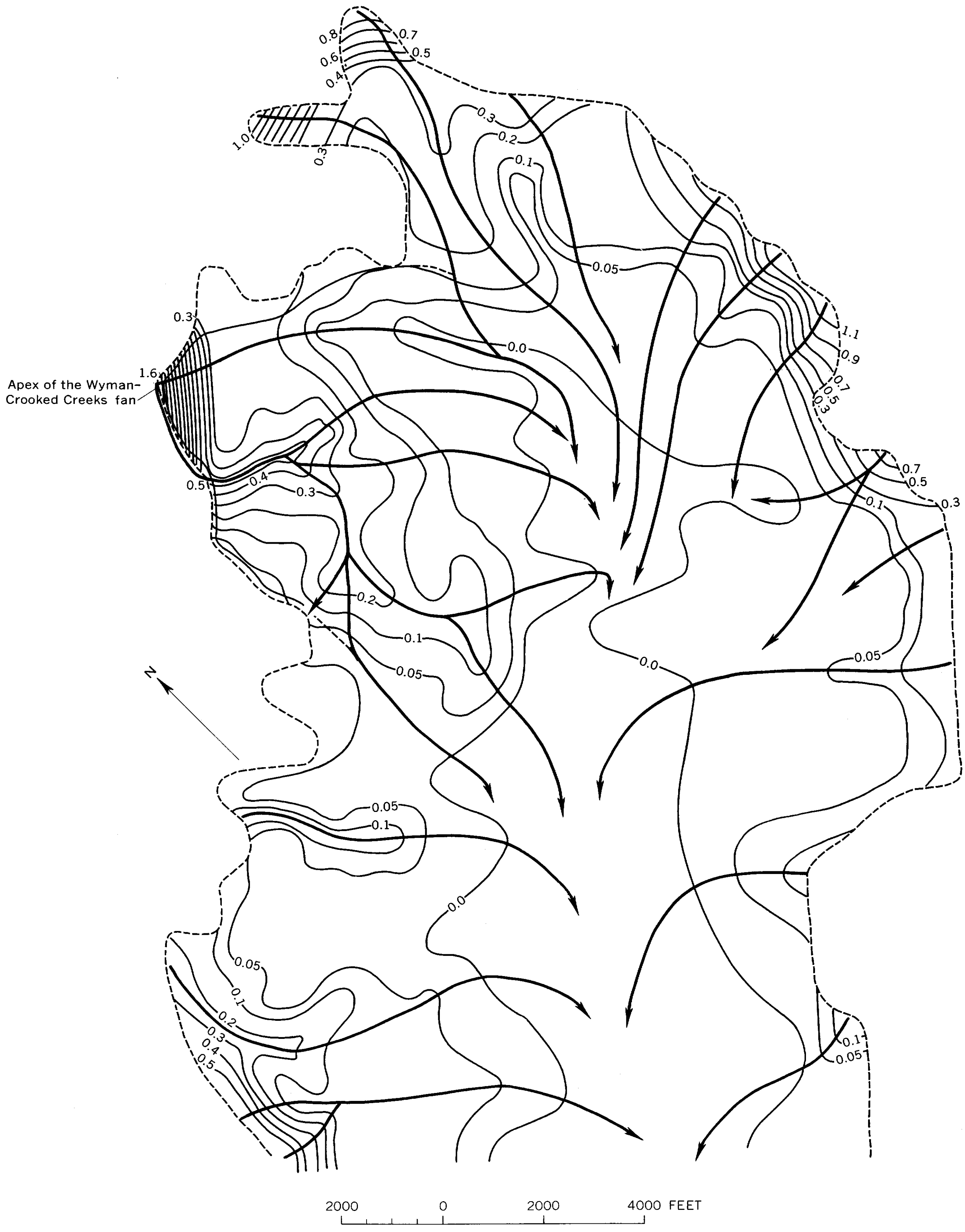

FIGURE 112.—Map showing estimated tractive force and inferred sediment transport paths in the north end of Deep Springs Valley. Isopleth values are slope-maximum particle-size products. Arrows indicate inferred paths of sediment transport. Dashed line represents approximate bedrock basin boundary.

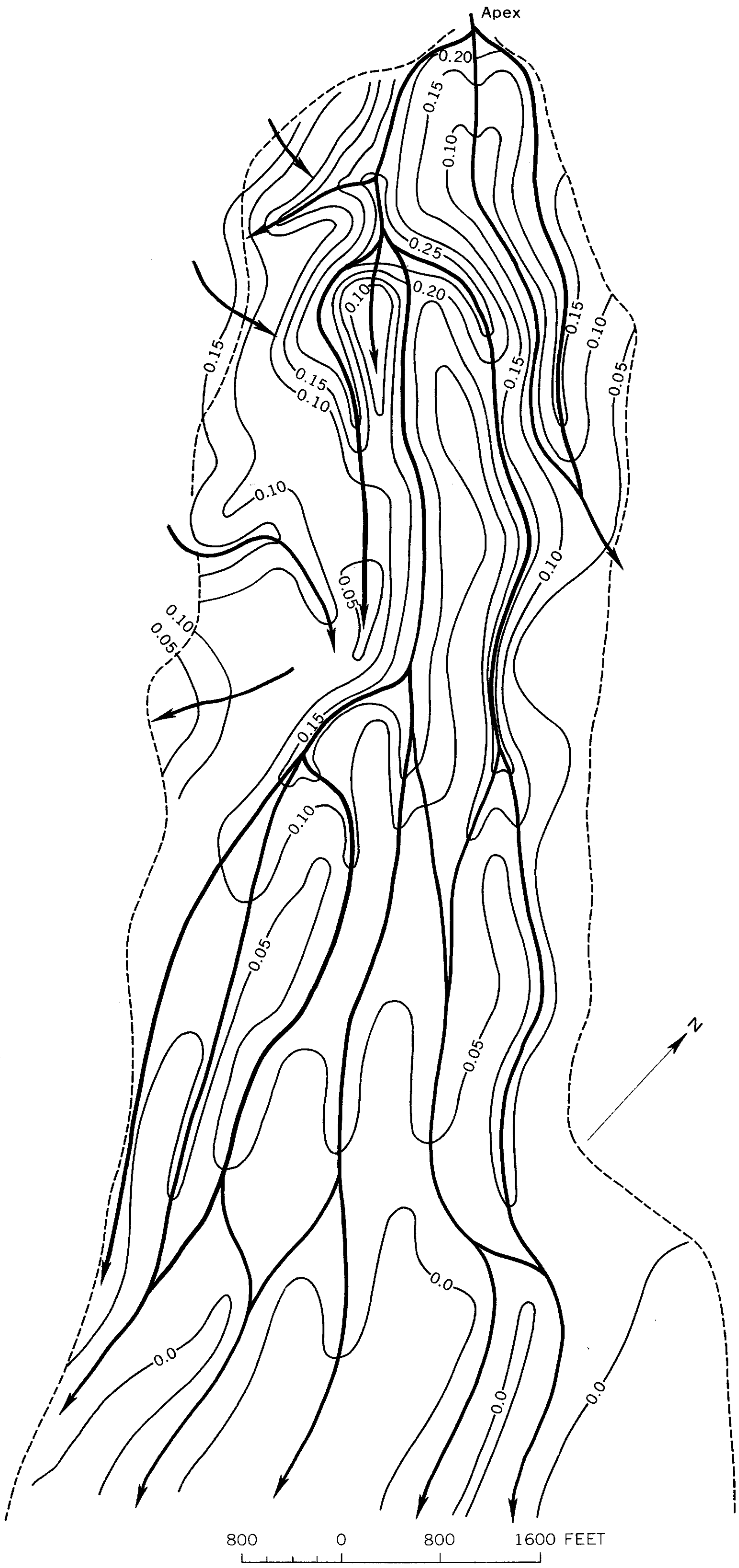

FIGURE 113.—Map showing estimated tractive force and inferred sediment transport paths on the Antelope Springs fan. Isopleth values are slope-maximum particle-size products. Arrows indicate inferred paths of sediment transport. Dashed line represents approximate fan boundary.

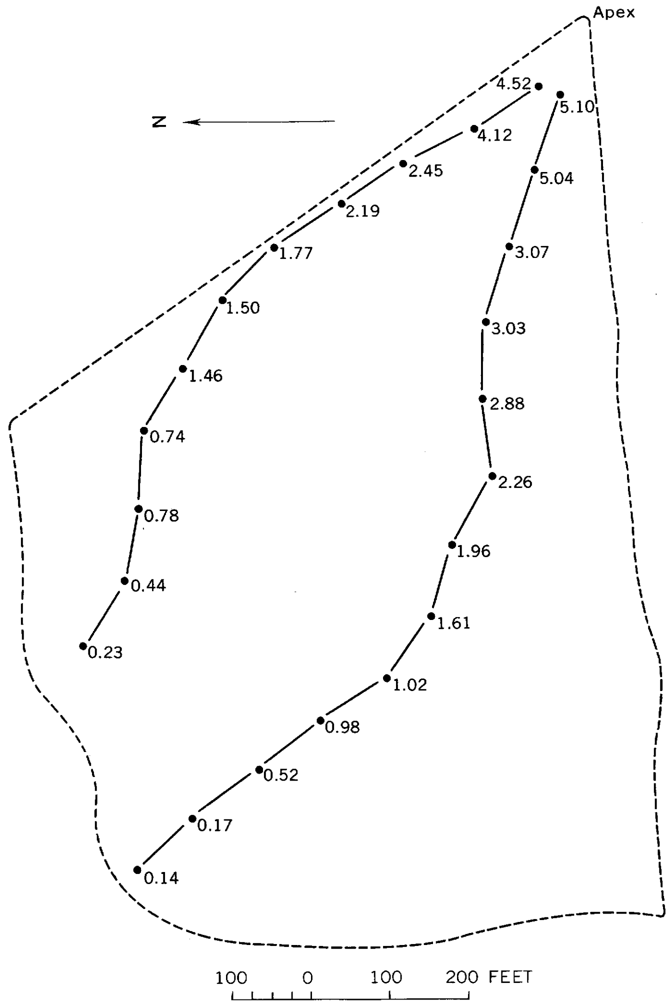

FIGURE 114.—Map showing the distribution of estimated tractive force on Paiute Chute. Values at the left were obtained from the fan surface at the indicated sampling stations; values at the right were obtained within the active channel. It is apparent that values in the channel are greater than on the surface. If estimated tractive-force isopleths were drawn, they would be convex downfan over the active channel.

The results show, however, that the values approximating competence again decrease downfan with greater regularity than does maximum particle size.

The evidence cited suggests that orthogonals drawn through estimated tractive-force isopleths delineate paths of sediment transport. Those orthogonals drawn through dominant trends in the contours coincide with active channels on alluvial fans, which are the most recent paths of sediment transport.

CHANNELS ON ALLUVIAL FANS

All the fans observed, both in Deep Springs Valley and in the neighboring basins noted above, contain channels that extend from trunk canyons far above their apex region to the middle or lower reaches of the fans. Channels are invariably entrenched and depths in the apex regions of the fans range from 20 to 200 feet below the surface. The sediment transported by modern floods is confined within these channels and can be readily distinguished in the field because of the color contrast with surface material. Regardless of actual color, the recent sediment is of lighter hue because weathering stain is absent; not only the more obvious channel on the Antelope Springs fan but parts of channels on adjacent fans can be distinguished by color contrast. (See fig. 85.)

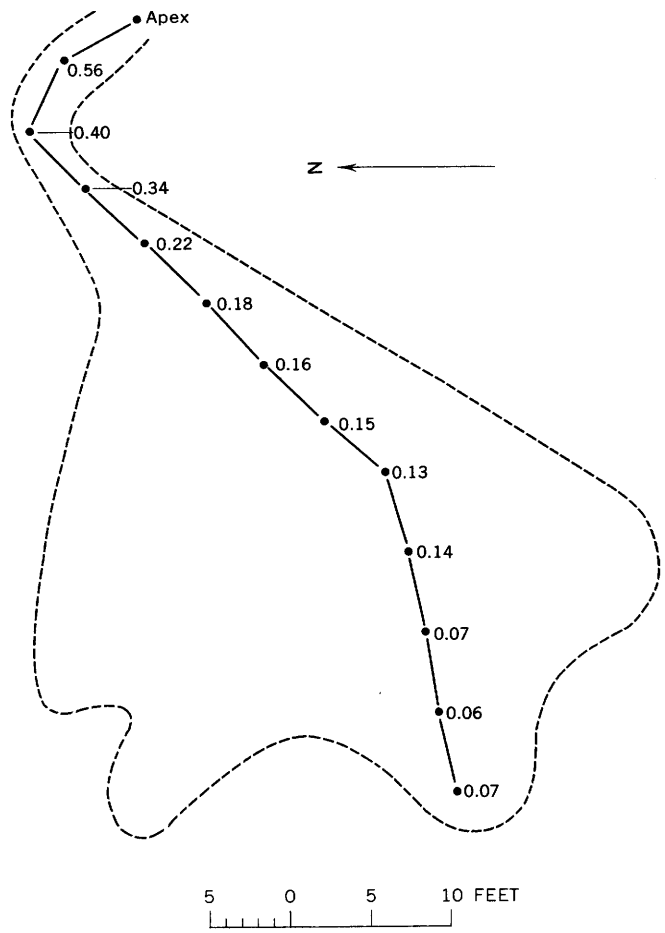

FIGURE 115.—Map showing the distribution of estimated tractive force in the channel of a miniature fan in Westgard Pass. Although the values fluctuate slightly, they decrease downfan far more consistently than maximum particle size.

