Incision and morphologic evolution of an alluvial channel recovering from hydraulic mining sediment

L. ALLAN JAMES *Department of Geography, University of South Carolina, Columbia, South Carolina 29208*

ABSTRACT

Hydraulic gold mining in the Sierra Nevada, California, produced such a large volume of sediment from 1853 to 1884 that channels could not carry a substantial proportion of the delivered sediment. The lower Bear River aggraded as much as 5 m and underwent a major avulsion during the 1870s. Channel morphology continues to respond to these changes more than 100 years after the cessation of most hydraulic mining. Textures of mine tailings and older alluvium are contrasted, and channel incision and channel-shape changes from 1930 to 1985 are documented.

Unlike the Yuba and Sacramento Rivers, the lower Bear River had not returned to premining base levels by 1950 as sediment loads decreased. Nor were incision rates or morphological change governed by reduced sediment loads below dams built in 1928 and in the mid-1960s. Channel stability was due to resistance of a cohesive stratum on the channel bed. An episode of channel down-cutting was instigated by a moderately large flood in December 1955 that pierced the cohesive layer and was sustained through the 1970s by moderate-magnitude floods. The 1955 flood did not directly erode large volumes of sediment, but it destabilized the channel, allowing smaller floods to erode the channel at an accelerated rate.

Episodic channel-bed incision is contrasted to progressive morphologic changes. Cross-section shape narrowed and deepened steadily from 1930 through the mid-1970s as the channel was superimposed onto cohesive older alluvium. Progressive deepening set up a positive feedback that ensured the ultimate penetration of the paved bed. Channel morphologic readjustment from the avulsion thus was both progressive and episodic.

INTRODUCTION

The Bear River flows out of deep narrow valleys of the northern Sierra Nevada foothills between the Yuba and American Rivers onto a broad alluvial plain that merges with the Sacramento Valley (Fig. 1). The channel continues to readjust to an avulsion that occurred in the 1870s as a result of episodic delivery of sediment produced by hydraulic gold-mining (Mendell, 1880; Kelley, 1959). Most of the unconsolidated mine tailings have been eroded from low-flow channels in the lower basin, exposing older, cohesive alluvium in the channel bed, although large volumes of tailings remain stored on flood plains.

This paper examines channel morphologic effects of contrasting sediment types at the alluvial boundary and the complex history of sedimentation at a cross section on the lower Bear River. Stream-flow measurement records from 1929 to 1985 from a gaging station and cableway near Wheatland (Fig. 2) are analyzed to determine the magnitude and frequency of flooding and the timing and character of changes in channel morphology and bed elevation. Cableway data ensure that measurements were repeated at the same location and provide an opportunity to reconstruct long-term morphological changes in a channel recovering from severe aggradation and avulsion.

Fluvial-channel erosion has historically been an important concern of geomorphologists, geologists, hydrologists, and engineers. Channel incision can destroy valuable land and structures (Lane, 1937; Wolman, 1959), determines longitudinal-profile development (Mackin, 1948), influences sediment budgets (Lane and Borland, 1954; Williams, 1982), complicates paleohydrologic reconstructions (Costa, 1983; Kochel and Baker, 1982), and has great bearing upon levee and bridge maintenance, stream-flow gaging, and flood prediction. Field evidence of 20th-century channel erosion in the lower Bear River includes scoured bridge pilings, over-steepened levee bases, and abandoned irrigation intakes perched above the present water surface. Three hypotheses of the cause of this channel incision are tested by this study: (1) rates of scour increased in response to trapping of sediment by reservoirs, (2) incision was progressive and was completed by 1950 (such as incision of the Sacramento and Yuba Rivers) presumably due to decreased sediment storage and reworking (Graves and Eliab, 1977; see also Gilbert, 1917), and (3) rates of incision and morphologic change were dominated by the resistance of cohesive alluvium in the bed and bank. The timing of floods and channel incision, and the character of bed and bank materials are documented and used to evaluate these competing hypotheses.

A test is also made of the hypothesis that the Bear River narrowed as it incised through noncohesive tailings into the clay-cemented, pre-mining alluvium. The morphology of many mid-latitude channels has evolved toward deeper, narrower shapes due to progressive decreases in sediment caliber during post-Pleistocene, valley-bottom aggradation (Fisk, 1944; Schumm, 1963). The possibility of a similar change in the lower Bear River is examined through analysis of channel cross-section shapes.

Sedimentation and Erosion History

Evolution of the Drainage Basin. A full understanding of recent channel morphologic responses depends on knowledge of the history and stratigraphy of the channel. The Bear River heads in structurally deformed Paleozoic and Mesozoic rocks of the Sierra Nevada belt (Clark, 1976; Schweickert and Cowan, 1975) and flows out onto relatively flat-lying Mesozoic and Cenozoic strata underlying the Sacramento Valley (Davis and Olmsted, 1952). The present east-west drainage of most Sierra valleys developed during periods of Miocene-Pliocene volcanism that filled earlier, north-striking Tertiary channel systems (Lindgren, 1911; Durrell, 1959; Marchand and Allwardt, 1981). Between 3.6 and 1.8 Ma, westward-flowing consequent channels deposited sediment at the edge of the Sacramento Valley in a broad plain comprising the Laguna Formation (Piper and others, 1939; Marchand and Allwardt, 1981; Busacca, 1982). Uplift of the Sierra block to the east led to trenching and extension of Laguna fans through the Quaternary, causing the steepened channels to incise along the Sacramento Valley margin

Geological Society of America Bulletin, v. 103, p. 723–736, 16 figs., 3 tables, June 1991.

