Downstream Changes in Stream Power in the Henry Mountains, Utah

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Abstract. Total stream power does not necessarily increase systematically in the downstream direction because of the conflicting influences of channel slope, width, and depth. Historical records and field data for arroyo systems of the northern Henry Mountains, south-central Utah, show that total stream power decreased in the downstream direction during a deposition period before 1896 and increased downstream during an erosion period thereafter. When total stream power declined in the downstream direction, channels were small and meandering, and the ten-year flood exceeded channel capacity, resulting in overbank deposition of sediment. After an especially erosive flood in 1896, total stream power increased in the downstream direction because channels were in the bottoms of arroyos that confined discharges, resulting in channel erosion and through-put of sediment. In 1980 deposition was occurring in the headward portions of the continuous arroyo system and in the large master stream, but not in mid-basin areas. Channels in the arid and semi-arid Henry Mountains do not exhibit mutual adjustment between form and process. Rather, at times of catastrophic system-wide events, fluvial processes control channel forms, whereas at other times channel forms control fluvial processes. Comparison of historical and modern conditions shows that mutual adjustment is unlikely to occur in the discontinuous operation of semi-arid fluvial systems.

Key Words: fluvial geomorphology, stream power, arroyos.

EXTENSIVE channel erosion in many semi-arid parts of the world has caused elimination of agriculturally productive floodplains, sediment pollution of downstream areas and reservoirs, and destruction of bridges, crossings, and irrigation works. The remaining unstable channel configurations pose continuing natural hazards for users of the channel and near-channel environment. Planners, engineers, and decision makers must deal with these dynamic systems with the aid of inadequate generalizations about the spatial and temporal changes in the channels and the arroyos they have created. The purpose of this paper is to develop generalizations about the geographic properties of unstable channel systems in a representative semi-arid environment.

The development of generalizations depends on construction of fluvial theory, which in many aspects is more advanced than other realms of geomorphology. An important gap in fluvial theory, however, is a lack of understanding of the spatial characteristics of stream power, especially on the regional or network scale. Established theory for spatial characteristics of sediment sizes and of channel morphology (Simons and Senturk 1978; Leopold, Wolman, and Miller 1964) would be more useful if it were complemented by a similar theory for power. Knowledge about temporal changes in stream power has also been lacking, despite radical adjustments in semi-arid streams during historical periods (for example, see Cooke and Reeves 1976). Using data from the Henry Mountains of south-central Utah, the following report attempts to provide some general outlines for theory development pertaining to spatial and temporal characteristics of stream power.

The Henry Mountains region is one of the classic localities in the development of modern fluvial geomorphology because of the research conducted there by G. K. Gilbert in 1875–1876 (Gilbert 1877). Like most of the American Southwest, the Henry Mountains

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373
area experienced widespread channel erosion in the late 1800s and early 1900s, but because of Gilbert’s early work the conditions in the Henry Mountains prior to this erosional phase are better known than those in other areas. Stream systems throughout the Henry Mountains have developed into arroyos that are deepest in mid-basin areas and become progressively more shallow upstream (Figure 1). On the local scale of a few kilometers, this upstream decline in depth is systematic (Graf 1982b). On a regional scale of several tens of kilometers, the arroyos also decline in depth in the downstream direction. This regional arrangement does not appear to be related to geologic materials, though there are some minor exceptions. The arroyo systems are continuous on a basin-wide scale, and their dimensions are not related to valley floor slopes, so their present configuration cannot be explained in terms of the formation of discontinuous gullies (Schumm and Hadley 1957; Patton and Schumm 1975). If stream power is computed for flows that are contained within the arroyo, it appears that power is likely to be spatially inconsistent in the downstream direction, initially increasing from the headwaters to mid-basin locations and then decreasing.

The initial field observations of channel changes in the Henry Mountains give rise to three basic research questions. First, what is the present spatial distribution of total stream power in the arroyo network? Second, how has that distribution changed over time? Third, what is the nature of the changing relationship between channel form and process that explains the spatial and temporal variation in stream power?