Because recent sediment transport has occurred along entrenched channels, no material has been added to the fan surfaces "at the mountain front" on any fan. From a purely qualitative viewpoint it is impossible to observe this and conceive that this state of affairs has always obtained; the modern process must differ in kind from that of the past. Further visual evidence is provided by the terraces that occur within the trunk canyons above fan apexes. These extend back into the mountains over distances on the order of miles and

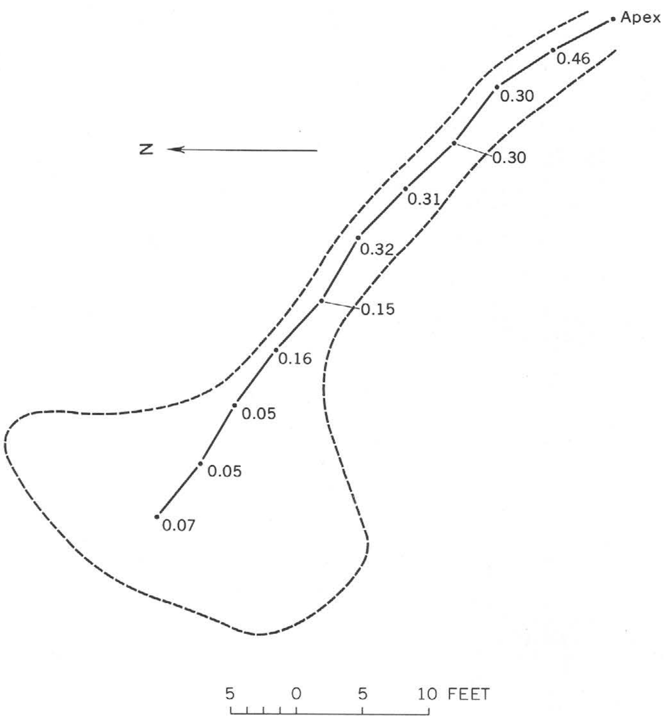

FIGURE 116.—Distribution of estimated tractive force in the channel of a second miniature fan in Westgard Pass. Although the values fluctuate slightly, they decrease downfan far more consistently than maximum particle size.

FIGURE 117.—View of the Antelope Springs channel, 1.5 miles above the mountain front. Note the terrace at *A*, about 8 feet above the bed along the right bank of the channel. This represents the former level of aggradation within the catchment area; it is continuous with the fan surface, below the mountain front. The heterogeneous character of the channel sediments and the occurrence of uprooted trees and shrubs are typical of the active channel.

FIGURE 118.—View of a low terrace in the Antelope Springs channel, 1 mile above the mountain front. Note that bedrock has been exhumed by channel cutting and that the part of the dipping strata below the terrace level is lighter in color than that above, which is weathered. The height of the low terrace is about 2.5 feet.

in some instances to the divide areas. The low terraces shown on figures 117 and 118 are about 1 mile above the apex region of the Antelope Springs fan. In the general vicinity of the mountain front, the highest terrace level present is continuous with the fan surface below; this is true for all fans. The aerial view of the Antelope Springs apex region (fig. 119) illustrates this continuity. The higher of the two terrace levels that occur on the right side of the channel in the photograph is clearly identical with the level of the fan surface. The hanging fan seen at this point will be discussed below. The two levels on the right side of the channel are more evident in a ground view (fig. 120). The terrace relationships within the Antelope Springs drainage system are similar to those observed elsewhere in the Basin and Range province by the writer, and these relations suggest that at some former time aggradation took place within catchment areas as well as on the fan surfaces. Degradation has occurred since that time, however, and the loci of deposition have shifted far downfan.

The active channel on the Antelope Springs fan (pl. 9) was mapped in its entirety and may be used as an example to substantiate this conclusion. The areas designated as zones of bankfull deposition represent the only parts of the fan where recent sediment has actually attained the level of the fan surface. At all points above the first bankfull zone, sediment has clearly been removed and not added. The low terrace

FIGURE 119.—Aerial view of the apex region of the Antelope Springs fan. Note the hanging fan (outlined) that is formed by the fill in the small tributary canyon to the right. Its surface is continuous with that of the Antelope Springs fan and with terraces farther upstream. The abandoned braided channels, visible on the fan surface in the foreground, are at a higher elevation than the floor of the active channel.

near the fan apex has a cover of sage and other vegetation, and the surface sediment exhibits a weathering stain. A view of this terrace in section near station 9 is shown in figure 121. A few hundred feet below station 9 and on the opposite side of the channel, this low terrace has been removed by erosion within the active channel. The fact that the active channel has indeed been deepened in the apex region is suggested by the abandoned channels (pl. 9) with floors at the level of the low terrace. Other abandoned channels, farther downfan, are visible in the aerial view of figure 119. It is apparent that the floors of all the abandoned channels, although below the fan surface, are at a higher elevation than is the floor of the active channel. This can also be seen on the Wyman-Crooked Creeks fan. The aerial view (fig. 87) shows several meandering and braided channels that are abandoned; the active channel is incised to a depth of about 40 feet below their floors. The ground view (fig. 122) indicates that the area occupied by the abandoned channels may again represent the level of a low terrace, as on the Antelope Springs fan, because there is a higher accordant level that was not apparent from the air.

The deposition that occurs today on the Antelope Springs fan is restricted predominantly to those areas at and below the first zone of bankfull deposition (pl. 9). It does not occur in widespread sheets over this entire area, but in various transitory locations that depend upon the migration of braided channels. The aerial view of the fan (fig. 85) clearly shows the point at which extensive braiding begins; this point corresponds to the location of station 126 on plate 9. The various channels that head in the bankfull depositional zones may result from the high-water stages that follow

FIGURE 120.—View of the hanging fan above the Antelope Springs channel. The surface level of the hanging fan (*A*) is continuous with the higher of the 2 terraces that are present and is about 25 feet above the floor of the active channel, which terminates the fan. The low terrace is labeled *B*. The bedrock in the lower right corner of the photograph has been exhumed.

FIGURE 121.—View of the low terrace at station 9 in the Antelope Springs channel. Note the alternation of beds of coarse and fine sediment in the terrace, which is about 5 feet high. The largest particles on the floor of the active channel generally exceed the size of those present in the banks and on the fan surface in this area.

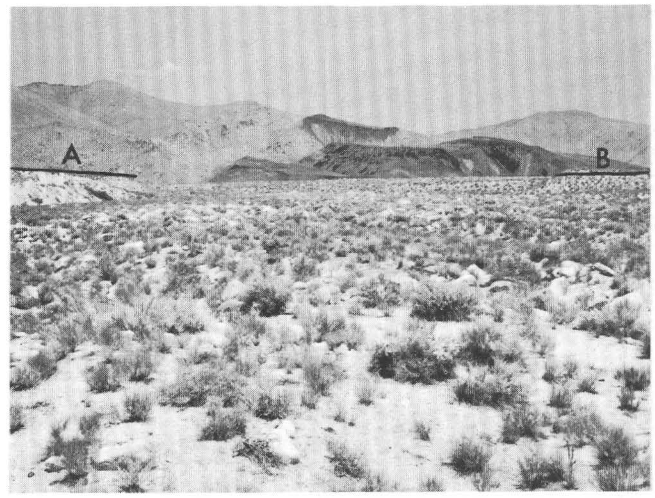

FIGURE 122.—Ground view of the Wyman-Crooked Creeks fan. The accordance of the surface of the ridge at *A* and the top of the low hill at *B* suggests that these may be remnants of a former continuous level. The active channel follows the base of the ridge at *A* and is incised to a depth of about 40 feet, below the general level seen in the foreground.

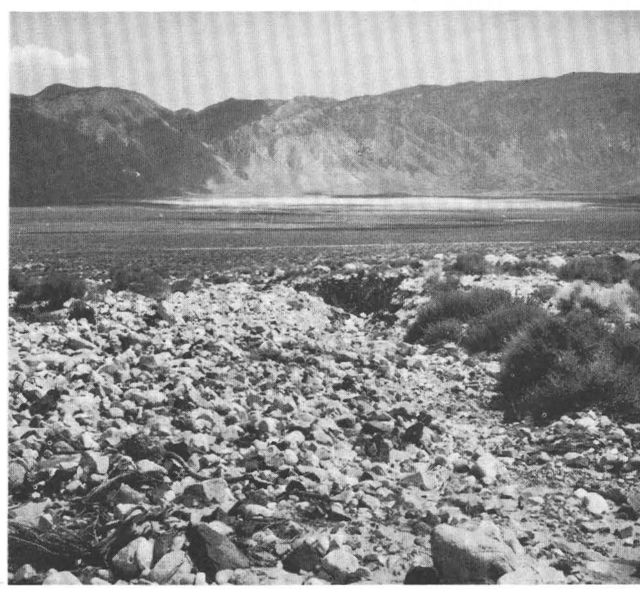

FIGURE 123.—View downfan at the bankfull deposition zone above station 126 on the Antelope Springs fan. There is a tendency toward convexity of profile in this region. The dissecting channel in the center of the photograph heads abruptly in the gravels, which are at, or slightly above, the elevation of the fan surface.

on the heels of mudflows (figs. 110, 111) and may be enlarged by seasonal runoff. Valley precipitation, as previously discussed, is probably insufficient to produce these features. The view downfan at station 126 (figs. 123, 124) shows the abrupt occurrence and rapid enlargement of one such channel. Two other channels, heading in the bankfull deposition zone just above the highway (pl. 9), are shown at the point of contact with the gravels in figures 125 and 126. The gravels are commonly openwork to considerable depth. Relatively clear water, flowing across and through these depositional sites, can account for both the winnowing of finer sediment and the fluctuation in mean size of the granule-to-clay fraction (fig. 102) noted above.

The last zone of bankfull deposition (fig. 127) is opposite station 318 (pl. 9) and clearly indicates that coarse debris has been transported far downfan in recent time. It is probable that deposition in the mid-

FIGURE 124.—Closeup view of the channel heading in the bankfull deposition zone at station 126. Note the rapid enlargement of the channel over a short distance and the convexity of profile.

FIGURE 125.—View of gravels, winnowed of fine sediment, on the Antelope Springs fan. The largest particle visible is about 4 feet in intermediate diameter. Note the poor sorting of the gravels and the abundance of finer sediment in the foreground, suggesting winnowing.

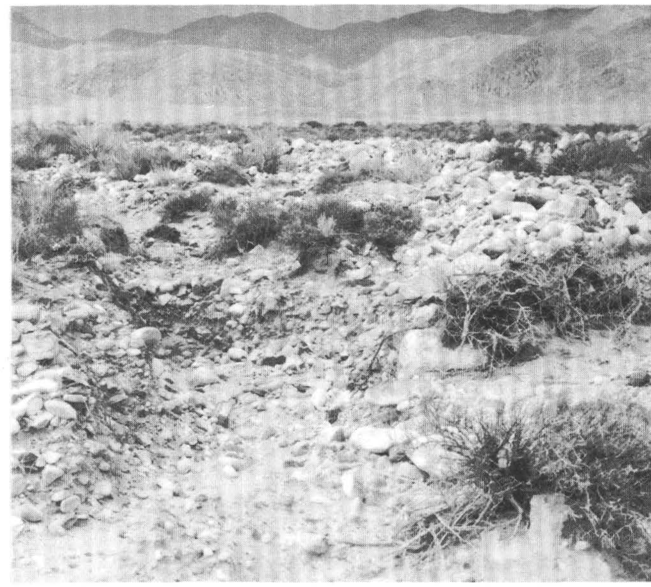

FIGURE 126.—View of a small plunge pool at the contact between a dissecting channel and the bankfull deposition zone above the highway. An abrupt break in profile at the contact is typical; here the plunge pool is about 2 feet deep. The fresh light-colored debris in the right background is slightly above the level of the fan surface.

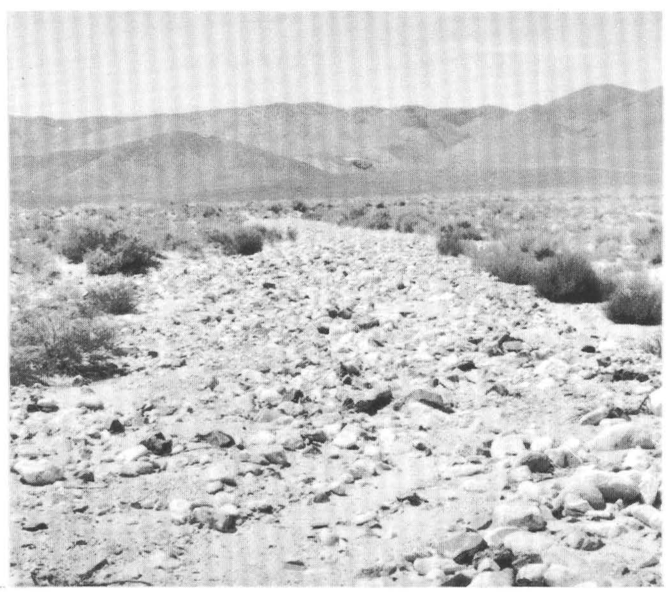

FIGURE 127.—View of the bankfull deposition zone near station 318 in the Antelope Springs channel. This fresh deposit is slightly above the general surface level. It represents sedimentation at the lower limits of the fan that tends to build it outward, into the valley proper.

dle and lower reaches of the Antelope Springs fan and degradation elsewhere is typical of most fans in the region today. This fan is entrenched only to a small extent in comparison with larger fans that contain channels as deep as 200 feet; and yet, abandoned channels and terrace levels occur as described. If the catchment and apex region has been subjected to degradation and if deposition occurs predominantly below midfan today then this fan is not being built up by modern processes in a manner suggested by the usual view of fan building. The question to be answered is "Why has trenching of fans occurred between the divide areas and midfan?"

CAUSE OF TRENCHING

The occurrence of trenches in alluvial fans is a fact noted by many authors, certain of whom have discussed the results produced by trenching rather than its cause. Eckis (1928), for example, thought that a geomorphic cycle was involved and that the onset of trenching signaled the beginning of the end for the fans. Buwalda (1951) considered the more effective transport of sediment to be expected, arguing that the trenches would act like flumes and funnel material downfan. Many authors have offered the view that deposition will occur where the trenches end and the width: depth ratio increases. This truth has little bearing upon the cause of trenching of alluvial fans, however. Two possible causes will be considered here, tectonic movements and variation in tractive force.

TECTONIC EXPLANATION

Tectonic explanations for a variety of phenomena have been invoked in local studies within the Basin and Range province and elsewhere. The trenching of fans referred to in this report, however, is extremely widespread and requires a regional rather than local explanation. If uplift induces downcutting, then approximately synchronous movements of most of the ranges in the Basin and Range province must be invoked to explain the trenching of fans. Even where uplift can be demonstrated for an individual range, however, several difficulties emerge from direct cause-and-effect correlation of uplift and the trenching of fans.