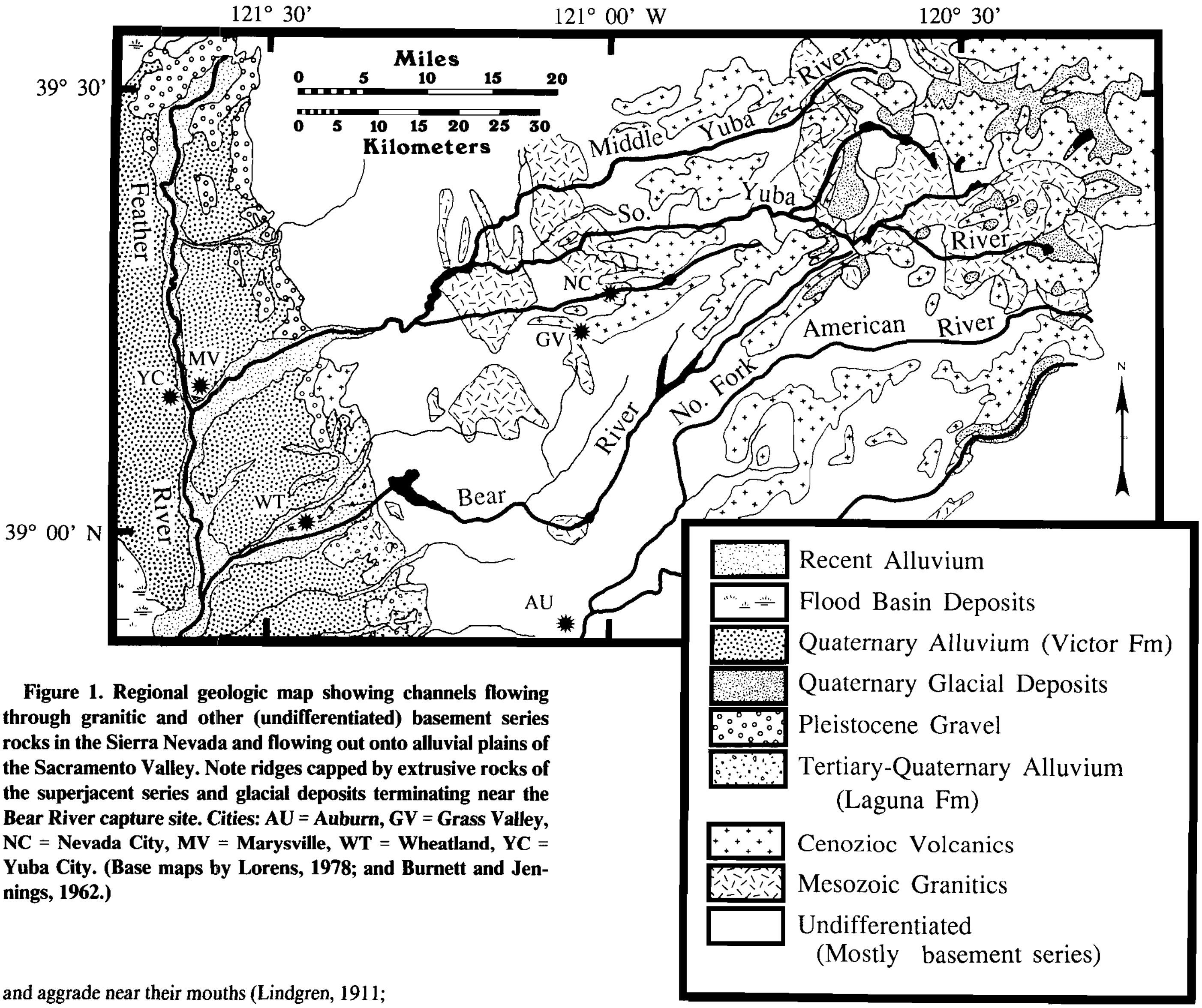


Figure 1. Regional geologic map showing channels flowing through granitic and other (undifferentiated) basement series rocks in the Sierra Nevada and flowing out onto alluvial plains of the Sacramento Valley. Note ridges capped by extrusive rocks of the superjacent series and glacial deposits terminating near the Bear River capture site. *Cities:* *AU = Auburn, GV = Grass Valley,* NC = Nevada City, MV = Marysville, WT = Wheatland, YC = Yuba City. (Base maps by Lorens, 1978; and Burnett and Jennings, 1962.)

and aggrade near their mouths (Lindgren, 1911; Piper and others, 1939; Axelrod, 1962; Christiansen, 1966). This dissected alluvial-fan surface forms the modern highlands known as the "Red Lands" (Bryan, 1923).

Most Sierra channels aggraded in response to Quaternary glacial and periglacial sediment production, and subsequent incision left alluvial terraces along most foothill channels where they enter the Central Valley (Bryan, 1923; Davis and Hall, 1959; Marchand and Allwardt, 1981). Glacial striae, geomorphic surfaces, and stratigraphic evidence indicate that the Bear River headed near the Sierra crest prior to late Quaternary capture by the South Yuba River (Fig. 1) (James, 1990). Quaternary terraces along the lower Bear River are inset between the higher Laguna surface and were buried by hydraulic mining sediment. This older alluvium has resisted erosion by the modern channel that was superposed upon them.

Historical Sedimentation. Hydraulic gold-mining from 1853 to 1884 generated high sediment loads and caused severe aggradation of many Sierra Nevada channels (Gilbert, 1917). As much as $270 \cdot 10^6 m^3$ of sediment was produced by mining in the Bear Basin, second only to the volume generated in the much larger Yuba Basin (Gilbert, 1917; James, 1989). Tailings were deepest in the mining districts where they attained depths up to 40 m in the late 1870s (Whitney, 1880). Very little sediment was stored at intermediate elevations of the Bear Basin due to steep slopes and narrow gorges (Fig. 2). Extensive deposition in low gradient reaches of the lower basin began in 1862 when an extreme flood filled low-flow channels with sand and gravel (Keyes, 1878; Mendell, 1881).

Sedimentation rates in the lower basin increased through the 1870s, when an avulsion shifted much of the channel up to 1 km (Fig. 3) (Hall, 1880; Mendell, 1881). Contemporary accounts indicate that the avulsion took place over a period of several years in the late 1870s, and that channel positions were influenced by sloughs and levees (Keyes, 1878; James, 1988). Between 1880 and 1882, a 1.8-m-high brush and rock sediment detention dam impounded ~735,000 m^3 of sediment (Mendell, 1881, 1882). By 1883, the railroad bridge near the Wheatland gage site had been raised three times, a total of more than 5 m (Von Geldern, 1891). Coring of deposits reveals that tailings deposited during the 19th century remain up to 5.3 m deep and extend 2 to 3 km across valley bottoms (Fig. 4) (James, 1989).

Channel Incision. Hydraulic mining was enjoined in 1884, and subsequent decreases in sediment loads resulted in incision by most

TABLE 1. MAJOR DAMS AND RESERVOIRS ON THE BEAR RIVER

Dam	Drainage area (km^2)	Year completed	Reservoir capacity* (m^3 · 10^3)
Camp Far West 1	738	1928	6,200†
Camp Far West 2	738	1963§	127,050
Van Geisen (Combie)	337	1928	6,852
Rollins	269	1965	81,411

Source: California Department of Water Resources, 1988.
†California Water Resources Board, 1955.
*Includes dead storage.
§Enlarged, 1963.