**Context of The Research**

Research into the behavior of stream systems in the arid and semi-arid western United States has focused in part on the processes of channel entrenchment that have occurred in many areas during the past century. Much scientific capital has been invested in efforts to explain the initiation of arroyo cutting or gully development (Cooke and Reeves 1976). The dynamic processes of discontinuous gullies have been investigated beginning with the work of Schumm and Hadley (1957), but the operation of fluvial processes in continuous arroyo networks remains little studied. This is the subject of the work reported here. An overview of general fluvial processes in the Henry Mountains appears elsewhere (Graf 1980). Graf (1982a) proposed in abbreviated form three avenues for exploration of spatial and temporal variation of arid/semi-arid fluvial processes. First, distance-decay functions were proposed as a tool for investigating the interaction among stream processes represented by discharge, distance to a site of disruption in the system, and the resulting channel erosion at various network locations. A further, more detailed analysis of this proposal showed the significant influence of geologic materials (Graf 1982b). Second, catastrophe theory was proposed as a method of describing temporal adjustments. The detailed analysis of this proposal is the subject of further continuing work and forthcoming reports. Third, downstream changes in channel shear stress were proposed as an important spatial characteristic.
of arid/semi-arid streams, a concept supported by limited, preliminary data. The present report uses an expanded set of data in a complete exploration of the third proposed approach, and extends the analysis of shear stress to the more useful measure of total stream power.

Streams of the area originate on the slopes of the Henry Mountains or in the high plateaus west of the Waterpocket Fold (Figure 2). In the study area the Fremont River is the only perennial stream, though in many years Sand and Sweetwater Creeks also flow continuously. The remaining channels are active only during the spring melt season or after occasional summer thunderstorms. In 1980 almost all major streams were in continuous arroyos that developed by headward erosion after a flood on the Fremont River in 1896. Although Hunt, Averitt, and Miller (1953) contended that the flood occurred in 1897, Gregory (1918) identified the date as 1896 from eye-witness accounts. Headward erosion was arrested in most systems by resistant bedrock outcrops in the channels near mountain slopes. Above these outcrops discontinuous arroyos, not analyzed in the present study, occur in scattered locations. The Henry

Study Area

The Henry Mountains form a series of five isolated peaks of granite rock rising above folded sedimentary rocks of the Colorado Plateau in south-central Utah (Figure 2). The climate of the area is mostly semi-arid, with vegetation communities commonly ranging from barren shale surfaces to pinyon pine and juniper on the uplands; spruce and fir forests cloak the high peaks. Picard (1980) provides the most recent regional overview.
Table 1. Data Sources for the Henry Mountains, Utah

<table>
<thead>
<tr>
<th>Year</th>
<th>Source, Data</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1853</td>
<td>John C. Fremont, 5th Expedition, Description</td>
<td>Carvallo 1857</td>
</tr>
<tr>
<td>1872</td>
<td>Almon H. Thompson, Kanab-Dirty Devil Expedition, Description</td>
<td>Gregory 1939</td>
</tr>
<tr>
<td>1873–1874</td>
<td>Almon H. Thompson and John L. Graves, Topographic map</td>
<td>Stegner 1954 (account)</td>
</tr>
<tr>
<td>1875–1876</td>
<td>G. K. Gilbert, Extensive Description and Analysis</td>
<td>Gilbert 1876, 1877</td>
</tr>
<tr>
<td>1883</td>
<td>Augustus D. Ferron, General Land Office Survey, Description, Map</td>
<td>Ferron 1883</td>
</tr>
<tr>
<td>1889–1954</td>
<td>Church of Jesus Christ of Latter-Day Saints Records, Description</td>
<td>Snow 1977</td>
</tr>
<tr>
<td>1909</td>
<td>James R. Stewart and Robert J. Bryant, General Land Office Survey, Description, Map</td>
<td>Stewart and Bryant 1909</td>
</tr>
<tr>
<td>1918</td>
<td>Herbert E. Gregory, Notes and Description</td>
<td>Gregory 1918</td>
</tr>
<tr>
<td>1933–1939</td>
<td>Charles B. Hunt, Description and Analysis, Map</td>
<td>Hunt, Averitt, and Miller 1953</td>
</tr>
<tr>
<td>1952–1953</td>
<td>John F. Smith Jr., Description and Analysis</td>
<td>Smith et al. 1963</td>
</tr>
<tr>
<td>1970</td>
<td>Benjamin L. Everitt, Description and Analysis</td>
<td>Everitt 1979</td>
</tr>
<tr>
<td>1979</td>
<td>Andrew E. Godfrey, Description and Analysis (non-channel areas)</td>
<td>Godfrey 1980a, 1980b</td>
</tr>
</tbody>
</table>