The tectonic hypothesis is predicated on the concept that uplift will increase channel slopes and the velocity of flow, thereby inducing erosion or degradation. The actual increase in slope that can occur from a given uplift is seldom cited, however. If the White Mountains (pl 8.) are used as an example, the calculation is instructive.

The range is about 20 miles wide, and the relief is roughly 5,000 feet. An approximation to the overall slope, from the divide to a fan apex, is therefore 0.094 foot per foot. The actual channel slope will exceed this value near the divide but will be less than this value near the mountain front. If an absolute uplift of 1,000 feet occurred instantaneously, this value would become 0.113, an increase of 0.019 foot per foot. This is a rather small increase in view of the extreme uplift conditions postulated. Epply (1961) has shown that even the most severe earthquakes with high Mercalli intensities are usually associated with throws along faults of no more than about 30 feet. The effect of the average uplift upon slope will therefore be extremely small.

Moreover, velocity and slope are related by the Manning equation, namely, $v=\frac{1.49}{n}\ R^{2/3}S^{1/2}$, where v is velocity, R is the hydraulic radius, and S is the slope. The exponents in this equation show that changes in the hydraulic radius, which is a function of discharge, have a greater effect upon velocity than changes in slope. In addition, it has been pointed out by many authors that velocities are not necessarily high in the mountains, where steep slopes occur. The mean velocity of rivers, in fact, tends to either increase or remain constant in a downstream direction (Leopold, 1962). This would be expected because drainage area and, consequently, discharge, both increase downstream.

It is true that local fault scarps can produce a nickpoint that will retreat through time, but aside from other considerations, this explanation for trenching would again require simultaneity of tectonic activity on a regional scale. It is probable, therefore, that the main role to be ascribed to tectonism in regard to trenching is the change induced in the pattern of precipitation and vegetation, both of which vary orographically and can affect the processes under discussion. Magnitudes of uplift cannot be directly equated with sedimentation or erosion. The cause of trenching will be further pursued by considering the implications of the tractive-force distribution.

TRACTIVE-FORCE EXPLANATION

The sediment transport orthogonals (figs. 112, 113), which coincide with fan channels, generally intersect estimated tractive-force isopleths where they are convex downfan. Because the estimated tractive-force values decrease downfan, this means that tractive force within the channels is greater than at adjacent points on the fan surfaces. This was shown to be true for Paiute Chute (fig. 114), for example. Because modern floods are confined within the channels, the tractive force associated with the modern process must be greater than that which accompanied transport in the past. If this is true, then one or more of the three component variables that comprise tractive force must be greater today than in the past.

Of the three component variables, it might appear that deviation of past and present slope would be simplest to ascertain. If an active channel is traced downfan, then a point will eventually be reached at which the slope of the channel intersects the fan surface; this is a zone of bankfull deposition (pl. 9). It would not, however, be correct to conclude that the slope today must be less than in the past, because the slope in question is that which obtained before deposition in these zones. The deposit is itself an indication of change in process. The information desired is whether change in slope altered tractive force and therefore initiated the trenching indirectly. In this sense the answer is indeterminate but field observations suggest that the slopes of the channels do not markedly differ from those of fan surfaces, except in the vicinity of zones of bankfull deposition.

The depth of flow may be greater today than in the past, if the depth of the confining channels or trenches is accepted as a criterion. This would imply greater discharge and a correspondingly greater tractive force in the channels. The fact that former lakes throughout the region are today playas, however, suggests that the discharge of modern floods is probably less than floods of the past. The modern floods, in fact, may be misfit (Dury, 1958) in relation to their confining trenches in the upper reaches of fans; evidence of overbank flooding in the vicinity of the mountain fronts is generally lacking. Moreover, if an increase in tractive force initiated trenching, then selection of depth of flow as the responsible variable would be equivalent to attributing both cause and effect to this factor.

The remaining component, specific weight of the medium, can be considered in terms of density. If the density of the average flow is greater today than in the past, then mudflows should be a more frequent mechanism of transport. Observations suggest that this may be true. It remains to be proved that sound reasons exist for postulating that a contrast in density characterizes the difference between the modern and former processes. The tractive-force hypothesis, however, can be summarized at this point. The implication of the maps of estimated tractive force is that the modern process is attended by a greater tractive force than that of the past. Greater tractive force provides a logical explanation for the cause of trenching; simply stated, the explanation is that the shearing stress on fan surfaces increased at some time in the past. The stress was produced during sediment transport, and each effective flood deepened the channels until the designation "trench" could more fittingly be applied; this is still true today. Although a combination of factors may have brought about the increase in tractive force, an increase in the density of flow, including buoyancy effects, is thought to be an important cause. The evidence for this suggestion will be considered in the following discussion.

CONTRAST BETWEEN THE MODERN AND FORMER PROCESSES

MUDFLOWS

Descriptions of mudflows, the magnitudes of the particles transported, and descriptions of attendant features such as levees, have been provided by Rickmers (1913), Pack (1923), Blackwelder (1928), Sharp (1942), Wooley (1946), Kesseli and Beaty (1959), and several others cited in this report. Such information is common knowledge today, but the important question is whether mudflows actually cut channels by reason of their high density and consequent high tractive force. Most reports convey the impression that mudflows serve only to build fans, but some information to the contrary does exist in the literature. This is worthy of stress because it is germane to any discussion of past and present processes.

A statement by Wooley (1946, p. 78) is revealing:

> The mudflow of 1938 scarred the surface of the fan in a course 60 to 125 feet wide from the apex down the crest of the slope. The courses of a number of earlier mudflows are marked by windrows or small ridges of alluvium, which remain along the edge of the flow as the channel is cut deeper.

And on page 80 of the same report he states:

> Although degradation by mudflows is not an uncommon natural process in the deserts and semiarid areas of the west, there is little information available as to the mechanics of mudflow operation.

In a recent paper treating erosion of the Wasatch Mountain front, Croft (1962) provides numerous examples and much pertinent data attesting to the fact that mudflows in recent time have caused much channel erosion. Two photographs taken from the literature and reproduced here serve to illustrate the point. Figure 128 shows one of the many trenches produced by the catastrophic flood of June 1957 in the basin of the Guil River, France; figure 129 shows the results of 1 of 100 mudflows that occurred in Grand Teton National Park, Wyo., in August 1941. These particular examples were chosen because neither area is semiarid. They therefore illustrate the fact that degradation in the upper reaches of the path of flow is a characteristic feature of mudflows. When they occur, erosion results whether the climate is humid or arid. If these two examples are compared with the view of Paiute Chute (fig. 88), it can be seen that the features of all three are identical.

The views suggest that debris is stripped from the upper reaches of each area and is transported to and deposited on the lower slopes. This is precisely the relationship observed on most alluvial fans in the Basin and Range province today. The catchment areas up to the divides, and the apex regions of the fans, are source areas for mudflows; deposition occurs below midfan.

The actual mechanics of mudflows is poorly understood, and the hypothesis set forth in this report, namely, that mudflows produce degradation primarily by reason of their greater tractive force, is based largely upon inference and deduction. As pointed out in the American Society of Civil Engineers, Task Force

FIGURE 128.—View of a channel cut by a mudflow in the Gull Basin, France (after Tricart, 1961, fig. 2). The region is fairly humid, but the results of the mudflow are comparable with the features of Paiute Chute (fig. 88); addition of sediment largely occurs in the lower reaches of the path of flow.

Committee on Hydromechanics report (1963), however, there can be no question that the effect of dispersions of sediment is to increase viscosity, which in turn, alters the bed form because of a reduction in fall velocity of the bed material. The report cites Le Conte (1896) on page 125 as follows:

> The results show conclusively the powerful influence of viscosity due to sediment in suspension. A channel lined with cobblestones will stand a clear water velocity of 6 to 8 feet per second, but when it is followed by a more viscous stream heavily charged with material in suspension, although at a reduced velocity of only 4 to 6 feet per second, the cobblestones will be at once picked up and transported with the flow.

Although the flows discussed by Le Conte are not the equivalent of modern mudflows, it is suggested that density flows can erode the floors of channels by the removal of bed material that is too large to be transported by clear water. The fact that there is no evidence to the contrary, coupled with the several observations previously cited, implies that greater tractive force is associated with the modern process as a result of an increase in the average density of flow.

THE FORMER PROCESS

If the onset of trenching was initiated by density flows of higher tractive force, such as the relatively infrequent modern mudflows, then the inference to be drawn from the tractive-force hypothesis is that sediment transport was formerly achieved by a medium of lesser density. Given flows of lesser density than a mudflow but of necessarily greater density than clear water, and the absence of the deep confining channels of today, it appears probable that the flows of the former process must have spread laterally below the mountain front and aggraded the apex regions that are being eroded today. It is suggested, therefore, that the former process was more nearly akin to sheet floods than to the channelized density flows of today. Discharge was probably greater in the past; but because of lesser density and shallow depth of flow, the resultant tractive force was less.

If the former flows were of lesser density, then the water: sediment ratios must have been greater. This accords with the concept that greater precipitation prevailed in the area during several former time intervals. From this point of view the relative scarcity of clay and abundance of silt in nearly all the samples studied is worthy of note.

One might argue that the surface samples have been winnowed of clays by local runoff, but clearly, the basin surface today represents a time plane. If these samples have been winnowed, then samples obtained from fan cross sections should also contain little clay. Bull (1960) has examined such sections in the attempt to find criteria that will aid in distinguishing among mudflow deposits, water-laid deposits, and intermediate types. Although clay content is probably not a diagnostic feature of deposits in other areas, it is instructive to note that the main criterion used by Bull to distinguish ancient mudflow deposits in Fresno County, Calif., is a high clay content. He gives a value of about 30 percent for the samples studied. The Owens Valley mudflow, however, is by no means an atypical example of mudflows of today. The nature of the flow can be seen from the photographs of the upper reach (figs. 110, 111) as well as of the lower reach (figs. 130, 131). Samples from this mudflow contained 2.41 and 4.22 percent clay; these values pertain to the granule-to-clay fraction alone and therefore are maxima for the flow.

Sample locations in the active channel of Paiute Chute are shown in figures 132 and 133. The matrix of a recent flow is still present in the channel banks or levees; data on clay content are therefore pertinent. The percentage of clay, again in the granule to clay fraction, is 2.10 at the apex and 3.00 at midfan.

FIGURE 129.—View of the aftermath of a mudflow in the Grand Tetons, Wyo. (after Fryxell and Horberg, 1943, pl. 2, fig. 1). Note the similarity of the chutes that occur here and those shown in figures 88 and 128. The large deposit in the foreground is derived from preexisting talus deposits that were trenched by the mudflow.

Similar low values were obtained from the mudflow of Wrightwood, Calif. (Sharp and Nobles, 1953); little clay was present.

These data suggest that former transport conditions may have differed from those of today by reason of greater clay content. One might argue that a greater clay content in former flows implies still greater density of flow, but it should be recognized that the rate of production of clays is, in good part, a function of precipitation. Little clay will be produced if precipitation is deficient; from this point of view, if more clay was associated with the former process, then this supports the inference that water: sediment ratios were formerly greater. It should also be noted that given a greater abundance of clay in the surface sediments of fans, cohesion and resistance to trenching will be increased.

A second factor suggesting that fluctuation in regime may have occurred is the contrast between the organic content of mudflows today and that of older deposits. Despite the typical dark-brown to deep black color of the Owens Valley mudflow (figs. 110, 111, 130, 131), the organic content was quite low. Weight percentages are 1.89 and 2.80 in the granule to clay fraction for the 2 samples obtained. Wyman Canyon is choked with vegetation and contains a perennial stream. Predictably, the thin atypical Wyman mudflow contained a much higher percentage (14.05) of organic matter. Eardley and Gvosdetsky (1960) reported evidence of about six thin mudflow deposits in core samples from the Great Salt Lake. They state that a high humus content is present in these old flows but do not give percentages; such data are extremely rare. Croft (1962) reported the organic content of soils developed on old and recent mudflow deposits, but these values are not comparable to the content of mudflows. The paucity of data permits only the conclusion that the organic content of modern flows is not great. This results, of course, from the fact that vegetation is not abundant today. It is probable that both clay and vegetation were formerly more abundant, because both require the same type of climatic regime. Higher precipitation is not the only characteristic of the former regime, however; the frequency of precipitation must also be considered.

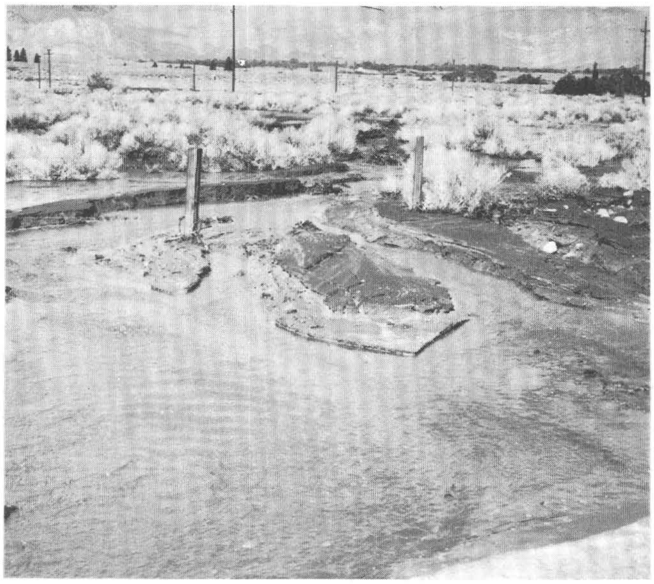

FIGURE 130.—View of the lower reach of the Owens Valley mudflow showing clear-water dissection. Although the deposit is undergoing dissection, it is apparent that sedimentation has occurred across a wide flat area at the lower boundary of the fan. The fenceposts shown parallel the main highway through the valley, which is about 8 miles below the mountain front of the Sierra Nevada at this point. Note the dark color of the deposit.