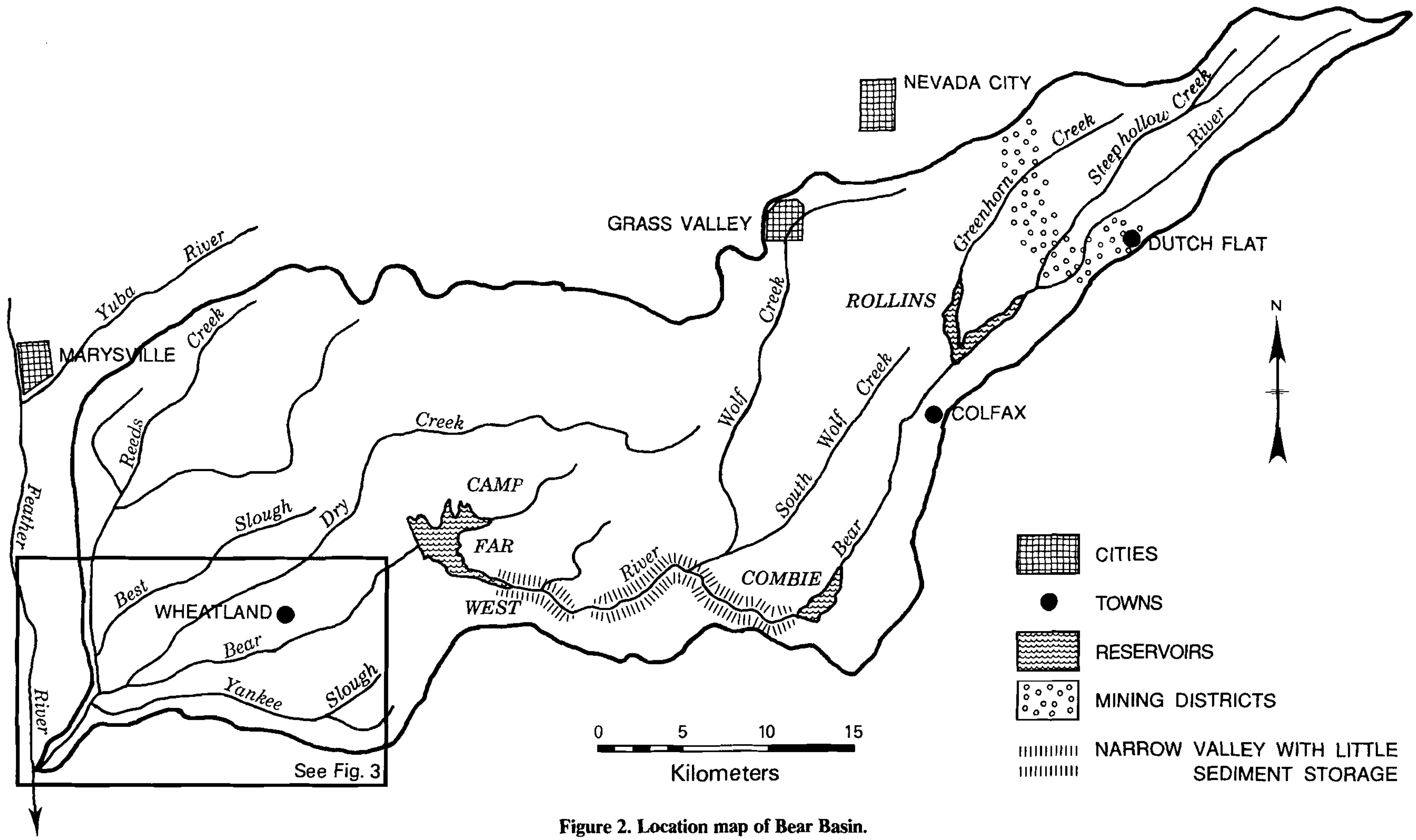


Figure 2. Location map of Bear Basin.

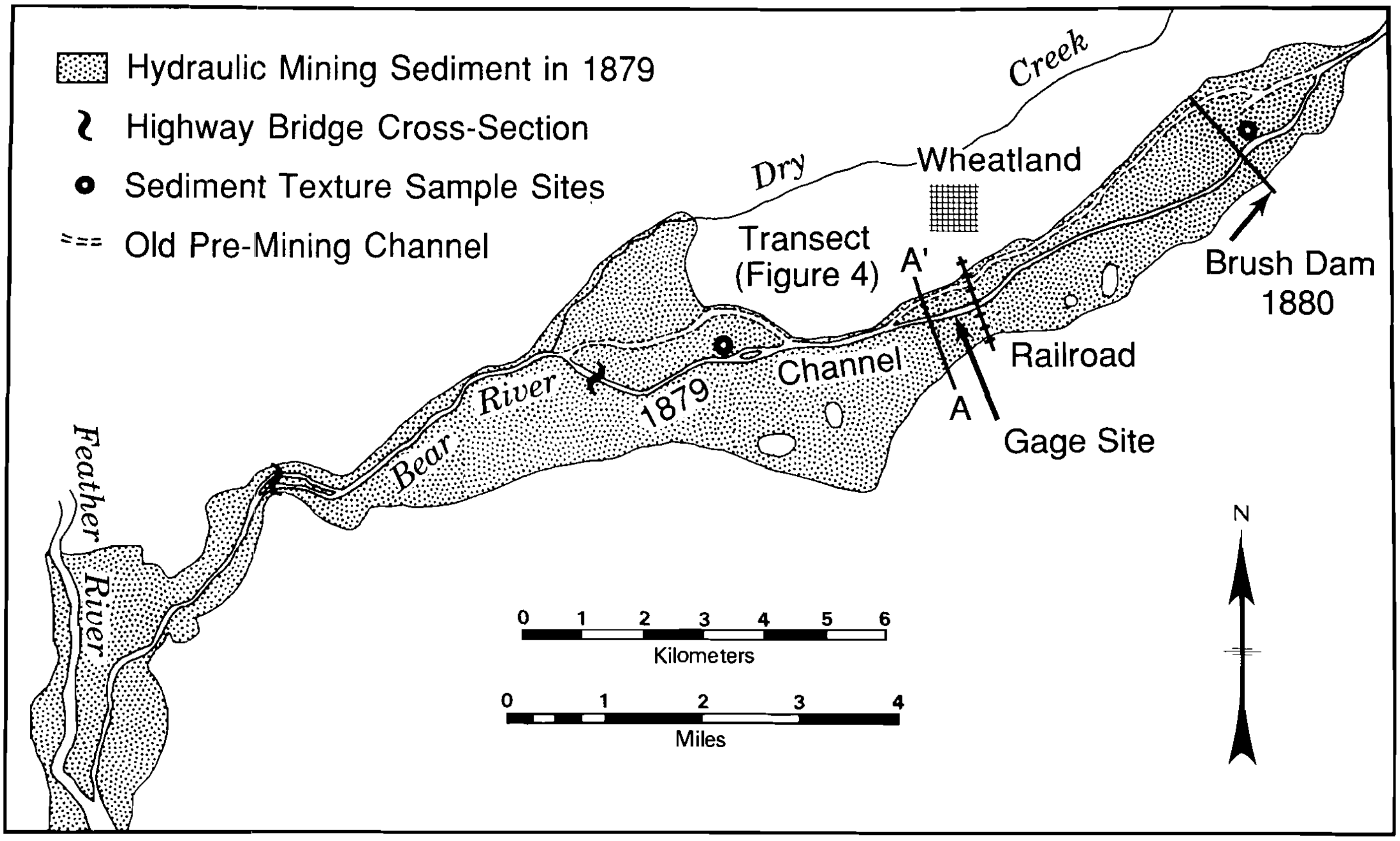


Figure 3. Map of lower Bear River 1879 survey of hydraulic mining sediment deposits by Wm. H. Hall (Mendell, 1880). Highway 70 Bridge survey site is downstream of the Pleasant Grove Bridge site.

channels draining hydraulic mines (Gilbert, 1917). The channel at the Wheatland gage site had begun to degrade by 1890, although concurrent overbank deposition probably continued (Von Geldern, 1891). By 1908, low terraces had formed in the Bear "piedmont," suggesting that channel incision continued to dominate along the lower Bear (Gilbert, 1917, p. 28).

Prior to 1928, sediment was transported to the lower Bear by reworking of large deposits in the upper basin, but this sediment is now detained by dams. Both Camp Far West and Van Giesen (Combie) dams were built in 1928; in the mid-1960s, Camp Far West was greatly enlarged, and Rollins Dam was constructed (Table 1; Fig. 2). Trapping of sediment behind these dams lowered sediment loads and presumably encouraged channel incision in the lower Bear. Some incision has continued to the present, although most tailings along the lower Bear River remain stored in an extensive flood plain pro-