Note: Several unpublished theses and dissertations have also addressed topics in the area. Most are stored at the University of Utah or at Johns Hopkins University.

Mountains are also valuable as a study area because of the availability of data on historical changes in channels and general environmental conditions (Table 1).

Stream Power

The analysis of channel form has frequently been useful in studies of fluvial systems, but the investigation of form alone would be insufficient in any study of continuous arroyos because the arroyo cross section represents a trench with a channel at the bottom. Because the total cross section is not a channel, bankfull depth of flow can rarely be defined; even the 100-year flood does not exceed the capacity of many arroyos. An analysis of sediment transport would be instructive regarding processes, but in the ephemeral channels of most arid/semi-arid regions such data are not available. An alternative approach is to analyze total stream power, which may be computed from channel dimensions and discharge, and is related to total sediment transport (Bagnold 1966, 1977; Graf 1971). Because total stream power is a function of both channel dimensions and discharge, it is more valuable in process analysis than are the individual channel dimensions considered separately.

Total stream power is the power exerted by flowing water over a unit length of channel (Bagnold 1966, 1977). It is defined as unit stream power times the channel width,

\[ \Omega = \omega W = \rho g R S V \]  

where \( \Omega \) is the total stream power in newtons, \( \omega \) the unit stream power in newtons per meter, \( \rho \) the density of the fluid in kilograms per cubic meter, \( g \) the acceleration of gravity in meters per second per second, \( R \) the hydraulic radius in meters, \( S \) the dimensionless energy slope approximated by the channel bed slope, \( V \) the mean velocity of flow in meters per second, and \( W \) the width in meters. Equation (1) as-
sumes a steady uniform flow, which probably is not true for individual points in time, but which is acceptable for mean conditions. In applying (1) it is assumed that \( g = 9.807 \) newtons per cubic meter (Carson 1971). At-a-station sediment transport and stream power values are usually recorded in unit form (e.g., Leopold and Emmett 1976); but in the present study, total power is used to assess total sediment transport. Total stream power is geomorphologically significant because it is directly related to total sediment transport, given a steady supply of sediment that exceeds transport capacity (Graf 1971) and relatively constant roughness. Within historical periods, sediment availability has always exceeded transport capacity because the arroyos are excavated in alluvial fill and because bedrock outcrops that limit excavation are rare. The channels are generally separated from canyon walls by alluvial surfaces and do not receive direct contributions of sediments from hillslopes.

The literature on stream power is confusing at best. The present analysis uses Bagnold’s (1966) approach because it is based on applied physics and because it provides scientific explanation rather than a “black box” prediction. When Bagnold’s unit stream power (total power divided by channel width) has been cited by other authors, the mathematical form has remained consistent, but verbal definitions have been variable. Compare, for example, the statements by Leopold, Wolman, and Miller (1974, 178), Bagnold (1966), Graf (1971, 209), Allen (1977, 127), and Simons and Senturk (1978, 574). The definitions and concepts used in this paper follow Bagnold’s original statement.