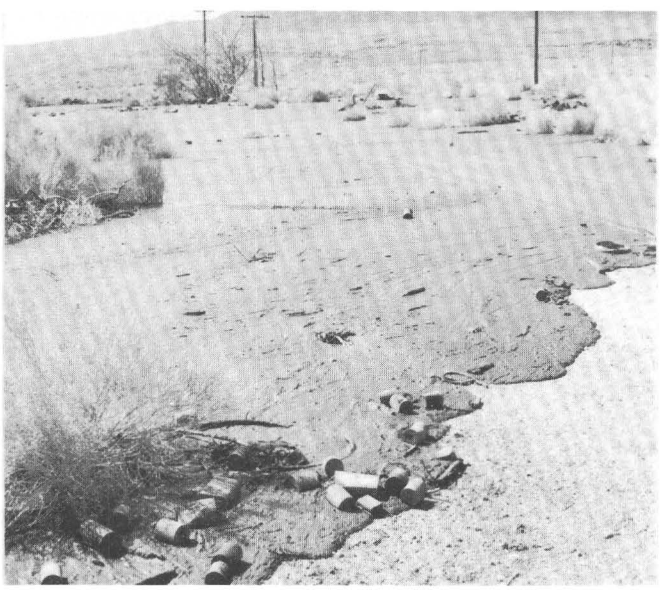

FIGURE 131.—View of the lower reach of the Owens Valley mudflow showing floating debris on the thin deposit. The flow was quite dense and viscous in this area (approx. 0.25 mile south of the clear-water flow shown in fig. 130), and in addition to the cans seen in the foreground, many cobble-size particles were also rolled and carried across a wide area of low slope.

FIGURE 132.—View of the apex region of the active channel in Paiute Chute. The intermediate axis of the large boulder to the right of center is 10.4 feet in length. The divide of the small catchment area extends along the ridge summit shown. Fine sediment is rather scarce.

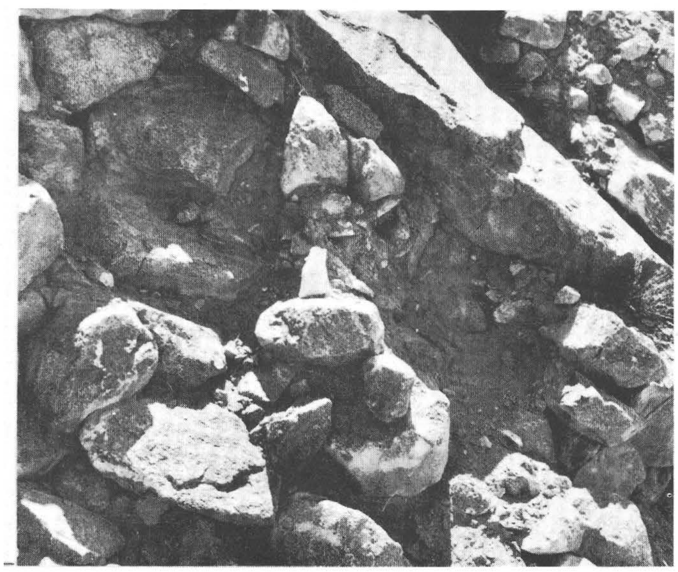

FIGURE 133.—View of the fine matrix in the levee wall along the active channel of Paiute Chute. The occurrence of the fine matrix surrounding the tightly wedged granitic boulders indicates that a mudflow produced the deposit and that Paiute Chute is not of rockslide or similar origin. The sample bag in the center of the photograph is 4 by 6 inches.

FREQUENCY OF PRECIPITATION

In semiarid or arid regions much of the annual precipitation today falls in few and infrequent storms. The slopes are largely bare of vegetation, little clay is present, and resistance to erosion is low. This combination of conditions favors the occurrence of mudflows. Given a more equitable distribution, as well as an absolute increase of annual precipitation, this situation will be reversed, and mudflows will become less frequent. The precipitation that accompanied the mudflows in the Grand Tetons (Fryxell and Horberg, 1943) represented a great departure from the norm for that area. Rainfall throughout the summer of 1941 was above average and was higher in each succeeding month; it was five times the average during the month of August, when the flows occurred. During June, July, and August no more than 3 successive days passed without recorded precipitation. The mudflows in the Guil Basin (Tricart, 1961) were also brought on by uncommonly high precipitation. Heavy rainfall over a period of 48 hours, entirely without precedent in the region, produced the flows. It cannot be claimed, therefore, that mudflows can only occur where the frequency of precipitation is low, but observations suggest that they are much more common in such areas. Variations in the frequency of precipitation are known to have occurred in semiarid areas (Leopold, 1951) and, by implication, in arid areas as well. It is thought that such changes in the frequency of precipitation can account for the prevalence of a former flow regime that was characterized by higher clay content, organic content, and water : sediment ratios.

Although the requirement of greater frequency of precipitation is often stated during discussions of former processes, it is difficult to prove on the basis of field evidence. One can, of course, easily argue in the case at hand that if the frequency of precipitation remained at its present value in the Great Basin while annual precipitation increased, then the only result would be more intense cloudbursts and still greater degradation by mudflows. It is thought, however, that the hanging fan shown in figures 119 and 120 does constitute field evidence that a greater frequency of precipitation must formerly have prevailed.

The fill in the small tributary canyon to the right of the main channel represents the level of aggradation that was attained throughout the Antelope Springs catchment area and by the fan proper. This is the high terrace level. The fill has been truncated by erosion within the main channel, and it may best be designated a hanging fan. This same phenomenon was observed in several small canyons that are tributary to the lower reaches of Wyman Canyon and in basins other than Deep Springs Valley. The occurrence of these hanging fans suggests that precipitation and sediment transport no longer occur in every canyon within the catchment areas. The infrequent cloudbursts of today may follow preferred tracks, as noted previously, and most of the precipitation will occur in the divide areas. The mudflows and the resultant trenches occur primarily in the trunk canyons and in the larger tributary canyons that head near divide areas. The precipitation pattern that accompanied uniform aggradation in the catchment areas must have differed from that which pertains today. Precipitation was certainly widespread rather than localized and, hence, was probably more frequent in occurrence.

The relationship discussed above may well be relevant to the problem posed by many hanging valleys that have been interpreted in terms of tectonism. Uplift of a range may result in more frequent precipitation in the divide areas and a decreased probability of precipitation in certain tributary valleys. A contrast in both discharge and frequency of floods could account for greater erosion within the main valleys and the bypassing of tributary valleys. For reasons cited previously, this explanation for hanging valleys would appear much more reasonable than any assumed linear relationship between uplift and downcutting.

PROCESS RATES

A charcoal pod was found at station 400 in the Antelope Springs channel (pl. 9). A general view of the channel in this area is shown in figure 134. The pod occurred 6 feet above the channel floor in the right bank, which is 12 feet high at this point (fig. 135). The presence of fire-blackened stones directly above the charcoal indicated that the pod was a fire hearth. A radiocarbon date of 1,380 ±250 years B.P. was obtained for a sample from this hearth. The field relationship therefore indicates that 6 feet of deposition and 12 feet of erosion have occurred since this date.

FIGURE 134.—View of the Antelope Springs channel at station 400. The view is downchannel from the fire-hearth site. Note the fresh character of the cobbles to the left of the meandering channel, indicating that modern transport has extended far out into the valley. The sage ranges from about 1 to 3 feet in height.

Any estimate of the process rates involved will depend upon the establishment of some arbitrary ratio between the rate of deposition and the rate of erosion. If it is assumed that the rate of erosion is equal to the rate of deposition, then 18 feet of either process has occurred in the last 1,130–1,630 years. This yields respective rates of 0.016 and 0.011 foot per year. If it is assumed that the rate of deposition was only half as great as the rate of erosion, which may be more reasonable at this location, then the equivalent of 24 feet of process has occurred during the same period. The respective rates in this case would be 0.021 and 0.015 foot per year.

FIGURE 135.—View of the fire-hearth site in the Antelope Springs channel. The channel bank is 12 feet high at this point, and the bottom of the hearth is 6 feet below the top of the bank. Note the fire-blackened stone resting directly upon the charcoal pod, thus indicating that it is in fact a fire hearth.

These rates, ranging from about 1 to 2 feet per hundred years, should properly be applied only to events in the vicinity of the sample. Some estimate of the time of the last expansion of Deep Springs Lake can therefore be made, because the linear ridges that indicate this expansion (fig. 85) are in this general area. If it is assumed that the 6 feet of deposition in this area coincided with the expansion of the lake, then the linear ridges on the basin surface are 300–600 years younger than the fire hearth, depending upon the rate of deposition chosen. The high stand represented by the ridges occurred, therefore, between 700 and 1000 B.P., if the radiocarbon date is correct. Langbein (1961) has estimated the date of the last expansion of several lakes based upon the time represented by the total dissolved salt content. The value obtained for Owens Lake was 1,700 years, and a few other lakes expanded still more recently. The estimated date of the last expansion of Deep Springs Lake is therefore not unreasonable.

If the relatively minor climatic changes of Recent time could bring about the expansion and contraction of lakes and trenching to a depth of 12 feet at the fire-hearth location discussed above, then the much greater changes that marked the Pleistocene Epoch in this area may have coincided with the initiation of trenching in catchment and upper fan areas. It is worthy of note that an erosion rate of about 3 feet per hundred years would suffice to produce even the deepest trenches that occur in the region.

To obtain some future data on process rates in Deep Springs Valley, chains were installed in the channel bed at five locations in the Antelope Springs channel (pl. 9). At each section, particles in the channel were painted in order that some information on sediment transport might also be obtained in the course of the monitoring program. At sections *A–A′*, *B–B′*, and *C–C′*, entire gravel bars were painted; the first mentioned is shown in figure 136. Section *C–C′* is in the vicinity of the fire hearth. At sections *D–D′* and *E–E′* no gravel bars were present, and particles that occurred on the channel floor were alined perpendicular to the

FIGURE 136.—View of the painted gravel bar at section *A-A′* in the Antelope Springs channel. The bar is about 70 feet long and consists mainly of particles as large as cobbles; few larger particles occur.

direction of flow. The information gained from this program over a period of years will not only provide data on modern process rates, but it should also serve to confirm the contention that trenching is the dominant long-term process in the upper fan reaches today.

FORMATION OF THE ALLUVIAL FANS

The formation of alluvial fans has been discussed ever since these striking sedimentary deposits were first noted during the early western exploratory surveys. The opinions expressed to explain the formation of fans may be conveniently grouped into three general categories, namely (1) evolutionary, (2) equilibrium, and (3) climatic hypotheses. Although the writer favors the climatic hypothesis, each holds some measure of truth, and because the truth often occupies middle ground in such problems, each hypothesis is treated here.

THE EVOLUTIONARY HYPOTHESIS

The concept that all landforms result from a geographic cycle that proceeds through the successive stages of youth, maturity, and old age is today commonly termed "Davisian philosophy." The specific choice of anthropomorphic stage names was probably influenced by the publication of the doctrine of evolution in the 19th century; the concept can therefore be appropriately designated an evolutionary hypothesis. Davis (1905) applied this concept to alluvial fans by emphasizing the importance of stage and thereby denigrating process. He suggested that alluvial fans would form in basins adjacent to uplifted mountain ranges in desert regions. The stage of youth was thought to be indicated by V-shaped canyons in the mountains and by small but rapidly growing fans. As the mountains wore down, Davis reasoned, the fans would continue to enlarge, and thus would fill the basin with sediment, and in the final stage—old age—the mountain remnants would be buried beneath a continuous cover of alluvium.

The inevitable reaction to one of the several failings of the concept of cycles, namely to treat processes and thereby gain some insight into the mode of formation of any given landform, set in with the advent of quantitative geomorphology. The modern view of the geographic cycle has, to a large degree, been summarized by Chorley (1962), who equates Davisian philosophy to closed-system thinking or the single-cause school of thought. Davis was quite correct in two important particulars, however.

First, despite the writer's disenchantment with tectonic explanations for magnitudes of sedimentation and erosion, as previously explained, it cannot be denied that slope and discharge are both intimately related to the production of sediment. An infinite discharge over a zero slope will have little more effect than zero discharge down a vertical rock wall. From this point of view, there can be no question that some uplift must first occur to provide the requisite differential relief for the formation of an alluvial fan at some initial time t_o.

Second, unless an infinite duration of uplift is postulated, then a given basin must completely fill with sediment at some finite time t_f. It is therefore reasonable to assume that at time t_f some mountain remnants will indeed be buried beneath an alluvial cover. The extent of this cover will depend upon coincidence, or its absence, of time t_f in each of many basins. The peneplain conceived by Davis (1889) will probably not result, but it is undeniable that old age, as Davis defined it, must ultimately be attained by any given basin.

The evolutionary hypothesis, admittedly based upon deduction, provides a reasonable description of alluvial fans in their formative and ultimate stages of development or morphology. Largely omitted, however, is consideration of the time between t_o and t_f and of the processes involved in fan formation.

THE EQUILIBRIUM HYPOTHESIS

The equilibrium hypothesis relates the morphology of a landform, or Davisian stage, with the processes acting upon it in terms of dynamic systems in which mass and energy are considered as functions of time. The two general systems defined by Von Bertalanffy (1950) are closed systems, in which materials and energy cannot be exchanged beyond the confines of a well-defined boundary, and open systems, in which materials, energy, or both can be exchanged with outside environments. As pointed out by Strahler (1952), most landforms represent open dynamic systems that tend toward equilibrium or a steady state. This view is widely accepted today (Leopold and Langbein, 1962; Chorley, 1962).

Although the equilibrium hypothesis is basically valid, certain difficulties of both application and interpretation arise when considering alluvial fans. An assumption must be made that a balance exists between erosion of the mountains and deposition in an adjacent basin. This is obviously reasonable, but if alluvial fans are in dynamic equilibrium, then a second, more particular assumption must be made, namely that the rates of sedimentation and erosion on a given fan are equal. Unless these rates are equal, then a fan will either grow or diminish in size as a function of the difference between the two rates. Denny (1965) contends that these rates are equal, or nearly so, and that alluvial fans are therefore in dynamic equilibrium with the processes acting upon them. Aside from correlations between planimetric measurements of source areas and their respective fans, which are somewhat suspect for reasons cited above (p. 133), little evidence exists to substantiate this contention. Rates of sedimentation and erosion on alluvial fans are not known. The amount of local runoff required to remove sediment from fans in order to balance the sediment added is also not known. Local runoff, however, can only be produced by either flows from the mountains or precipitation in the valley proper. It is not obvious that either of these sources will suffice to achieve the required balance; mountain floods must overcome seepage into the permeable alluvium, and local precipitation is not great. Because of orographic effects, the precipitation on most basin floors, and presumably on the fans as well, is very low, ranging from less than 2 inches per year in Death Valley to about 5 inches per year in basins at higher elevations. Moreover, the fact that highly erodible sediments, which represent former lake levels, have existed for thousands of years on the lower reaches of many fans does not support the view that a balance has been achieved or dynamic equilibrium attained.