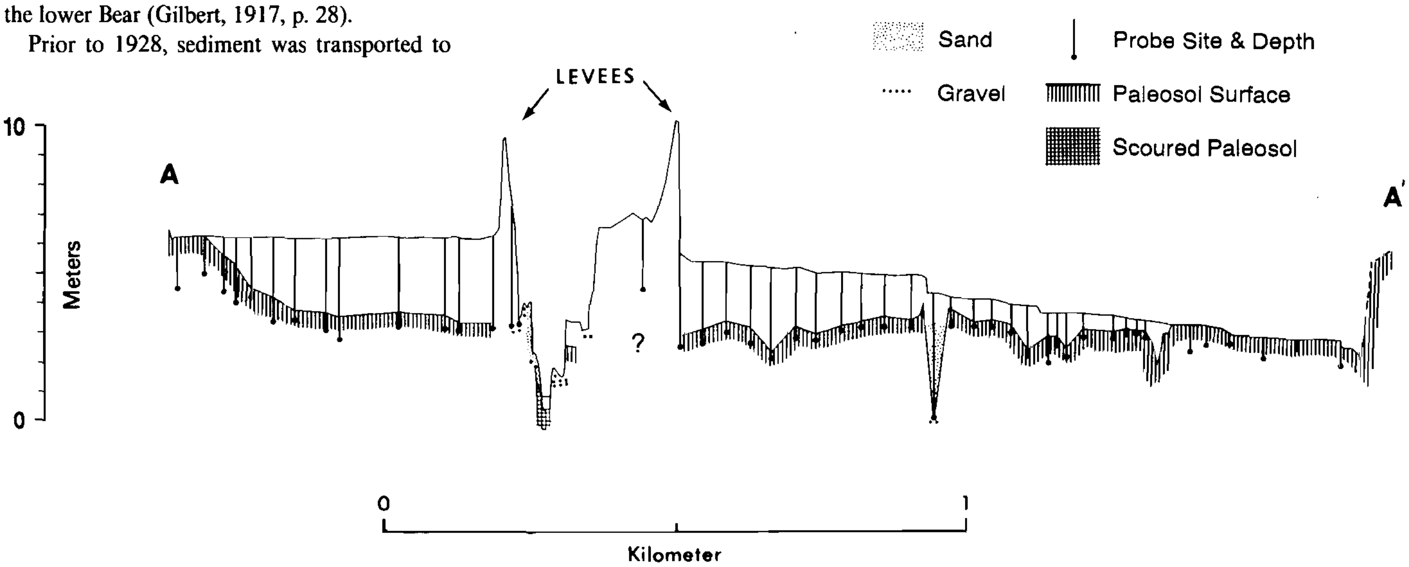


Figure 4. Valley cross section about 200 m downstream of the Wheatland gage. Depths of mining sediment, determined by coring, are about 3 m at this site but reach as much as 5.3 m downstream (see Fig. 7a). Note deep, narrow pre-mining channel about 0.7 km north of present channel.

tected from channel erosion by levees (Fig. 4). Portions of the channel between the Wheatland gage and Camp Far West Dam were actively migrating during the 1980s.

DATA AND ANALYTICAL METHODS

Channel-incision rates may have responded to (1) dam construction, (2) decreasing sediment loads, or (3) resistance of channel boundary materials. To identify the controls on channel incision, the magnitude and frequency of historic floods are examined, the timing of incision is compared with dam construction dates, and textures of tailings and underlying alluvial strata are contrasted. Finally, the timing of changes in channel cross-section shape is examined and compared with the timing of channel incision.

Flood Frequency Analysis

The annual maximum flood series at the Wheatland gage (U.S. Geol. Survey Station #11424000, 1929–1986) was analyzed to characterize the frequency and variability of large flows at the cross-section site. Statistical analysis was conducted with Consolidated Frequency Analysis, a Fortran computer program (Pilon and others, 1985). Data were tested for nonrandom trends to ensure that assumptions of flood-frequency analysis were not violated (Water Resources Council, 1981). The annual series was split at 1965 and a Mann-Whitney difference of means test was used to test the null hypothesis that there is no difference between floods occurring in the two periods before and after dams were closed upstream of the gage. This test was performed both with and without the record 1986 flood.

Spearmans rank-order serial correlation coefficient was calculated to test for serial (lag one) autocorrelation in the flood series and to ensure that the assumption of independence between events is not violated. A Mann-Whitney difference of means test based on month of annual flood occurrence was used to test for significant differences in floods occurring in different seasons. The flood series was divided into two groups: floods occurring from November to March, and those occurring from April to July. Finally, flood frequencies were calculated using the entire annual maximum series with a three-parameter lognormal frequency distribution (Pilon and others, 1985).

Sediment Textures

Mining sediment in the lower Bear Basin can be seen in the field to be texturally distinct and less cohesive than the older alluvium upon which it rests. To demonstrate these distinctions, 29 samples were collected from 2 sites (Fig. 3). By coring a 5-m-thick deposit at a site 4 km downstream of the gage, 15 tailings samples and 3 buried soil samples were collected. In addition, 11 pre-mining soil samples were collected at 10-cm vertical intervals below the A horizon of a soil exposed in a stream bank about 6 km upstream of the gage.

Mechanical particle-size analysis was performed by standard methods (Day, 1965). Sample preparation included removal of gravel by wet sieving through a 2-mm sieve, oxidation of organic matter with 30% H_2O_2, and disaggregation with sodium hexametaphosphate. Sands were sieved to 1-phi intervals with a sonic sifter, and silt-clay percentages were determined by hydrometer analysis. Sand-silt and silt-clay divisions are defined at 4.0 and 8.0 phi (62 and 4 microns), respectively (Wentworth, 1922). Graphic inclusive moments were calculated for particle-size distributions using equations given by Folk and Ward (1957).

Channel Incision

Channel incision from 1930 to 1985 is documented through graphical and statistical analysis of U.S. Geological Survey stream-flow and sounding data at the Wheatland gage (hydrographers' notes, U.S. Geol. Survey Archives). The channel at this site is relatively straight (sinuosity = 1.07), although pool and gravel-bar formation in recent decades has resulted in increased sinuosity of the low-flow channel. Cross-section, flow-stage, and thalweg elevations are expressed relative to a zero value equivalent to the local gage datum at 21.92 m above mean sea level. Vertical control has been well maintained on these sections, and longitudinal position is fixed by the cableway which has been maintained at the same cross section since its establishment in 1930 (Thomas Hankins, U.S. Geol. Survey, Sacramento). Effects of downstream drift of current meters during the gaging of high flows is minimal due to the U.S. Geological Survey practice of using weights and correction factors if necessary (Rantz and others, 1982). To ensure that the entire channel boundary is represented, flow data were sampled by choosing the largest flood measurement for most years plus a random sampling of smaller events. The selection of samples is biased toward large floods, therefore, but constitutes a random sample of flows in all other respects.

Statistical analysis is conducted on the morphological variables to identify long-term trends. Discharge explains much of the variance in flow width, depth, velocity, or stage at a station (Leopold and Maddock, 1953; Brush, 1961), and this variation due to discharge must be isolated before temporal changes in the other parameters can be identified. Discharge is standardized, therefore, by regressing morphological variables on flow magnitude and analyzing the regression residuals (difference between observed and predicted values of the dependent variable). The integrity of this method is constrained by the validity of the regression model used. Power functions have been defended on practical and theoretical grounds (Leopold and Maddock, 1953; Williams, 1978) but have been criticized elsewhere (Richards, 1976). Power functions are employed because (1) they are simple, practical, and conventional; (2) alternative polynomial models are less robust; and (3) Bear River morphologic changes are so pronounced that differences between power function and polynomial models have little bearing on results. Power functions were derived with univariate regressions of natural log values of hydraulic parameters and discharge (Leopold and others, 1964). Regression residuals thus represent the observed minus the predicted natural logs of dependent variables.

Cross Sections. More than 30 cross sections were plotted from soundings made at the cableway between 1931 and 1984 to depict explicit morphological changes through time. Horizontal distances are constrained by a 1984 topographic survey conducted with rod and level for this study.

Stage-Discharge Relationships. Stage-discharge data pairs (N = 103), determined by concurrent flow measurements and stage observations (U.S. Geol. Survey Water-Supply Papers), provide an independent means of assessing changes in channel morphology. Plots of flow stage versus discharge (rating curves) may identify periods of channel aggradation and degradation. Stage depends not only on discharge and bed elevation, however, but integrates the effects of several hydraulic variables over the entire channel boundary, including channel slope, roughness, cross-section area, and flow velocity (Leopold and others, 1964; Hey, 1978; Hey and Thorne, 1986). Nevertheless, long-term changes in the stage of a given discharge can reveal channel scouring or filling trends when used in conjunction with other observations such as cross sections.

Thalweg Elevations. Elevations of the deepest point on each channel cross section (thalweg) were determined from sounding data at the cableway (1931–1984). Observations are primarily from large floods, but several measured low-flows were included to test for channel refilling during low-water periods. Thalweg elevations (relative to the gage datum) augment information from cross-section or stage-discharge relationships by revealing depths of scour and fill.