Data Sources

Field measurements of channel morphology combined with discharge data can be used to calculate values of total stream power using (1). Field measurements for the study area are generally available for a variety of cross sections from three time periods: 1883, 1909, and 1979–80. The 1883 and 1909 data were extracted from the General Land Office Surveys, whereas the 1979–80 data were collected by the author, and for simplicity are hereafter referred to as the 1980 data. Width \( W \) and mean depth \( D \) in 1883 and 1909 were recorded in the surveyors’ notes, which are unusually complete. Width and mean depth in 1980 were measured in the field using a tape and optical ranging devices.

Discharge has not been gauged at any site in the remote study area, but discharges of various return intervals can be calculated using functions developed by Berwick (1962), and used by the Utah State Highway Department for bridge and culvert design. The present analysis employs the ten-year discharge because such events are small enough to have occurred several times in the study period (1872–1980), but also large enough to have performed significant amounts of geomorphic work. Discharge data for a fixed recurrence interval are required because the concept of bankfull discharge is meaningless in the case of arroyos that are not true channels.

According to the Manning equation,

\[
V = n^{-1} R^{0.67} S^{0.5}
\]

(2)

where \( V \) is the mean velocity in meters per second, \( R \) the hydraulic radius in meters, \( S \) the dimensionless friction slope approximated by channel bed slope, and \( n \) the estimated Manning roughness coefficient. Although the roughness coefficient is the product of a field estimate, the estimates can be made in a reasonably standardized fashion based on published guidelines (Aldridge and Garrett 1973; Barnes 1967; Benson and Dalrymple 1967; Ree and Palmer 1949). Hydraulic radius \( R \) in all cases was calculated as cross-sectional area divided by wetted perimeter:

\[
R = (W D) / (2D + W)
\]

(3)

where \( W \) is the width in meters, and \( D \) the mean depth in meters. Width was obtained from the 1883 or 1909 survey or measured in the field, whereas mean depth was determined by distributing the ten-year discharge across the cross section using the formula

\[
D = \left( \frac{Q n}{k W S^{0.5}} \right)^{0.6}
\]

(4)

where \( Q \) is the ten-year discharge in cubic meters per second, and \( k \) is a constant related to channel cross section shape (see Graf 1979a). The depth of flow given by (4) is then used in (1) to (3), provided that the arroyo is not filled. Where the ten-year discharge ex-
ceeded the capacity of the arroyo, \( D \) was set equal to the depth of the arroyo in the calculations using (1) to (3). Channel slope was calculated from topographic maps, with a limited sample being surveyed in the field as an accuracy check. Topographic maps proved useful in determining the channel gradients of all but the smallest streams. For historical channels the gradient of the alluvial surface obtained from modern topographic maps was combined with measurements of sinuosity on historical planimetric maps to calculate channel gradient. Topographic maps were used rather than field surveys to obtain channel gradient data because the mean gradients yielded by the maps were deemed more meaningful than the highly variable local gradients measured in the field.

Data were collected from arroyo cross sections ranging in size from small tributaries to the Fremont River. Data were available for 18 cross sections from the 1883 surveys and for an additional 18 cross sections from the 1909 surveys; all 36 historical sites were resurveyed in 1980 (Figures 3 and 4). Confidence in the analysis of the 1980 situation was enhanced by the inclusion of an additional 88 cross sections for which no historical data were available.

The historical survey data for channels in the northern Henry Mountains are of exceptional quality. Because of the nature of the surveyor contracts with the General Land Office, the surveyors’ notes include channel data where channels crossed the section lines of the township and range survey system (U.S. Department of Interior 1934). In other areas of the intermountain west, the surveyors often recorded only channel width (Cooke and Reeves 1976) or their data are so inaccurate as to be useless (Bryan 1954). In the Henry Mountains the surveyors occasionally reported channel depths. Analyses of abandoned channels that were mapped and measured in 1883 and 1909 and that appeared as remnants on the landscape in 1980 attest to the accuracy of these depth data and indicate that the surveyors measured widths orthogonal to the direction of the flow. This conclusion is supported by the fact that widths are given for channels that are aligned along section lines rather than across them. In these
cases the surveyors noted where the section line entered the channel, where it left the channel, and added to the notes a figure for channel width. Channel widths were more frequently noted than depths. Where both measures were available, they corresponded closely with the widths and depths measured on remnants noted by Hunt, Averitt, and Miller (1953) and by the author in 1980. The present work uses only those sites where width and depth were recorded by the earlier surveyors. The quantity of data is thereby decreased, but the quality is preserved.