The primary problem involved in the application of the equilibrium hypothesis, therefore, is to demonstrate conclusively that dynamic equilibrium has indeed been attained, rather than to merely assume that this is true. Even if one grants that alluvial fans have attained a steady state, however, a second problem will arise, namely the proper interpretation of this information.

Possible misinterpretation can be illustrated by consideration of the formation of playas. If their widespread occurrence, time contemporaneity, and the existence of strand lines at higher elevations are ignored, then it may be erroneously concluded that climatic change or other cyclic phenomena are not pertinent and that the interrelationship of their morphology with present processes is sufficient for an understanding of their formation.

Morphology and processes may be clearly interrelated, but the demonstration of dynamic equilibrium at any instant in time, even if conclusively proven, does not preclude the possibility that changes in both the intensities and types of processes may have occurred through time. Moreover, unlike water bodies, many landforms adjust to such changes very slowly, and although they tend toward equilibrium with modern processes, they may in fact be far removed from a steady state today. Leopold and Langbein (1962) state that it is improbable that time periods as long as geologic epochs could elapse before the attainment of equilibrium by fluvial processes in adjustable channels. Although the absolute duration of an epoch depends upon the geologic period considered, their meaning is clarified by a declaration of agreement with Hack (1960), who thought that no important time period is necessary. These statements do not, however, refute the fact that Pleistocene relicts that are thousands of years old exist today in all parts of the world. Hanging valleys, raised beaches, river terraces, dry waterfalls, the strand lines cited previously, and many other features attest to a lack of equilibrium with modern processes; in the absence of proof to the contrary, it is not unreasonable to suppose that some trenches in alluvial fans may be akin to the misfit river channels discussed by Dury (1958). Why, then, in the absence of any substantial proof should alluvial fans be thought to have attained dynamic equilibrium?

The concept of dynamic equilibrium of landforms represents a major advance beyond the application of Davisian principles. As previously stated, however, this does not mean that certain truths should be ignored merely because they are founded upon deduction. The equilibrium hypothesis requires blending with caution, lest a new generation of workers seek everywhere to demonstrate equilibrium as those before them sought cycles and peneplains. Dynamic equilibrium for any given landform must be conclusively proved, rather than simply claimed because the concept is in vogue, and if demonstrated to exist, then equilibrium conditions should not be cited as proof that the modern processes are necessarily identical with those of the past.

THE CLIMATIC HYPOTHESIS

The evidence suggesting that present climatic conditions are not identical with those of the past is legion. In response to changes in temperature, precipitation, evaporation rates, and the location of cyclonic storm tracks, changes in the regimen of rivers and lakes and in the distribution of plants and animals occurred. Jamieson (1863) recognized more than a century ago that climatic change regulated the expansion and contraction of saline lakes in arid regions. In the Great Basin, the existing lakes are largely confined to the area north of the 39th parallel. Hubbs and Miller (1948) have suggested that this comparatively wet region is representative of the former pluvial conditions that prevailed in those parts of the province that are driest today. Studies of pluvial lakes in New Mexico, such as Lake Estancia (Leopold, 1951), for example, indicate that this is true.

The occurrence of alternating climatic episodes, based upon various kinds of evidence, has been suggested by Bryan (1941), Judson (1953), and Miller and Wendorf (1958) in New Mexico, by Hack (1942) in Arizona, by Richmond (1962) in Utah, by Hunt (1954) in Death Valley, and by many others. Morrison (1961) is one of several workers that has suggested that climatic changes have been virtually synchronous throughout the Great Basin. Most of the reports cited treat climatic fluctuation within Recent time. Although less drastic than the Pleistocene fluctuations, changes during Recent time have been sufficient to produce recognizable imprints upon the sedimentary record, and four or five distinct climatic episodes are thought to have occurred within this time interval.

Data on the age of the alluvial fans in the Basin and Range province are generally unavailable. Although all the fans are not necessarily of equal age, it seems most probable that climatic changes have occurred during their formation. Because climatic change induces changes in the regimen of streams and alters the abundance of vegetation and edaphic characteristics, the climatic hypothesis requires that features thought to attest to such changes be widespread and common to most fans in the region. The features and characteristics of the fan environment that are thought to be indicative of climatic change have been previously discussed in this report. They include the following:

1. The loci of deposition on alluvial fans have shifted from areas well within catchment basins to the middle or lower reaches of the fans, far below the mountain front.
2. Trenches in the apex regions of the fans are misfit relative to present flow conditions by reason of their excessive depth, which is as much as 200 feet.
3. Paired terraces that are continuous with fan surfaces occur within catchment areas and in some places extend to the divide.
4. Abandoned channels occur on fan surfaces at higher elevations than the floors of the active channels.
5. The fan surfaces and the abandoned channels exhibit weathering stain and (or) desert varnish, whereas in the active channels and on modern deposits these are absent.
6. Hanging fans occur in tributary canyons within the catchment areas, and their surfaces are continuous with terraces and fan surfaces below the mountain front.
7. The estimated tractive force is greater within the active channels than on the fan surfaces.
8. The percentages of clay and organic material in both surface sediments and modern mudflows are extremely low.

The formation of the type of alluvial fan discussed in this report according to the climatic hypothesis is shown on figure 137. The observations and data that have been presented suggest the occurrence of the following events during one climatic cycle.

During a period of fan building (fig. 137*A*) aggradation occurred within catchment areas and on fan surfaces at continuous levels. Precipitation was both more abundant and more frequent than it is today. In response to this climatic regime, the abundance of both clay and vegetation was greater, and the fan surfaces were more resistant to erosion. Although discharge was greater, water: sediment ratios were higher, and the medium of transport had a lesser tractive force. Frequent floods spilled over numerous shallow channels onto the fan surfaces, spreading laterally below the mountain front. Sediment was deposited over wide areas of the fans as a consequence of these conditions.

Climatic change, manifested by a change from widespread to local or concentrated precipitation and by a lesser frequency of precipitation, produced a reduction of vegetation and a decrease in the abundance of clay. Catchment slopes and surface deposits became more easily erodible, thus contributing toward a reduction in water: sediment ratios and a greater frequency of mudflows. These flows of greater density, and correspondingly greater tractive force, deepened the trunk channels in catchment areas and trenched the upper reaches of the fans (fig. 137*B*). Terraces and hanging fans in the catchment areas were produced by these processes, and the sediment removed was deposited farther downfan, far below the mountain front.

The trenching of fans does not signal their demise within a geographic cycle, as suggested by Eckis (1928). During an episode of trenching, in response to the climatic conditions outlined above, the shift in the loci of deposition results in the addition of material at the lower limits of the fans. During any part of a climatic cycle, the fans are therefore constantly growing; they grow upward during intervals of general aggradation within the catchment areas and below the mountain front, and are built outward into the basin during periods of trenching in the upper reaches.

The sediment added at the lower margins of the fans may be redistributed and obscured by a subsequent rise in lake level within a given basin. This will produce an alternating alluvial fan-lacustrine facies, such as those commonly reported from core studies. At this time the conditions postulated in figure 137*A* will again occur; the trenches will fill with sediment, and deposition on fan surfaces will again build them upward, provided that the magnitude of the climatic shift is sufficient to accomplish this result before conditions change once again.

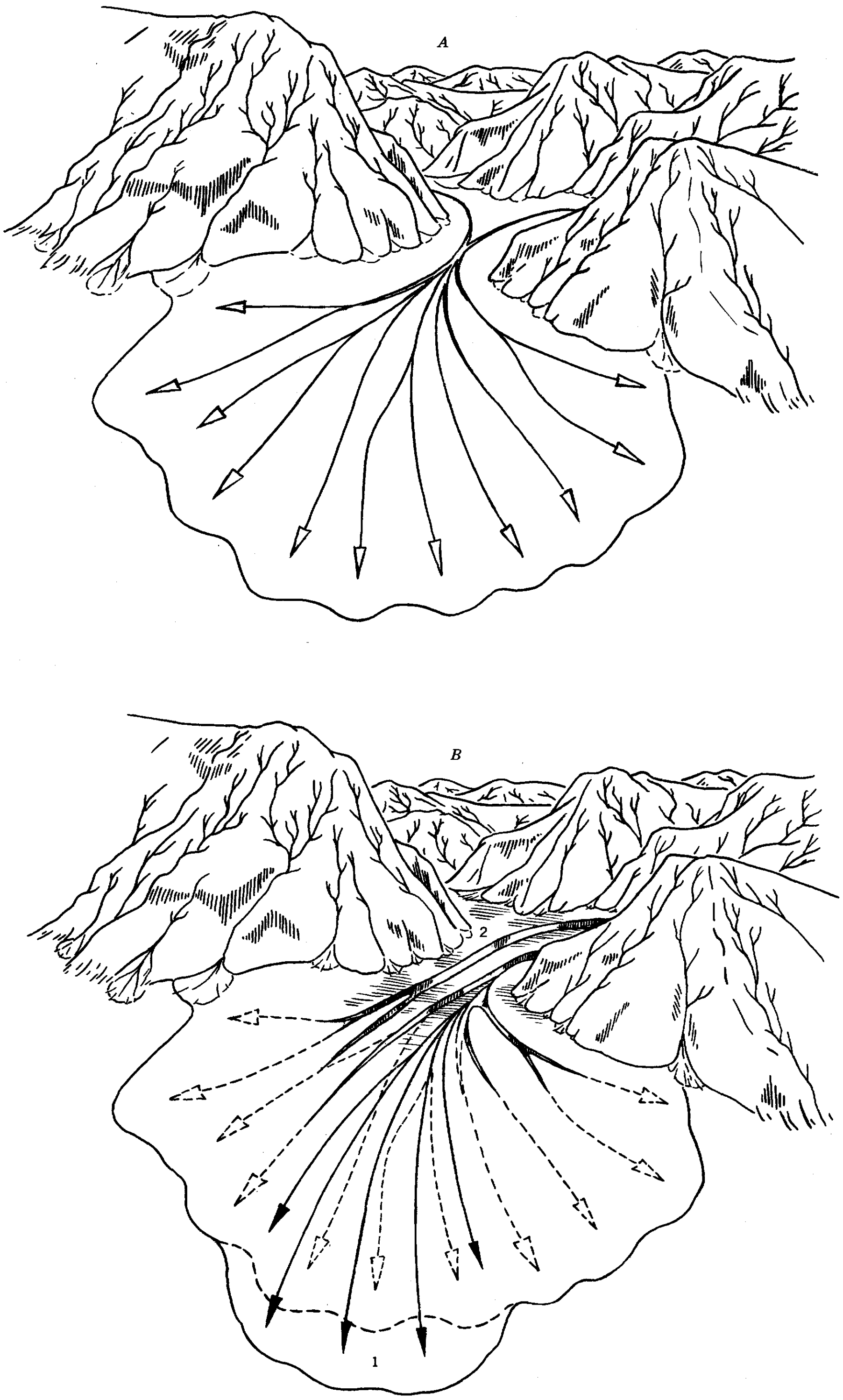

FIGURE 137.—Diagram showing the formation of alluvial fans during one cycle, according to the climatic hypothesis. *A*. During more humid or pluvial periods aggradation occurs within the catchment area and below the mountain front at continuous levels. Frequent floods with high water: sediment ratios spill over the shallow braided channels below the mountain front and deposit sediment over wide areas, building the fans upward. *B*. During a subsequent drier climatic period trenching occurs along main channels within the catchment area and on the upper reaches of the fan. Mudflows and other infrequent density flows transport the sediment they erode to the lower reaches of the fan, building it outward, into the valley (1). The terraces within the catchment area and the abandoned channels on the fan surface are produced as a consequence of this change in process. The small tributary canyon (2) receives little of the infrequent and localized precipitation. Because of trenching in the main channel, the previously deposited sediment is left as a hanging fan at the level of the high terrace as shown.

Both the Davisian and equilibrium theories therefore contain elements of truth. An understanding of the formation of alluvial fans, however, requires recognition of changes of processes through time that are accompanied by consequent changes in fan morphology. Because the most likely cause of such changes is climatic fluctuation, known to have occurred throughout the region, the climatic hypothesis is thought to provide the most reasonable explanation of the formation of fans. It should not be concluded that the equilibrium and climatic hypotheses are mutually exclusive from the foregoing treatment. As summarized so well by Braithwaite (1955, p. 339)—

> The world is not made up of empirical facts with the addition of the laws of nature: what we call the laws of nature are conceptual devices by which we organize our empirical knowledge and predict the future. From this point of view any general hypothesis whose consequences are confirmed by experience is a valuable intellectual device; and the profitable use of such a hypothesis does not presuppose that it will not at some future time be subsumed under some more general hypothesis in a more widely applicable deductive system, nor that the facts that it explains will not some time be explicable by a quite different hypothesis in another deductive system.