Bridge Resurveys. Extrapolations of conclusions about channel morphologic change to other locations are complicated by local idiosyncratic responses such as shifting pools, riffles, and bedforms, or knickpoint retreat (Colby, 1964, p. 2; Schumm, 1977). Cross-section surveys were made downstream of the Wheatland gage by the California Highway Department (CalTrans) at the Highway 70 Bridge in 1961 and at the Pleasant Grove Bridge in 1972 (CalTrans unpub. data) (Fig. 3). As a check of the ubiquity of channel cross-section changes at the Wheatland gage, these surveys were repeated by lowering a weight on the end of a tape to measure depth below a specified vertical datum on the bridges.

Channel Shape. Changes in channel shape at the cableway were examined by analysis of hydraulic geometry and width/depth ratios. One hundred field measurements of channel-top width, mean velocity, cross-section area, and discharge were made at the cable site from 1930 to 1984 (hydrographers' notes in the U.S. Geol. Survey Archives). Channel widths, mean velocities, and mean depths (ratio of flow cross-section area to top width) were expressed as power functions of discharge (Leopold and others, 1964), and regression residuals were plotted as time series to reveal temporal changes in channel morphology (Knighton, 1974). Width-depth (W/D) ratios were calculated using top widths and mean depths of flows. This ratio can be calculated for non-equilibrium channels, because it requires no identification of the bankfull channel. The ratio varies with discharge, however, and so variations in shape with discharge were removed by regression and analysis of residuals.

CHANGES IN HYDROLOGIC REGIME AND CHANNEL MORPHOLOGY

Flood Series

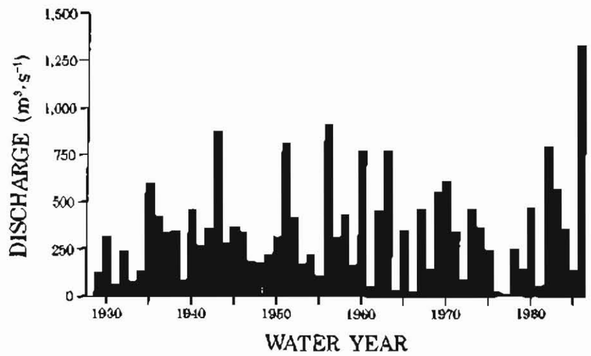


Figure 5. Annual maximum flood series at Wheatland gage, 1929–1986.

The annual maximum flood series is characterized by moderately large floods in water years 1943, 1951, 1956, 1960, 1963, 1982, and 1986 (Fig. 5). Flood-frequency analysis indicates that the December 1955 (water year 1956) flood has a return period of only about 36 yr (Table 2). This computation is based on inclusion of all data from 1929 to 1986 in spite of possible nonstationarity in the hydrologic series. Significant (α = 5%) long-term changes were detected between flood series before and after 1965 at the Wheatland gage. Mitigation of extremely high or low flows by reservoir storage does not explain this difference, because flood variability is greater *after* dam closures and enlargement. The apparent change in flood statistical properties may be due to (1) climate change, (2) land-use changes in the upper basin (for example, logging), (3) reservoir operating policies, or (4) inclusion of the large 1986 flood. No statistically significant difference was found between the pre- and post-1965 data subsets when the 1986 flood was excluded.

No other trends were detected in the flood series. The data show no significant (α = 5%) serial correlation or difference between seasons (Table 2). Floods in many Sierra Nevada channels are derived from two populations: (1) rainfall from November to March and (2) snowmelt from April to July (Water Resources Council, 1981, p. 16), but the Bear River heads at too low an elevation for substantial snowmelt contributions to runoff.

Sediment Textures

Mining sediment in the lower Bear River is much less cohesive or pebbly than older allu-

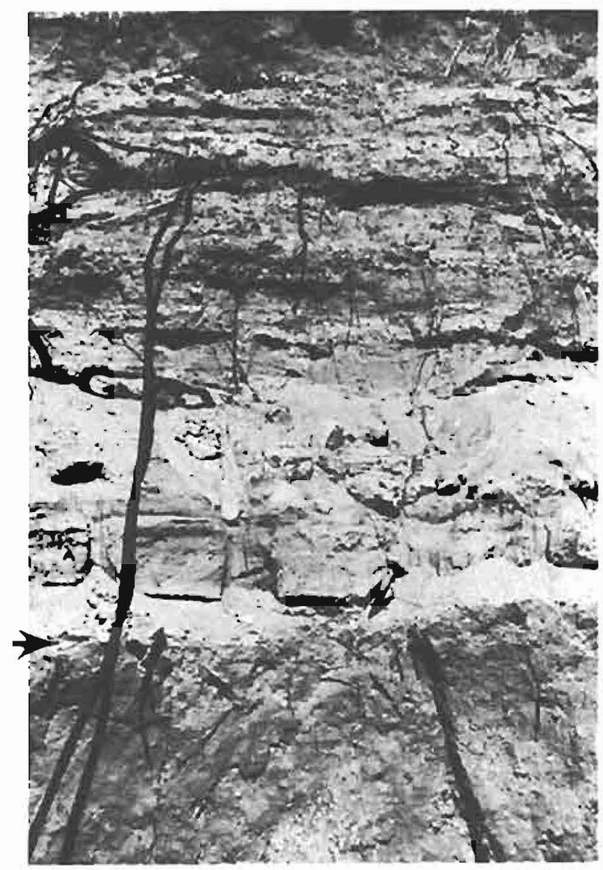


Figure 6. Mining sediment is quartzose, micaceous, and highly stratified. Underlying alluvium (below arrow) is considerably denser.

TABLE 2. ANNUAL MAXIMUM FLOOD SERIES STATISTICS AT WHEATLAND GAGE

Years	Discharge ($m^3 \cdot s^{-1}$)				Nonrandom trends		
					Mann-Whitney*		Spearmans
	N	Max. peak	Mean annual	Std. dev.	Time†	Season†	serial c.c.§
1929–1986	58	1,359	351	270	Sig	NSig	NSig
1929–1965	37	934	346	240	-	NSig	NSig
1966–1986	21	1,359	360	321	-	NSig	NSig

*Difference of means tests (i) before and after 1965, and (ii) Nov.–March versus April–July.
†Sig and NSig represent significant or not significant at the 5% level of confidence, respectively.
§Spearmans rank order serial correlation coefficient.

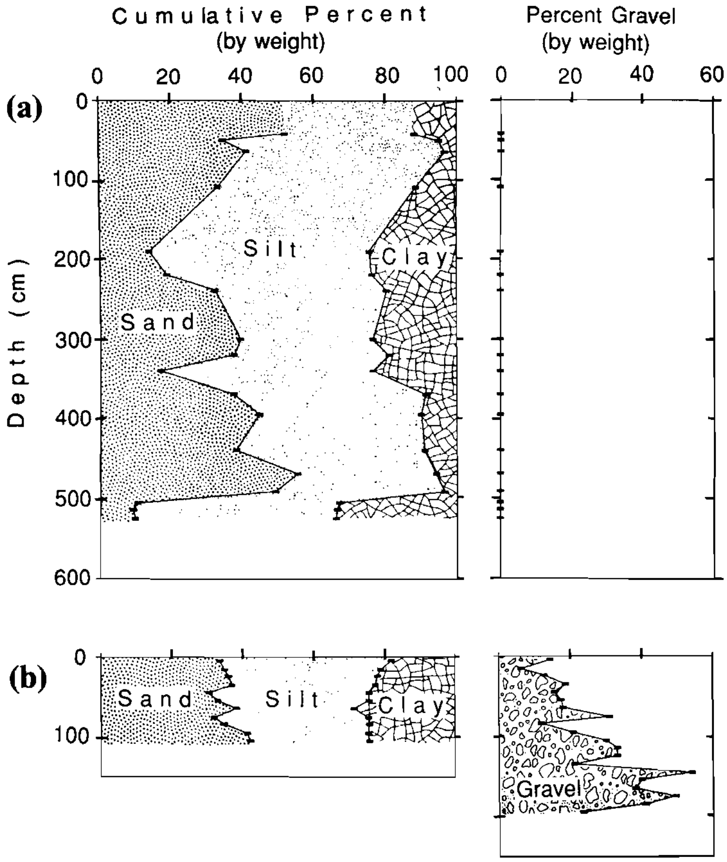


Figure 7. Sediment textures from two vertical sections. (a) Core downstream of Wheatland gage through tailings into A horizon of pre-mining soil at 5-m depth. Mining sediment is siltier and less clayey than pre-mining soil but contains no pebbles or gravel at this site. (b) Samples of pre-mining alluvium from a bank exposure upstream of the Wheatland gage have relatively uniform sand, silt, and clay, and as much as 55% by weight coarser than 2 mm (pebbles and gravel).