Two major problems with the analysis of downstream changes in total power involve sediment and climatic stationarity. An analysis of stream power alone does not take into account temporal changes in the sediment supplied from hillslopes. In the semi-arid Henry Mountains, land use has had a significant but unquantified impact on hillslope systems. Grazing and mineral exploration have produced surficial disruptions and probable changes in sediment production, but the dimensions of the effects are unknown.

In this study the calculated ten-year discharge is the most useful measure of discharge. The ten-year event can be calculated with reasonable confidence in the absence of gauging data, but it remains a calculated rather than an observed value. The technique relies on data primarily from the pre-1960s period, and its extension to periods such as the 1880s involves an implicit assumption of climatic stationarity. Instrumented records in other parts of the intermountain west reveal climatic changes of varying magnitude, and such changes undoubtedly occurred in the Henry Mountains (Bradley 1976). However, lack of detailed climatological studies prevents this climatic variation from being taken into account in the present study. It is likely that climatic and land use changes have affected the magnitudes of downstream power changes but not the trends.

**Downstream Changes in Stream Power**

In the late 1800s most of the streams of the Henry Mountains areas flowed on surfaces of extensive alluvial deposits. Although some gullies were reported by Thompson in 1872 (Gregory 1939), most valley floors had well-watered soils and dense grass covers with grazing potential (Snow 1977). Work by Gil-
Figure 5. An unnamed creek below Avery Springs (see Figure 2 for location). Top: Hunt's field camp on the left bank during a flood in 1934. During this period when the reach had a downstream increase in total stream power, erosion is pronounced. Hunt Photo 462, U.S. Geological Survey Photographic Library, Denver. Bottom: same view as above, except as seen in 1980. The reach now has less of a downstream increase in total stream power and has experienced considerable deposition. Photo by author.
Figure 6. View looking south across Muddy Creek into the mouth of the Fremont River (see Figure 2 for location). Top: in 1939 both streams were relatively powerful and transported large quantities of material. Northern Henry Mountains in the background. Hunt Photo 840, U.S. Geological Survey Photographic Library, Denver. Bottom: same view as above, except as seen in 1980. The Fremont River is relatively less powerful than Muddy Creek, and sediment is accumulating in the Fremont channel. Photo by author.
bert (1876, 1877) and Ferron (1883) showed that channels were meandering and were much smaller than are modern channels in the area (Figures 3 and 4). In 1896 severe floods on the master stream, the Fremont River, excavated a relatively deep, wide arroyo and destroyed irrigation works, town sites, and floodplain fields (Gregory 1918). Several floods during the next two decades extended the arroyo by headward erosion through tributary systems (Hunt, Averitt, and Miller 1953). Within twenty years, sediments that had been stored throughout the channel network were evacuated. Between 1940 and 1980, however, a reversal of the process has occurred in the Fremont River, and sediments delivered to the master stream by tributaries have not been carried farther downstream (Figures 5 and 6) (Hunt, Averitt, and Miller 1953). It appears that this historical change in sediment distribution involved changes in the total stream power distribution.

Historical changes in total stream power reflect the influence of trends in unit stream power combined with channel width. In 1883 stream channels were relatively small and narrow, so that the combination of unit power and channel width produced a decline in total power in the downstream direction (Figure 7 and Table 2). As previously noted, 1883 was a period of system-wide deposition, a circumstance not surprising in view of downstream decline in total power. Irrespective of sediment sizes, increasing amounts of sediment would have been supplied to the system as drainage areas increased downstream, yet declining amounts of total power were avail-