TABLE 1.—*Slope and maximum particle-size measurements in Deep Springs Valley, Calif.*

Sample	Slope (feet per foot)	Maximum particle size (feet)	Sample	Slope (feet per foot)	Maximum particle size (feet)
North end of Deep Springs Valley (fig. 90)					
−5, −6	0.160	3.2	−2, 1	0.010	0.3
−4, −6	.075	.2	−1, 1	.020	.1
−3, −6	.035	3.0	−4, 2	.130	.7
−2, −6	.035	2.1	−3, 2	.060	.2
−1, −6	.010	.2	−2, 2	.020	1.6
−6, −5	.160	3.7	−1, 2	.020	.1
−5, −5	.100	.2	−5, 3	.105	1.0
−4, −5	.035	1.6	−4, 3	.145	.6
−3, −5	.040	2.3	−3, 3	.030	.6
−2, −5	.035	1.3	−2, 3	.020	1.9
−1, −5	.030	2.0	−1, 3	.020	3.8
−6, −4	.110	2.7	−4, 4	.050	.2
−5, −4	.065	3.2	−3, 4	.020	5.3
−4, −4	.040	1.8	−2, 4	.025	4.0
−3, −4	.025	3.7	−1, 4	.030	4.7
−2, −4	.030	1.7	−5, 5	.075	.2
−1, −4	.025	.6	−4, 5	.085	2.3
−7, −3	.035	7.0	−3, 5	.055	3.7
−6, −3	.060	1.7	−2, 5	.035	1.7
−5, −3	.040	.1	−1, 5	.035	3.9
−4, −3	.050	2.1	−6, 6	.080	.5
−3, −3	.030	1.7	−5, 6	.040	3.2
−2, −3	.035	1.5	−4, 6	.050	2.3
−1, −3	.015	.1	−3, 6	.055	2.5
−7, −2	.105	1.9	−2, 6	.035	1.9
−6, −2	.070	.4	−1, 6	.025	2.5
−5, −2	.060	.3	−6, 7	.070	7.7
−4, −2	.030	.2	−5, 7	.040	6.4
−3, −2	.035	.2	−4, 7	.065	4.6
−2, −2	.020	.2	−3, 7	.045	5.4
−1, −2	.005	.1	−2, 7	.040	2.8
−6, −1	.120	.5	−1, 7	.030	.3
−5, −1	.060	.1	−6, 8	.060	3.5
−4, −1	.065	.1	−5, 8	.070	5.1
−3, −1	.035	.2	−4, 8	.095	4.8
−2, −1	.030	.6	−3, 8	.035	3.3
−1, −1	.000	.1	−2, 8	.025	2.0
−5, 0	.080	.2	−1, 8	.025	.6
−4, 0	.060	1.7	−7, 9	.110	16.5
−3, 0	.110	1.4	−6, 9	.050	2.2
−2, 0	.005	2.8	−5, 9	.095	2.9
−1, 0	.010	.9	−4, 9	.075	2.8
−6, 1	.100	.5	−3, 9	.040	.6
−5, 1	.090	.4	−2, 9	.030	3.3
−4, 1	.080	.2	−1, 9	.030	.7
−3, 1	.065	.2	−6, 10	.060	2.7

TABLE 1.—*Slope and maximum particle-size measurements in Deep Springs Valley, Calif.*—Continued

Sample	Slope (feet per foot)	Maximum particle size (feet)	Sample	Slope (feet per foot)	Maximum particle size (feet)
North end of Deep Springs Valley (fig. 90)—Continued					
−5, 10	0.085	3.5	4, 3	0.025	0.2
−4, 10	.050	.8	5, 3	.060	.5
−3, 10	.040	.4	6, 3	.085	.6
−2, 10	.040	.1	7, 3	.090	.5
−1, 10	.020	.4	8, 3	.130	1.8
−6, 11	.080	.3	0, 4	.010	.3
−4, 11	.130	1.4	1, 4	.010	.3
−3, 11	.050	1.5	2, 4	.015	.4
−2, 11	.050	1.7	3, 4	.010	1.3
−1, 11	.045	3.5	4, 4	.010	1.0
−2, 12	.055	4.4	5, 4	.065	.3
−1, 12	.050	5.6	6, 4	.065	.2
−2, 13	.045	4.8	7, 4	.080	.3
−1, 13	.055	4.5	8, 4	.140	.9
−5, 14	.160	6.2	0, 5	.015	.4
−4, 14	.055	7.4	1, 5	.020	.8
−3, 14	.050	3.7	2, 5	.015	.2
−2, 14	.100	2.5	3, 5	.020	.2
−1, 14	.080	3.3	4, 5	.010	.7
−3, 15	.070	4.4	5, 5	.030	.7
−2, 15	.065	4.7	6, 5	.050	.4
−3, 16	.190	4.2	7, 5	.065	.3
0, −6	.005	.1	8, 5	.110	1.4
1, −6	.000	.1	0, 6	.000	1.2
2, −6	.010	.0	1, 6	.015	.3
3, −6	.000	.1	2, 6	.015	.1
4, −6	.010	.3	3, 6	.010	.1
5, −6	.030	.2	4, 6	.005	1.2
6, −6	.055	.2	5, 6	.035	.1
0, −5	.005	.2	6, 6	.055	.6
1, −5	.010	.1	7, 6	.070	.7
2, −5	.010	.2	8, 6	.135	1.9
3, −5	.005	.0	0, 7	.020	1.3
4, −5	.005	.0	1, 7	.020	.1
5, −5	.030	.1	2, 7	.010	.1
6, −5	.060	.2	3, 7	.010	.1
0, −4	.000	.1	4, 7	.000	1.6
1, −4	.000	.0	5, 7	.030	.1
2, −4	.005	.1	6, 7	.095	1.0
3, −4	.010	.1	7, 7	.140	5.0
4, −4	.005	.0	0, 8	.010	.1
5, −4	.045	.1	1, 8	.025	.1
6, −4	.050	3.0	2, 8	.025	.1
0, −3	.005	.1	3, 8	.010	1.3
1, −3	.005	.1	4, 8	.010	1.8
2, −3	.000	.1	5, 8	.010	.9
3, −3	.005	.1	0, 9	.020	.2
4, −3	.005	.0	1, 9	.035	.1
5, −3	.065	.3	2, 9	.020	1.0
0, −2	.005	.3	3, 9	.015	.8
1, −2	.005	.1	4, 9	.010	.6
2, −2	.010	.1	5, 9	.055	.7
5, −2	.065	.6	6, 9	.135	5.2
0, −1	.005	.3	0, 10	.030	1.0
1, −1	.000	.1	1, 10	.040	.5
2, −1	.005	.0	2, 10	.025	.4
5, −1	.110	1.1	3, 10	.030	.3
0, 0	.000	1.2	4, 10	.015	.4
1, 0	.005	.1	5, 10	.085	1.3
2, 0	.005	.1	6, 10	.180	7.0
0, 1	.005	1.1	0, 11	.050	2.5
1, 1	.005	.2	1, 11	.035	1.3
2, 1	.000	.4	2, 11	.030	2.0
7, 1	.110	.3	3, 11	.035	1.0
8, 1	.160	.6	4, 11	.065	1.2
0, 2	.010	1.8	5, 11	.180	6.7
1, 2	.005	.2	0, 12	.035	2.6
2, 2	.005	.1	1, 12	.055	2.5
3, 2	.005	.3	2, 12	.050	2.7
4, 2	.050	.3	3, 12	.110	2.8
5, 2	.060	.3	4, 12	.130	5.2
6, 2	.140	.2	0, 13	.040	1.2
7, 2	.100	.3	1, 13	.075	1.8
8, 2	.110	.9	2, 13	.065	2.4
0, 3	.015	.1	3, 13	.205	1.7
1, 3	.010	.1	0, 14	.160	2.3
2, 3	.005	.2	1, 14	.090	3.9
3, 3	.005	.1	2, 14	.105	1.9
Paiute Chute; on fan surface (fig. 114)					
1 [1]	0.080	2.9	7	0.290	6.1
2	.130	3.4	8	.465	4.7
3	.135	5.8	9	.680	3.6
4	.245	3.0	10	.490	8.4
5	.235	6.2	11 [2]	.435	10.4
6	.230	6.5			

See footnotes at end of table.

TABLE 1.—*Slope and maximum particle-size measurements in Deep Springs Valley, Calif.*—Continued

Sample	Slope (feet per foot)	Maximum particle size (feet)	Sample	Slope (feet per foot)	Maximum particle size (feet)
Paiute Chute; in recent channel (fig. 114)					
1[1]	0.035	4.1	8	0.370	6.1
2	.055	3.0	9	.400	7.2
3	.100	5.2	10	.505	6.0
4	.140	7.0	11	.530	5.8
5	.175	5.8	12	.630	8.0
6	.215	7.5	13[2]	.680	7.2
7	.200	9.8			
Westgard Pass (fig. 115)					
1[1]	0.170	0.4	7	0.270	0.6
2	.160	.4	8	.230	.8
3	.230	.3	9	.430	.5
4	.280	.5	10	.380	.9
5	.320	.4	11	.500	.8
6	.250	.6	12[2]	.800	.7
Westgard Pass (fig. 116)					
1[1]	0.220	0.3	6	0.400	0.8
2	.240	.2	7	.440	.7
3	.230	.2	8	.500	.6
4	.400	.4	9	.590	.5
5	.390	.4	10	.660	.7
Antelope Springs fan (fig. 91)					
1	0.075	2.8	36	0.060	2.5
2	.080	2.2	37	.060	1.5
3	.070	1.7	38	.070	1.9
4	.070	2.1	39	.080	2.0
5	.080	2.0	40	.060	1.4
6	.080	1.2	41	.060	1.6
7	.065	2.4	42	.070	2.3
8	.050	3.7	43	.080	2.1
9	.040	2.6	44	.050	1.8
10	.055	1.5	45	.080	2.7
11	.070	1.8	46	.050	2.0
12	.090	2.2	47	.075	1.6
13	.080	2.7	48	.070	1.7
14	.065	1.7	49	.060	1.7
15	.075	1.2	50	.070	2.1
16	.070	2.6	51	.040	1.9
17	.070	1.0	52	.060	2.1
18	.090	1.8	53	.075	2.0
19	.070	1.4	54	.060	2.1
20	.070	1.7	55	.055	1.7
21	.075	3.2	56	.055	2.1
22	.070	1.8	57	.060	2.3
23	.065	1.1	58	.060	1.6
24	.080	2.0	59	.065	1.0
25	.080	2.5	60	.050	1.3
26	.050	2.6	61	.050	1.3
27	.075	2.4	62	.060	2.4
28	.080	3.4	63	.065	2.3
29	.080	2.6	64	.050	2.8
30	.070	1.5	65	.050	2.5
31	.055	1.6	66	.070	1.2
32	.065	2.6	67	.050	3.0
33	.070	2.3	68	.055	2.5
34	.070	1.0	69	.055	1.5
35	.050	2.1	70	.050	1.6

TABLE 1.—*Slope and maximum particle-size measurements in Deep Springs Valley, Calif.*—Continued

Sample	Slope (feet per foot)	Maximum particle size (feet)	Sample	Slope (feet per foot)	Maximum particle size (feet)
Antelope Springs fan (fig. 91)—Continued					
71	0.055	0.8	137	0.035	1.6
72	.045	1.0	138	.030	1.3
73	.055	.9	139	.035	1.3
74	.070	2.1	140	.030	1.5
75	.060	1.2	141	.040	.2
76	.070	1.2	142	.035	1.3
77	.060	2.9	143	.040	1.4
78	.070	2.3	144	.030	1.0
79	.060	1.5	145	.035	1.2
80	.040	1.0	146	.030	2.2
81	.050	1.7	147	.035	1.3
82	.050	1.8	148	.040	1.4
83	.050	1.7	149	.020	1.9
84	.035	1.7	150	.030	1.0
85	.060	1.6	151	.035	1.7
86	.050	1.3	152	.035	1.8
87	.050	1.5	153	.025	.9
88	.055	1.0	154	.025	1.5
89	.040	1.3	155	.020	.9
90	.050	2.0	156	.020	.1
91	.055	1.5	157	.020	1.1
92	.045	1.7	158	.025	.1
93	.070	2.1	159	.030	.1
94	.065	2.0	160	.020	.1
95	.050	1.9	161	.010	.5
96	.050	.9	162	.020	.1
97	.060	2.8	163	.020	.1
98	.050	2.6	164	.020	.1
99	.045	2.3	165	.020	.1
100	.055	1.7	166	.015	.1
101	.050	3.2	167	.030	1.0
102	.040	2.2	168	.020	.1
103	.060	1.2	169	.025	1.6
104	.035	1.4	170	.020	1.5
105	.040	1.1	171	.025	1.3
106	.035	1.0	172	.030	1.1
107	.050	1.7	173	.020	.1
108	.050	.9	174	.025	.1
109	.055	1.6	175	.020	.2
110	.050	1.5	176	.030	.7
111	.050	1.5	177	.015	.3
112	.050	1.4	178	.020	.1
113	.050	1.9	179	.015	.1
114	.045	1.1	180	.010	1.2
115	.050	1.6	181	.015	1.1
116	.050	2.4	182	.010	.2
117	.035	3.0	183	.015	.2
118	.055	1.8	184	.020	.1
119	.030	1.5	185	.020	.1
120	.030	1.6	186	.020	.1
121	.035	1.7	187	.010	.1
122	.040	1.6	188	.015	.1
123	.040	1.2	189	.020	.2
124	.020	1.0	190	.010	.1
125	.035	1.8	191	.010	.5
126	.045	1.4	192	.015	.4
127	.045	1.4	193	.020	.1
128	.040	1.3	194	.010	.7
129	.040	1.6	195	.005	.1
130	.050	1.5	196	.015	.1
131	.050	1.3	197	.010	.1
132	.040	1.4	198	.010	.1
133	.035	1.2	199	.020	.2
134	.045	1.6	200	.020	.1
135	.025	1.0	201	.005	.1
136	.020	1.0	202	.010	.1

[1] At toe of fan. [2] At fan apex.