vium exposed in channel cuts. Tailings occur in horizontally bedded deposits with abrupt contacts between distinct strata (Fig. 6). The older alluvium is also stratified, but contacts are gradational due to pedogenic translocation of clay and iron. The top 15 tailings samples from a 5.3-m core have less clay than the 3 samples of the underlying soil A horizon (Fig. 7a). The bottom 3 m of the tailings at this site fines upward, but the upper 2 m is relatively sandy. Eleven representative samples of pre-mining alluvium from a bank exposure upstream of the gage site show less textural variation and a more uniformly high clay and sand content than the tailings (Fig. 7b).

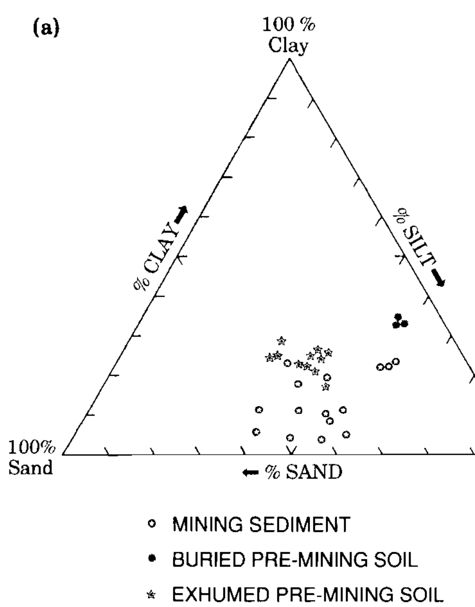
All pre-mining soil samples from the bank exposure contained from 6% to 30% gravel by weight (averaging 18%) in a fining-upward sequence. Tailings samples contained no pebbles or gravel. A similar contrast in pebble abundance can be seen at the gage site, although gravel textures in the pre-mining alluvium are finer. Mining sediment ranges from loam to silt loam to sandy loam (Fig. 8a). In contrast, the pre-mining alluvium samples were all loams with less silt and more clay than the tailings.

Graphic inclusive sorting and skewness of textural frequency distributions are highly variable in the tailings but uniform in the older alluvium (Fig. 8b). Differentiation between the two sediment populations may be possible by textural analysis alone. Tailings samples are relatively well sorted and both positively and negatively skewed. Pre-mining alluvium is poorly sorted and slightly fine skewed due to fine clay in a sandy matrix. This clay combines with iron to cement the sediment into a dense, cohesive mass.

Channel Incision

Cross Sections. Six representative flood cross sections at the cableway, superimposed on the 1984 stratigraphic section, reveal relationships between channel morphology, bed positions, and alluvial units (Fig. 9). Sparse grass dominates the banks with only a few trees and shrubs nearby, and so erosional resistance is imposed primarily by boundary materials. Banks are composite with several strata of varying characteristics, but unlike cantilevered banks described by Thorne and Lewin (1979), the only weak stratum is a sandy loam at the present base of the right bank; all other pre-mining strata are cohesive.

By the time of the earliest available cross-section data in 1931, the channel bed had already eroded through the tailings and was resting on a resistant layer of iron-cemented, pebbly clay, over which it remained for the next 25 yr. This resistant layer was partially penetrated by a moderately large flood in 1943 but was not severely eroded until the moderately large December 1955 flood, the largest flow on



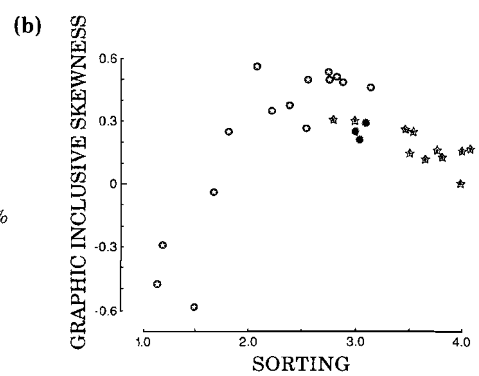


Figure 8. Textures of tailings and pre-mining alluvium after removal of coarse material. (a) Tailings are silty and sandy but low in clay; pre-mining alluvium has 18% to 34% clay. (b) Pre-mining alluvium is poorly sorted and uniformly skewed due to very fine clay cement.