**Table 2. Calculations for Stream Power**

<table>
<thead>
<tr>
<th>Regression Statistic</th>
<th>1883</th>
<th>1909</th>
<th>1980</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a$</td>
<td>3.46</td>
<td>2.57</td>
<td>2.74</td>
</tr>
<tr>
<td>$b$</td>
<td>-0.24</td>
<td>0.65</td>
<td>0.58</td>
</tr>
<tr>
<td>Number of Observations</td>
<td>18</td>
<td>18</td>
<td>122</td>
</tr>
<tr>
<td>Correlation Coefficient</td>
<td>-0.32</td>
<td>0.96</td>
<td>0.86</td>
</tr>
<tr>
<td>Level of Significance</td>
<td>0.29</td>
<td>0.001</td>
<td>0.001</td>
</tr>
<tr>
<td>Standard Error Estimate</td>
<td>0.42</td>
<td>0.09</td>
<td>0.27</td>
</tr>
</tbody>
</table>

Notes: Calculations for function $\log \Omega = a + b \log Q_{10}$, where $Q_{10}$ = ten-year peak discharge in $m^3 s^{-1}$ and $\Omega$ = total stream power in $N$. 1980 data include 36 relocated and resurveyed cross-sections plus 88 additional sites.
able for sediment transport. Basin-wide storage of sediment resulted.

In 1909, during the major erosion episode, the situation reversed, with total power increasing in the downstream direction (Figure 7 and Table 2). If the sediments forming the bed and banks consisted of small enough particles to be readily entrained and were supplied in quantities that did not exceed total transport capacity, system-wide transportation and erosion would have been the logical consequence of such a distribution of total power. Such conditions did indeed prevail, according to witnesses (Hunt, Averitt, and Miller 1953).

The least-squares line overpredicts total stream power for the largest and the smallest streams at cross sections where sedimentation was occurring in 1980 (Figures 8 and 9, Table 2). Wide, shallow channels with low total stream power occurred at these extremes. The least-squares line underpredicts total stream power for many cross sections in the region where the ten-year discharge is 100 to 300 m³s⁻¹ and that were spatially coincident with erosion zones in 1980. Narrow, deep channels with high total stream power were the result. The least-squares line is therefore a simplification of a complex situation where there is no system-wide equilibrium.

**Influence of Individual Variables**

An explanation of the downstream variation in total stream power requires a disentangle-ment of the influences of a variety of channel dimensions. Two general sets of conditions may be distinguished: first, the pre-1896 period when the ten-year discharges were not contained within most channels and, second, the post-1896 period when the ten-year discharges were contained by most arroyos. The differentiation between the overbank and within-arroyo cases is significant because in the former situation only a small channel is available for sediment transport downstream. In the latter case, however, all the water is contained within the arroyo walls and contributes to downstream sediment movement. In the overbank case the equation of continuity was not preserved, and only the water remaining in the channel was analyzed for the power generated. The channel variables influencing total stream power are illustrated by substituting the Manning equation (Equation 2) into the equation defining total stream power (1). Simplification, combination, and substitution of mean depth \( D \) for hydraulic radius \( R \) produces

\[
\Omega = \rho g n^{-1.00} D^{1.67} S^{1.50} W. \tag{5}
\]
Table 3. Calculations for Channel Characteristics

<table>
<thead>
<tr>
<th>Dependent Variable (y)</th>
<th>Regression Statistic</th>
<th>1883 (n = 18)</th>
<th>1909 (n = 18)</th>
<th>1980 (n = 36)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth of Flow (m)</td>
<td>a</td>
<td>-0.49</td>
<td>-1.10</td>
<td>-1.03</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>0.26</td>
<td>0.54</td>
<td>0.49</td>
</tr>
<tr>
<td></td>
<td>r</td>
<td>0.69</td>
<td>0.73</td>
<td>0.74</td>
</tr>
<tr>
<td>Width of Flow (m)</td>
<td>a</td>
<td>0.20</td>
<td>-0.20</td>
<td>0.35</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>0.26</td>
<td>0.73</td>
<td>0.54</td>
</tr>
<tr>
<td></td>
<td>r</td>
<td>0.48</td>
<td>0.82</td>
<td>0.70</td>
</tr>
<tr>
<td>Channel Slope</td>
<td>a</td>
<td>-1.55</td>
<td>-1.43</td>
<td>-1.23</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>-0.28</td>
<td>-0.32</td>
<td>-0.43</td>
</tr>
<tr>
<td></td>
<td>r</td>
<td>-0.57</td>
<td>-0.87</td>
<td>-0.80</td>
</tr>
<tr>
<td>Roughness</td>
<td>a</td>
<td>—</td>
<td>—</td>
<td>-1.78</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>—</td>
<td>—</td>
<td>0.09</td>
</tr>
<tr>
<td></td>
<td>r</td>
<td>—</td>
<td>—</td>
<td>0.41</td>
</tr>
</tbody>
</table>