TABLE 2.—*Mean size, mean roundness, and lithology of pebbles in Deep Springs Valley, Calif.*

Sample	Mean size (mm)	Mean roundness	Lithology (per 50 pebbles)			Sample	Mean size (mm)	Mean roundness	Lithology (per 50 pebbles)		
			Granitic rocks	Sedimentary rocks	Basalt				Granitic rocks	Sedimentary rocks	Basalt
−5, −6	21.9	0.3	50	0	0	−3, −4	36.2	1.4	50	0	0
−4, −6	17.4	.2	50	0	0	−2, −4	28.3	1.3	49	1	0
−3, −6	15.1	.4	50	0	0	−1, −4	12.0	.7	50	0	0
−2, −6	13.6	.4	50	0	0	−7, −3	36.0	1.6	50	0	0
−1, −6	16.9	.9	50	0	0	−6, −3	31.0	1.5	49	1	0
−6, −5	31.5	.0	50	0	0	−5, −3	18.6	.5	50	0	0
−5, −5	19.3	.4	49	1	0	−4, −3	15.7	.5	50	0	0
−4, −5	17.9	.1	50	0	0	−3, −3	15.9	.1	50	0	0
−3, −5	14.2	.2	50	0	0	−2, −3	17.3	.2	50	0	0
−2, −5	13.1	.2	50	0	0	−1, −3	15.8	.5	49	1	0
−1, −5	15.9	.4	50	0	0	−7, −2	31.6	1.0	50	0	0
−6, −4	18.4	.1	50	0	0	−6, −2	16.1	.1	50	0	0
−5, −4	17.3	.4	50	0	0	−5, −2	18.7	.1	50	0	0
−4, −4	17.3	.2	50	0	0	−4, −2	14.2	.1	50	0	0

TABLE 2.—*Mean size, mean roundness, and lithology of pebbles in Deep Springs Valley, Calif.*—Continued

Sample	Mean size (mm)	Mean roundness	Lithology (per 50 pebbles)			Sample	Mean size (mm)	Mean roundness	Lithology (per 50 pebbles)		
			Granitic rocks	Sedimentary rocks	Basalt				Granitic rocks	Sedimentary rocks	Basalt
−3, −2	17.4	0.4	50	0	0	5, −5	12.6	0.3	50	0	0
−2, −2	15.2	.3	50	0	0	6, −5	17.6	.0	50	0	0
−1, −2	10.9	1.9	33	16	1	0, −4	8.9	2.2	41	6	3
−6, −1	17.6	.0	50	0	0	1, −4	10.9	2.2	38	12	0
−5, −1	18.5	.1	50	0	0	2, −4	9.1	1.8	47	1	2
−4, −1	15.9	.1	50	0	0	3, −4	11.1	2.4	38	7	5
−3, −1	15.6	.1	50	0	0	4, −4	8.3	2.6	32	9	9
−2, −1	19.3	.3	50	0	0	5, −4	12.9	.1	50	0	0
−1, −1	11.8	1.8	34	13	3	6, −4	21.2	.0	50	0	0
−5, 0	16.8	.0	50	0	0	0, −3	10.5	2.4	36	12	2
−4, 0	17.6	.1	50	0	0	1, −3	11.7	2.0	41	9	0
−3, 0	22.4	.1	50	0	0	2, −3	7.7	2.2	40	6	4
−2, 0	12.4	1.5	32	16	2	3, −3	7.4	2.1	39	8	3
−1, 0	9.5	2.0	38	7	5	4, −3	8.3	2.4	29	10	11
−6, 1	16.2	.1	50	0	0	5, −3	13.6	.0	50	0	0
−5, 1	16.6	.1	50	0	0	0, −2	10.2	2.3	37	12	1
−4, 1	30.4	.2	50	0	0	1, −2	10.9	2.5	33	10	7
−3, 1	20.5	.2	50	0	0	2, −2	7.7	2.3	42	5	3
−2, 1	19.6	2.1	28	15	7	5, −2	22.8	.0	50	0	0
−1, 1	13.3	2.2	29	17	4	0, −1	11.3	2.0	35	12	3
−4, 2	16.2	.0	50	0	0	1, −1	9.0	2.3	38	11	1
−3, 2	15.5	.1	50	0	0	2, −1	8.0	2.1	39	11	0
−2, 2	12.4	1.8	34	12	4	5, −1	21.3	.0	50	0	0
−1, 2	10.3	2.3	35	12	3	0, 0	15.3	1.8	27	21	2
−5, 3	23.9	.0	50	0	0	1, 0	11.8	2.2	25	23	2
−4, 3	17.3	.0	50	0	0	2, 0	11.9	2.2	29	17	4
−3, 3	20.6	1.4	39	11	0	0, 1	11.9	1.5	34	12	4
−2, 3	13.0	1.6	32	15	3	1. 1	10.0	2.1	34	12	4
−1, 3	11.6	2.0	24	19	7	2 1	16.8	2.2	29	16	5
−4, 4	15.1	.0	50	0	0	7, 1	18.9	.0	50	0	0
−3, 4	19.9	1.9	33	14	3	8, 1	22.4	.0	50	0	0
−2, 4	14.5	1.8	34	9	7	0, 2	14.1	1.7	35	14	1
−1, 4	14.6	1.6	31	18	1	1, 2	11.9	1.9	30	13	7
−5, 5	24.1	.1	50	0	0	2, 2	18.2	3.3	9	40	1
−4, 5	18.7	.5	50	0	0	3, 2	9.2	2.4	37	10	3
−3, 5	15.9	1.5	34	13	3	4, 2	14.6	.1	50	0	0
−2, 5	15.4	2.0	31	16	3	5, 2	14.3	.2	50	0	0
−1, 5	17.3	.9	38	10	2	6, 2	20.2	.0	50	0	0
−6, 6	19.2	.0	50	0	0	7, 2	17.8	.0	50	0	0
−5, 6	19.2	.2	47	2	1	8, 2	18.2	.0	50	0	0
−4, 6	21.8	2.0	29	19	2	0, 3	21.1	1.8	25	21	4
−3, 6	19.1	1.7	32	14	4	1, 3	17.3	2.7	8	42	0
−2, 6	17.5	2.4	9	40	1	2, 3	20.5	3.1	7	42	1
−1, 6	17.3	1.9	19	27	4	3, 3	11.9	2.1	27	18	5
−6, 7	16.1	.3	47	3	0	4, 3	12.1	.2	46	4	0
−5, 7	20.7	2.3	27	17	6	5, 3	12.4	.2	50	0	0
−4, 7	23.7	2.3	10	37	3	6, 3	20.1	.1	50	0	0
−3, 7	23.9	2.1	11	34	5	7, 3	16.4	.0	50	0	0
−2, 7	18.5	1.8	9	40	1	8, 3	26.6	.0	50	0	0
−1, 7	13.9	1.6	21	26	3	0, 4	14.8	1.5	21	25	4
−6, 8	26.9	2.1	28	17	5	1, 4	13.1	2.1	32	18	0
−5, 8	29.6	1.8	20	25	5	2, 4	10.6	2.0	29	16	5
−4, 8	28.0	1.9	11	29	10	3, 4	12.2	2.2	29	19	2
−3, 8	35.5	1.2	24	17	9	4, 4	17.3	2.1	13	2	35
−2, 8	18.2	2.1	15	35	0	5, 4	11.6	.6	43	6	1
−1, 8	13.7	.7	45	1	4	6, 4	14.4	.9	41	9	0
−7, 9	43.3	3.0	31	15	4	7, 4	15.5	.0	50	0	0
−6, 9	32.1	1.1	19	27	4	8, 4	18.6	.0	50	0	0
−5, 9	24.3	2.0	15	32	3	0, 5	15.6	1.8	22	26	2
−4, 9	25.8	1.8	11	34	5	1, 5	13.4	2.4	14	32	4
−3, 9	20.2	2.3	14	34	2	2, 5	17.3	2.4	9	39	2
−2, 9	16.6	1.4	26	24	0	3, 5	11.1	.7	46	3	1
−1, 9	15.7	.7	48	1	1	4, 5	14.4	1.8	20	1	29
−6, 10	16.6	1.3	18	24	8	5, 5	17.9	.2	50	0	0
−5, 10	28.0	2.1	14	30	6	6, 5	16.2	.1	49	1	0
−4, 10	19.2	1.3	27	20	3	7, 5	17.6	.1	49	1	0
−3, 10	14.8	1.3	23	25	2	8, 5	17.9	.2	48	2	0
−2, 10	19.4	.9	43	4	3	0, 6	14.3	1.7	34	14	2
−1, 10	15.2	.7	47	1	2	1, 6	16.2	1.7	28	21	1
−6, 11	17.0	.6	39	7	4	2, 6	13.8	1.0	42	7	1
−4, 11	23.1	1.8	8	16	26	3, 6	13.1	.9	45	0	5
−3, 11	16.4	.9	31	13	6	4, 6	8.2	.9	46	1	3
−2, 11	18.2	.5	49	1	0	5, 6	14.6	.4	49	0	1
−1, 11	14.9	.3	43	0	7	6, 6	15.7	.1	50	0	0
−2, 12	18.6	.7	46	2	2	7, 6	21.6	.0	50	0	0
−1, 12	18.1	.2	48	1	1	8, 6	23.0	.0	50	0	0
−2, 13	17.2	.2	45	3	2	0, 7	16.4	1.7	30	19	1
−1, 13	18.6	.0	50	0	0	1, 7	13.6	.7	50	0	0
−5, 14	20.2	.1	44	0	6	2, 7	12.6	.7	50	0	0
−4, 14	17.9	.6	42	3	5	3, 7	11.0	.4	50	0	0
−3, 14	17.3	.5	43	2	5	4, 7	11.5	1.7	19	0	31
−2, 14	20.5	.2	50	0	0	5, 7	14.7	2.1	8	0	42
−1, 14	18.6	.2	49	0	1	6, 7	14.0	.0	50	0	0
−3, 15	18.9	.1	50	0	0	7, 7	17.7	.0	50	0	0
−2, 15	21.1	.0	49	0	1	0, 8	13.2	.8	47	2	1
−3, 16	24.5	.0	50	0	0	1, 8	18.1	1.0	48	1	1
0, −6	6.9	1.5	45	4	1	2, 8	13.5	.7	50	0	0
1, −6	8.8	2.0	40	8	2	3, 8	13.2	.9	50	0	0
2, −6	7.5	1.7	39	10	1	4, 8	13.5	.7	50	0	0
3, −6	7.1	1.7	43	6	1	5, 8	15.8	2.3	8	0	42
4, −6	9.0	2.0	27	16	7	0, 9	15.8	.6	49	0	1
5, −6	15.9	.1	50	0	0	1, 9	13.6	.8	50	0	0
6, −6	16.9	.0	50	0	0	2, 9	12.8	.8	50	0	0
0, −5	7.8	1.6	47	2	1	3, 9	13.1	.6	50	0	0
1, −5	8.4	1.8	40	6	4	4, 9	13.1	2.2	15	0	35
2, −5	8.5	2.0	39	9	2	5, 9	13.7	.1	50	0	0
3, −5	9.0	2.2	37	9	4	6, 9	17.3	.0	50	0	0
4, −5	7.7	2.0	39	6	5	0, 10	16.9	.4	47	2	1

TABLE 2.—*Mean size, mean roundness, and lithology of pebbles in Deep Springs Valley, Calif.*—Continued

Sample	Mean size (mm)	Mean roundness	Lithology (per 50 pebbles)			Sample	Mean size (mm)	Mean roundness	Lithology (per 50 pebbles)		
			Granitic rocks	Sedimentary rocks	Basalt				Granitic rocks	Sedimentary rocks	Basalt
1, 10	12.7	0.6	50	0	0	0, 12	16.6	0.4	50	0	0
2, 10	12.7	.7	49	0	1	1, 12	16.9	1.6	36	1	13
3, 10	16.8	1.5	33	0	17	2, 12	16.6	1.8	23	0	27
4, 10	18.5	2.2	16	1	33	3, 12	18.6	1.2	50	0	0
5, 10	15.3	.2	50	0	0	4, 12	27.1	1.7	33	4	13
6, 10	15.9	.0	50	0	0	0, 13	19.0	.7	50	0	0
0, 11	14.4	.4	46	4	0	1, 13	21.6	2.3	28	0	32
1, 11	13.3	.9	46	1	3	2, 13	19.6	1.2	24	0	26
2, 11	13.3	1.3	39	0	11	3, 13	18.2	.0	50	0	0
3, 11	15.0	1.9	25	0	25	0, 14	20.4	.8	40	0	10
4, 11	16.1	1.5	32	9	9	1, 14	25.3	1.0	13	0	37
5, 11	17.0	.1	50	0	0	2, 14	18.5	.2	50	0	0

TABLE 3. *Basic data on the tails of the granule-to-clay size distribution*

Sample	Granules (weight percentage)	Clay (weight percentage)	Silt/Clay	Sample	Granules (weight percentage)	Clay (weight percentage)	Silt/Clay
Antelope Springs active channel (pl. 9)							
9	4.21	1.55	6.44:1	291	9.44	2.76	2.26:1
62	9.39	.57	5.61:1	301	2.88	.00	∞
85	15.69	.83	6.44:1	318	2.92	1.24	3.67:1
126	.25	1.52	5.21:1	353	3.92	2.12	3.11:1
149	.14	1.17	21.59:1	400	2.24	1.41	4.94:1
199	14.21	1.42	4.97:1	437	1.28	.91	10.84:1
214	8.28	1.72	5.54:1	463	7.29	1.51	3.62:1
241	15.35	.54	9.75:1	480	.80	1.50	2.94:1
279	17.71	.20	14.83:1	512	.13	.77	4.80:1
				553	.99	1.74	1.65:1
Paiute Chute							
P-1 [1]	10.60	2.10	4.32:1	P-2 [1]	12.90	3.00	4.92:1
Mudflows							
BP-1 [3]	17.12	2.41	4.39:1	Wy-1 [5]	0.00	17.38	4.28:1
BP-2 [4]	8.42	4.22	3.57:1				
North end of Deep Springs Valley (fig. 90)							
−5, −4	6.06	1.43	8.36:1	0, 1	2.81	0.99	11.82:1
−4, −4	10.29	.49	26.58:1	7, 1	4.94	2.17	1.68:1
−7, −3	9.04	.98	7.23:1	2, 2	2.60	.75	11.47:1
−2, −3	16.76	1.71	6.30:1	4, 2	6.01	3.62	6.11:1
−5, 0	7.51	1.10	7.48:1	0, 3	5.27	3.79	3.05:1
−3, 0	8.49	4.12	2.96:1	6, 3	28.50	.00	∞
−1, 0	13.31	.34	32.95:1	8, 3	20.00	2.01	6.80:1
−1, 3	2.16	1.24	11.52:1	3, 4	3.14	2.46	4.92:1
−3, 5	6.85	.98	17.60:1	0, 5	8.30	.97	13.30:1
−5, 6	1.85	1.18	8.46:1	4, 5	3.27	3.11	5.05:1
−5, 7	4.76	2.15	4.76:1	6, 5	17.82	1.61	16.64:1
−4, 7	4.02	2.70	5.76:1	8, 5	21.19	1.72	8.00:1
−5, 8	5.69	.92	5.63:1	2, 6	7.95	.00	∞
−7, 9	13.32	1.67	8.90:1	0, 7	5.34	1.11	11.50:1
−5, 9	6.08	2.14	7.10:1	4, 7	.34	1.48	10.15:1
−5, 10	1.79	6.37	3.91:1	6, 7	23.60	.64	27.28:1
−3, 10	3.04	1.83	5.76:1	2, 8	5.75	1.45	7.67:1
−1, 12	8.76	1.31	8.46:1	3, 8	2.84	2.34	6.48:1
−2, 14	14.99	.00	∞	0, 9	10.91	.78	15.30:1
0, −6	5.02	2.35	7.04:1	3, 9	8.60	1.36	11.19:1
3, −6	1.90	2.89	11.05:1	5, 9	8.02	1.37	10.78:1
4, −5	2.62	3.93	5.81:1	1, 10	3.52	1.07	5.95:1
6, −5	20.40	.70	12.00:1	4, 10	3.94	2.18	7.91:1
0, −4	5.70	1.65	8.00:1	6, 10	15.19	1.74	10.06:1
2, −4	1.74	3.91	11.12:1	0, 11	11.39	.84	11.23:1
3, −4	5.61	.00	∞	3, 11	4.16	1.68	7.16:1
5, −4	11.88	1.04	19.18:1	5, 11	22.65	1.68	6.28:1
4, −3	12.58	3.53	8.16:1	2, 12	9.99	.67	7.30:1
1, −2	11.19	2.48	6.78:1	4, 12	9.99	1.28	4.16:1
2, −2	3.68	.00	∞	1, 13	16.58	.88	10.44:1
1, −1	5.59	3.22	10.87:1	3, 13	15.04	2.50	6.70:1
5, −1	3.79	26.58	1.06:1	0, 14	10.21	.89	9.96:1
2, 0	2.28	1.29	7.81:1	2, 14	16.05	1.26	9.44:1

[1] At fan apex.
[2] At midfan.
[3] 2 miles below mountain front.
[4] 8 miles below mountain front.
[5] At BM 5220.