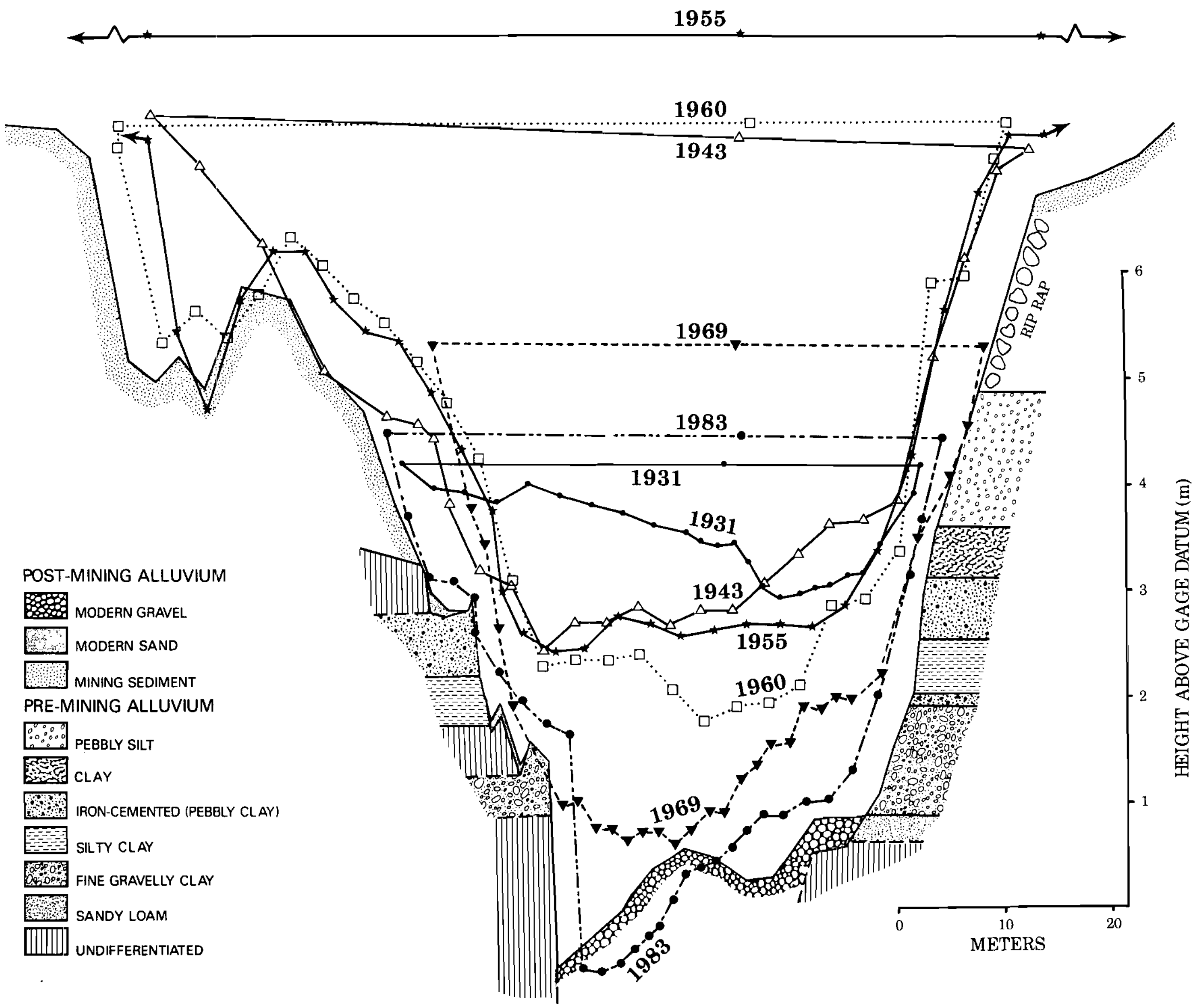


Figure 9. Cross sections and stratigraphy at Wheatland cable (1931–1983), showing more than 3 m of degradation since 1955. Resistant pebbly clay layer is cemented by pedogenic iron. Section surveyed in 1984. High-water conditions with view downstream (vertical exaggeration: 10 ×).

record at that time. During the next 14 yr, the channel down-cut more than 2 m through less-resistant layers of pebbly, silty, and gravelly clay. Incision continued through 1983 as the channel thalweg deepened and migrated toward the left bank, and a gravel bar formed in the channel center.

Stage-Discharge Relationships. The stage-discharge data independently confirm the magnitude of channel incision documented by channel cross sections and elaborate upon the timing of these changes. Discharge explains 55% of the variance in stage at the gage (1930–1985) when stage is expressed as a power function of discharge (Fig. 10a). Stage-discharge regression residuals reveal a distinct pattern of decreasing flow stages through time corresponding to the channel degradation documented by cross sections (Fig. 10b). Residuals for the 3 smallest discharges sampled ($Q < 5\ m^3 \cdot s^{-1}$) are not shown due to poor fit of the power-function model on the low tail of the distribution. The channel bed was relatively stable from 1930 to 1955 as it rested above the resistant stratum, and there is no indication of channel incision following the 1928 closure of Camp Far West dam. Following the 1955 flood, however, the channel began to degrade at an approximately constant rate. Incision persisted into the mid-1970s, when stage-discharge relationships became highly variable. There is no indication that reservoir construction or enlargement in the mid-1960s affected flow stages. Hydrographers' descriptions from 1964 to 1967 (U.S. Geol. Survey Archives) included "mud" along with the gravel, sand, and "hardpan" noted in most years, but dam closure apparently did not cause an acceleration in channel incision.

Thalweg Elevations. The timing of thalweg lowering reinforces conclusions drawn from the cross-section and stage-discharge data that channel incision began during the 1950s, continued until the 1970s, and did not respond to damming in 1928 or the mid-1960s. Thalweg elevations were stable from 1931 to the mid-1950s, incised progressively into the mid-1970s, and fluctuated thereafter (Fig. 11).

Thalweg elevations can be used to distinguish a paved bed from mobile alluvium, because scour by high flows and redeposition during waning flows should result in an inverse correlation between thalweg elevation and discharge. Thalweg elevations were not related to discharge before 1964 (Fig. 12), which suggests a

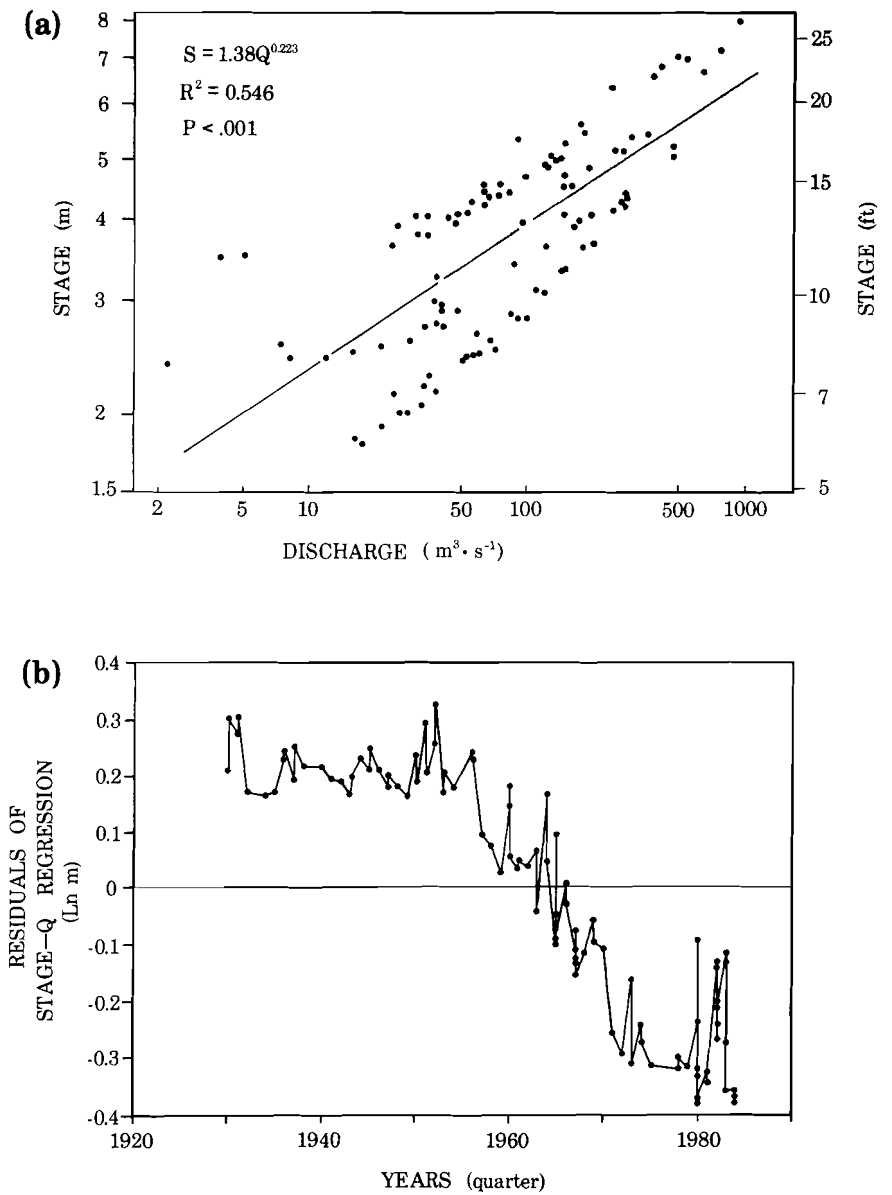


Figure 10. Flow stage versus discharge at the Wheatland gage (1929–1984). (a) Stage (relative to gage datum) as a power function of discharge suggests long-term shifts in rating curves. (b) Time series of residuals (Ln m) shows channel degradation. The bed was stable from 1929 to 1955, degraded steadily from 1955 to 1970+, and high variability began in the 1970s.