Notes: Calculations for the function log y = a + b log \( Q_{10} \), where \( Q_{10} \) = ten-year discharge and y = channel characteristic. Insufficient data available for roughness evaluation for 1883 and 1909. Roughness estimated with the Manning roughness coefficient (values from 0.017 to 0.065). The 1980 sites are the relocated versions of the 1883 and 1909 sites.

Assuming \( \rho g \) is constant, each of the variables on the right side may be considered in turn. Judging from the exponents in (5), total power appears to be most sensitive to changes in depth and slope, and somewhat less sensitive to changes in roughness and width. However, Table 3, which shows the relationships among depth, width, slope, and roughness, indicates that depth and slope change in the downstream direction at about the same rate, but in opposite directions. Consequently, they approximately cancel one another out in their influence on total power. No data are available for roughness in either 1883 or 1909, but in modern channels it varies very little downstream (exponent value of only 0.09 when related to discharge), and sparse descriptions from historical sources suggest that it may be eliminated as a useful explanatory variable. It therefore appears that the remaining variable of width is the one most useful in understanding temporal and spatial variations in total stream power.

When erosion of the pre-1896 Fremont River channel occurred, the narrow meandering configuration was widened by a factor of five to ten. Tributary channels were widened to a lesser degree. The exponents (b values) in Table 3 show that before the erosion episode, width increased in the downstream direction at only one-third the rate of increase in 1909, and at only one-half the rate in 1980. Depth increased at only about one-half the post-erosion rates.

In the second set of conditions, those where the ten-year flows were contained within the arroyo, a different approach to determining the influence of variables is possible. If the arroyo capacity is not exceeded, the equation of continuity is preserved, and

\[
Q = WDV. \tag{6}
\]

Assuming that mean depth \( D \) is a reasonable surrogate for hydraulic radius (a safe assumption, given that the channels have widths 20 or more times their depths) and substituting (6) into (1), total stream power may be computed from

\[
\Omega = \rho g QS. \tag{7}
\]

Assuming that \( \rho g \) is constant, in the 1909 and 1980 cases where the ten-year peak flow is contained within the arroyo, total stream power varies with discharge and slope. Slope declines in the downstream direction, whereas discharge increases, so the two variables operate in opposition to one another in their influence on total stream power. As shown in Tables 2 and 3, the rate of downstream increase in discharge is greater than the rate of decline in slope, so total power increases downstream in the 1909 and 1980 cases.

The results of the analysis show the importance of channel morphology in the sediment
transport process. In the 1883 channels, ten-year floods were dissipated across floodplains and the total stream power in the channels was significantly limited by their small sizes. After the catastrophic flood of 1896 altered the configuration of the channels, all the waters of the ten-year flood were confined between the arroyo walls and were available for downstream sediment transport. Under 1883 conditions, floods larger than the ten-year discharge operated in the same fashion as those analyzed here, except for the very large 1896 flood (with a 250-year or greater recurrence interval), which combined with human actions in and near the channel to destroy completely the original channel system. Under the 1909 and 1980 conditions, large floods are contained within the arroyos and do not have so great an impact on the channel system. Large floods in the 1920s and in 1981 did not cause extensive erosion along the channel. Slow accumulation of sediments, a process that has already begun, presumably will eventually refill the arroyos. (Figures 5 and 6; see also Emmett 1974; Leopold 1976).