TABLE 4.—*Sedimentary parameters of the granule-to-clay size fraction, in phi units*

Sample	Median	M_z	σ_I	Sk_I	K_G	Sample	Median	M_z	σ_I	Sk_I	K_G
Antelope Springs active channel (pl. 9)											
9	1.95	1.96	2.12	0.12	1.62	301	2.40	2.43	1.55	0.21	1.49
62	1.28	1.22	1.53	−.30	1.16	318	1.67	1.71	1.35	.11	1.34
85	1.65	1.27	2.01	−.15	1.10	353	1.60	1.66	1.86	.62	1.16
126	2.14	2.24	1.46	.27	1.60	400	1.73	1.56	1.65	.23	1.38
149	3.32	3.78	1.59	.40	1.34	437	1.99	2.14	1.63	.27	1.41
199	1.02	1.02	2.10	.15	1.28	463	1.71	1.59	1.92	.03	1.34
214	2.26	2.13	2.23	−.02	1.73	480	2.07	2.11	1.19	.12	1.42
241	1.30	1.18	1.97	.02	1.06	512	1.65	1.72	1.03	.19	1.22
279	.80	.70	1.69	−.01	.90	553	1.73	1.77	1.02	.09	1.12
291	1.40	1.43	2.01	.12	1.33						
Paiute Chute											
P-1 [1]	1.64	1.53	2.17	0.07	1.59	P-2 [2]	1.30	1.67	2.67	0.27	1.04
Mudflows											
BP-1 [3]	0.95	1.13	2.47	0.25	1.15	Wy-1 [5]	6.30	6.29	1.67	0.02	0.83
BP-2 [4]	1.53	2.01	2.73	.31	1.19						
North end of Deep Springs Valley (fig. 90)											
−5, −4	1.70	1.80	2.20	0.18	1.24	7, 1	1.12	1.27	1.56	.19	1.28
−4, −4	.74	1.25	2.34	.39	1.13	2, 2	1.51	1.71	1.83	.27	1.17
−7, −3	1.01	1.20	2.03	.27	1.22	4, 2	2.90	2.89	2.77	.04	1.14
−2, −3	.49	1.03	2.45	.41	1.11	0, 3	1.93	2.03	2.24	.20	1.31
−5, 0	1.00	1.27	2.03	.29	1.24	6, 3	−.18	.09	1.71	.33	1.01
−3, 0	1.49	1.72	2.50	.37	1.20	8, 3	.55	1.11	2.65	.40	.96
−1, 0	.70	1.07	2.29	.34	1.19	3, 4	2.51	2.38	1.96	.04	1.49
−1, 3	1.78	2.02	2.04	.28	1.17	0, 5	1.61	1.70	2.30	.17	1.08
−3, 5	1.70	2.35	2.44	.25	1.05	4, 5	2.00	2.28	2.46	.25	1.15
−5, 6	1.56	1.73	1.93	.28	1.17	6, 5	2.73	2.49	3.13	−.03	.77
−5, 7	1.50	1.61	2.12	.24	1.20	8, 5	.48	1.06	2.62	.41	.97
−4, 7	1.98	2.28	2.36	.24	1.19	2, 6	1.22	1.44	2.09	.22	1.44
−5, 8	.81	1.34	1.64	.29	1.22	0. 7	2.31	2.14	2.09	.00	1.22
−7, 9	1.35	1.54	2.55	.22	1.02	4, 7	2.37	2.44	1.90	.22	1.22
−5, 9	1.68	1.97	2.42	.25	1.08	6, 7	.60	1.25	3.77	.40	.83
−5, 10	2.89	3.39	2.60	.28	.97	2, 8	1.72	1.78	2.15	.16	1.14
−3, 10	1.65	1.83	2.06	.26	1.12	3, 8	2.21	2.34	2.18	.20	1.25
−1, 12	1.51	1.61	2.24	.18	1.18	0, 9	1.30	1.49	2.27	.22	1.11
−2, 14	.71	1.02	2.22	.30	1.09	3, 9	1.40	1.73	2.67	.28	1.08
0, −6	2.10	2.30	2.38	.24	1.18	5, 9	1.92	1.88	2.35	.15	1.09
3, −6	3.13	3.35	2.65	.12	.87	1, 10	1.80	1.82	1.71	.10	1.28
4, −5	3.10	3.23	2.91	.10	1.20	4, 10	2.46	2.51	2.39	.11	1.17
6, −5	.57	.79	2.23	.27	1.24	6, 10	1.20	1.67	2.78	.30	.95
0, −4	2.11	2.01	2.21	.07	1.14	0, 11	.71	1.12	2.23	.36	1.18
2, −4	3.80	3.83	2.74	.00	1.01	3, 11	1.65	1.84	2.14	.27	1.16
3, −4	1.31	1.27	1.28	−.06	1.09	5, 11	.42	.86	2.48	.38	1.04
5, −4	1.95	2.11	2.72	.15	.93	2, 12	.51	.67	1.61	.24	1.26
4, −3	3.20	2.98	3.13	−.06	.84	4, 12	.70	.87	1.85	.25	1.25
1, −2	1.89	2.04	2.67	.16	1.04	1, 13	.35	.81	2.22	.41	1.33
2, −2	1.52	1.61	1.59	.15	1.28	3, 13	1.52	1.81	2.80	.23	1.14
1, −1	3.30	3.38	2.93	.03	.82	0, 14	1.20	1.35	2.13	.20	1.19
2, 0	1.99	2.10	1.83	.21	1.33	2, 14	.94	1.33	2.46	.28	1.02
0, 1	1.62	1.84	1.97	.28	1.85						

[1] At fan apex.
[2] At midfan.
[3] 2 miles below mountain front.
[4] 8 miles below mountain front.
[5] BM 5220.

TABLE 5.—*Sample modes and their weight percentage of the granule-to-clay size fraction*

Sample	Primary mode	Weight percent	Secondary mode	Weight percent
Antelope Springs active channel (pl. 9)				
9	Medium sand	26.31	None	
62	do	29.38	do	
85	do	25.55	Granules	15.69
126	do	31.84	None	
149	Very fine sand	33.79	do	
199	Medium sand	22.75	Silt	7.03
214	Fine sand	26.36	Granules	8.28
241	Medium sand	22.49	do	15.35
279	do	23.88	do	17.71
291	do	24.92	None	
301	Fine sand	30.21	Granules	2.88
318	Medium sand	38.26	None	
353	do	21.95	do	
400	do	31.49	do	
437	do	33.05	do	
463	do	26.50	do	
480	do	33.52	do	
512	do	45.93	do	
553	do	43.43	do	
Paiute Chute				
P-1 [1]	Medium sand	28.41	Granules	10.60
P-2 [2]	Coarse sand	16.91	Silt	14.70
Mudflows				
BP-1 [3]	Medium sand	17.46	Granules	17.12
BP-2 [4]	do	18.80	Silt	15.07
Wy-1 [5]	Silt	77.25	None	
North end of Deep Springs Valley (fig. 90)				
−5, −4	Medium sand	21.88	None	
−4, −4	Coarse sand	22.90	Silt	13.00
−7, −3	do	22.24	None	
−2, −3	Very coarse sand	23.07	Silt	10.78
−5, 0	Coarse sand	25.15	None	
−3, 0	Very coarse sand	19.81	Silt	12.20
−1, 0	do	22.71	do	11.20
−1, 3	Medium sand	24.15	do	14.31
−3, 5	Coarse sand	19.12	do	17.29
−5, 5	do	25.40	None	
−5, 7	do	22.09	do	
−4, 7	Medium sand	21.28	Very fine sand	16.17
−5, 8	Coarse sand	29.63	None	
−7, 9	Very coarse sand	17.34	Silt	14.87
−5, 9	Coarse sand	19.86	do	15.21
−5, 10	Silt	24.81	Medium sand	19.43
−3, 10	Coarse sand	23.27	Very fine sand	15.44
−1, 12	Medium sand	20.04	None	
−2, 14	Coarse sand	20.64	Silt	9.65
0, −6	Medium sand	18.67	Very fine sand	17.39
3, −6	Silt	31.95	Medium sand	14.25
4, −5	Very fine sand	24.87	None	
6, −5	Coarse sand	22.50	Granules	20.40
0, −4	Very fine sand	18.95	None	
2, −4	Silt	40.35	Very coarse sand	7.45
3, −4	Medium sand	32.30	None	
5, −4	Silt	19.95	Medium sand	13.47
4, −3	do	28.80	Granules	12.58
1, −2	Very fine sand	17.17	Coarse sand	15.68
2, −2	Medium sand	31.97	None	
1, −1	Silt	35.10	Medium sand	12.37
5, −1	do	28.05	Coarse sand	15.39
2, 0	Medium sand	26.60	None	
0, 1	do	25.04	do	
7, 1	Coarse sand	28.51	do	
2, 2	do	25.16	do	
4, 2	Very fine sand	22.33	Medium sand	12.94
0, 3	Medium sand	20.71	Very fine sand	16.99
6, 3	Granules	28.50	Silt	4.62
8, 3	Very coarse sand	20.49	do	13.72
3, 4	Very fine sand	22.90	None	
0, 5	Coarse sand	18.24	Very fine sand	15.84
4, 5	Medium sand	18.76	do	17.09
6, 5	Silt	26.80	Granules	17.82
8, 5	Granules	21.19	Silt	13.75
2, 6	Coarse sand	21.06	Very fine sand	13.79
0, 7	Very fine sand	21.69	None	
4, 7	Medium sand	24.30	Very fine sand	21.00
6, 7	Granules	23.60	Silt	13.45
2, 8	Medium sand	19.71	None	
3, 8	do	21.15	do	
0, 9	Coarse sand	20.15	Silt	11.91
3, 9	do	20.60	do	15.29
5, 9	Very fine sand	17.37	Medium sand	16.68
1, 10	Medium sand	28.28	None	
4, 10	Very fine sand	21.12	Medium sand	17.14
6, 10	Silt	17.48	Very coarse sand	17.26
0, 11	Coarse sand	22.99	Silt	9.45
3, 11	do	21.76	Very fine sand	15.21
5, 11	Granules	22.65	Silt	10.46
2, 12	Coarse sand	30.39	None	
4, 12	do	24.99	do	
1, 13	Very coarse sand	24.92	Silt	9.20
3, 13	Silt	16.80	Very coarse sand	15.75
0, 14	Medium sand	21.32	None	
2, 14	Coarse sand	18.02	Silt	11.89

[1] At fan apex.
[2] At midfan.
[3] 2 miles below mountain front.
[4] 8 miles below mountain front.
[5] BM 5220

TABLE 6.—*Clay-mineral ratios of related samples, in parts per ten*

Sample	Montmorillonite group	Illite	Kaolinite and (or) chlorite	Sample	Montmorillonite group	Illite	Kaolinite and (or) chlorite
Antelope Springs active channel (pl. 9)							
9	5.2	3.8	1.0	301	2.1	5.8	2.1
85	1.3	6.5	2.2	318	2.2	5.7	2.1
126	2.6	5.6	1.8	353	3.3	4.9	1.8
149	3.1	4.8	2.1	400	2.1	5.5	2.4
199	.7	7.0	2.3	437	3.1	4.2	2.7
214	2.4	5.3	2.3	463	3.2	4.2	2.6
241	4.0	4.3	1.7	480	3.4	4.7	1.9
279	2.5	6.0	1.5	512	1.3	7.1	1.6
291	3.7	3.9	2.4	553	2.2	5.6	2.2
Paiute Chute							
P-1	7.1	1.9	1.0	P-2	5.5	3.4	1.1
Mudflows							
BP-1 [1]	2.4	5.2	2.4	Wy-1 [2]	5.7	3.5	0.8
BP-2 [1]	4.8	2.9	2.3				
North end of Deep Springs Valley (fig. 90)							
2, 14	3.3	4.1	2.6	−5. 8	1.1	6.8	2.1
3, 13	5.8	3.4	.8	3, 8	4.2	4.5	1.3
4, 12	6.0	2.5	1.5	−5, 7	4.3	3.9	1.8
5, 11	5.1	3.0	1.9	−4, 7	3.4	5.4	1.2
6, 10	6.0	2.6	1.4	−5, 6	3.9	4.5	1.6
−5, 10	5.6	4.2	.2	−3, 5	4.3	4.3	1.4
4, 10	4.8	4.0	1.2	−1, 3	2.6	5.7	1.7
−5, 9	3.1	5.1	1.8	1, −1	4.0	4.3	1.7
5, 9	4.7	3.6	1.7	0, 1	4.5	4.5	1.0
3, 9	3.0	5.1	1.9	4, 7	3.4	5.2	1.4

[1] At Big Pine.
[2] BM 5220

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