The present results also indicate that, although visually the most remarkable change in the channels as a result of the 1896 flood occurred in depth, interactions among width, discharge, slope, and stream power are the more significant from a geomorphological point of view. This situation is partly due to the fact that in many arroyos some of the depth is unused in the sense that it is rarely occupied by flowing water, whereas the width is fully occupied by relatively frequent flows. In the pre-erosion period, overbank flows were common, so the channels carried relatively little sediment and most materials were stored in floodplain deposits.

Cause and Effect

The direction of causality between channel morphology (represented in the present study by channel dimensions) and channel processes (represented by stream power) is difficult if not impossible to establish in the multi-century framework of a continuously operating stream system (Embleton and Thornes 1979, 2). Because fluvial systems are characterized by strongly developed feedback mechanisms (Chorley and Kennedy 1971), the establishment of the direction of causality is reasonable only for limited periods of time. For humid-area streams, constant mutual adjustment may occur between form and process, but in arid and semi-arid streams the discontinuous operation of the fluvial system precludes such mutual adjustments. The evidence from the Henry Mountains streams suggests that two situations prevail: one when processes control forms during catastrophic flow events, and the other when forms control processes during smaller flow events.

During high-magnitude, low-frequency events, such as the 1896 flood in the Fremont River, so much water is available that previously defined channels are completely destroyed. Such events could be viewed as transgressions of geomorphic thresholds (Schumm 1979), or as transgressions of folds in catastrophe surfaces (Graf 1979b); in either view the system undergoes drastic change. The change is followed by an adjustment or relaxation period, during which a new order is established (Graf 1977). In the case of the Henry Mountains, headward erosion from the Fremont River arroyo established a new order within one or two decades.

Radical adjustments by the channel network might be caused by either a large flood or internal adjustments, as suggested by Schumm (1979) and Hey (1979). The excavation of arroyos in the Henry Mountains is clearly related to a single flood event (Gregory 1918; Hunt, Averitt, and Miller 1953). The series of events outlined above that give rise to the transportation and deposition of sediment in the Henry Mountains is similar to the series outlined by Hey (1979), though the backwater effects that he mentioned do not appear to be significant in the present study area.

Conclusions

Data from the Henry Mountains permit tentative answers to the questions with which this paper began. First, total stream power increases systematically downstream under present conditions in response to downstream increases in discharge and the containment of water by arroyo walls. Second, total stream power distributions have changed
radically in the past century. In the 1880s, when the area was experiencing system-wide deposition, total stream power declined in the downstream direction. Overbank flows were common. In 1896 the master stream experienced a catastrophic flood, which disrupted the system and initiated arroyo development. In the early 1900s, as a result of system-wide erosion, total stream power increased downstream, an arrangement conducive to sediment evacuation. In 1980 the rate of downstream change in total power was intermediate between the depositional conditions of the 1880s and the erosional conditions of 1909; deposition was occurring in the smallest and largest channels, but not in mid-basin areas (Figures 1, 5, 6). Third, processes in semi-arid streams appear to control channel morphology only during extreme events; most of the time forms control processes.

The results of this study lead to some additional conclusions of a general nature. Since 1896 the arroyo systems of the northern Henry Mountains have not transgressed major geomorphic thresholds or catastrophe-surface folds in their operations. Significant adjustments have occurred within the confines established by the episode of erosion beginning in 1896 and there continues to be substantial spatial variation of process intensity during the between-threshold or non-fold period. In the 1960s and 1970s research in fluvial geomorphology emphasized thresholds and changes through time (e.g., Williams 1978; Schumm 1979). The results of the present study indicate that fruitful avenues of inquiry for the 1980s lie in the realm of change across space with an emphasis on disequilibrium.

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References


Bryan, K. 1954. The geology of Chaco Canyon, New Mexico, in relation to the life and remains of the prehistoric peoples of Pueblo Bonito. Smithsonian Miscellaneous Collections, Washington: Smithsonian Institution.