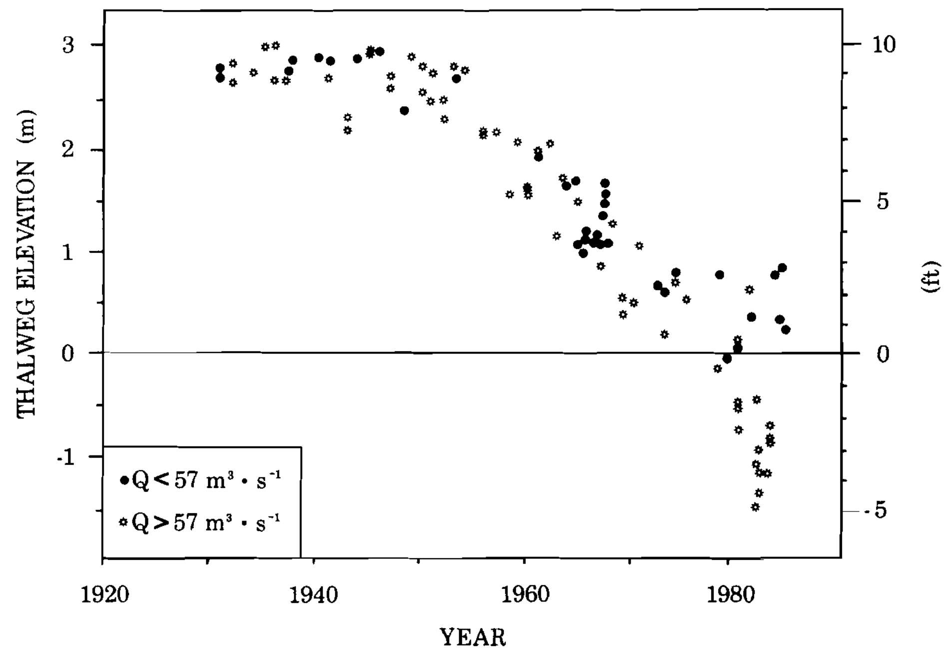


Figure 11. Thalweg elevations above the Wheatland gage datum show that low-flow channel incision began about 1955, was constant to 1970, and then varied. This corresponds closely to the response of the entire channel boundary as inferred from the stage-discharge data.

stable bed with negligible storage and remobilization of sediment at the gage site during this period. A weak insignificant correlation between thalweg elevations and discharge began to emerge in 1965, strengthened in 1972 ($p < .05$), and became highly significant ($p < .001$) after 1978, a period of high variation in bed elevation. These bed-elevation changes with discharge indicate scour during floods, refilling during low flows, and large volumes of bed material passing through the reach. The onset of these bed fluctuations may date the initiation of severe lateral channel migration observed a few kilometers upstream between 1983 and 1989.

Refilling of the channel bed in the 1980s may also signify return to an equilibrium base level. Probing of the pre-mining channel (Fig. 4) suggests that the original bed elevation was not substantially different from the present bed elevation. This preliminary assessment requires further substantiation but may indicate that, by ~110 yr after its avulsion, the channel had returned to its pre-mining base level (Fig. 4).

Bridge Resurveys. Resurveys at two highway bridges downstream of the gage do not conflict with the interpretation that the timing of channel incision at the gage was ubiquitous in lower basin channels. The resurvey at the Highway 70 Bridge (Figs. 13a, 3) documents about 1.5 m of low-flow channel incision during the period from 1961 to 1983, which includes most of the period of rapid incision at the Wheatland gage site. The small degree of down-cutting (about 2/3 the depth of incision at the Wheatland gage during this period) is to be expected, due to the proximity of the bridge site to base-level controls of the Feather River. The resurvey at the Pleasant Grove Bridge indicates channel stability during this later period from 1972 to 1983 which follows the end of rapid incision at the Wheatland gage (Fig. 13b). These surveys are not conclusive, but they support a longitudinally extensive interpretation of the channel-incision data.

Channel Cross-Section Shapes

Channel cross-section shapes support the hypothesis that the channel at the Wheatland gage

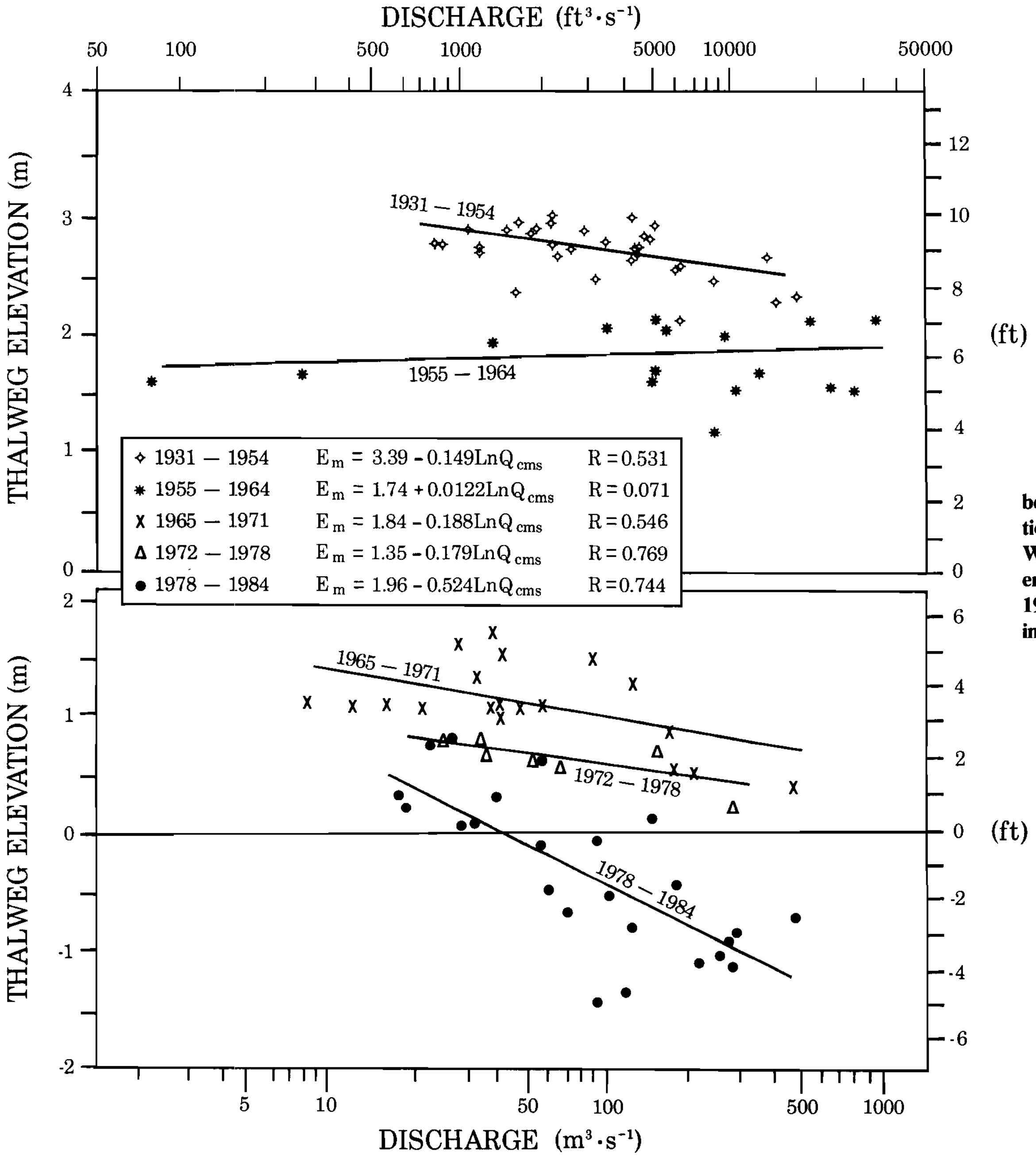


Figure 12. Correlation between thalweg elevations and discharge at the Wheatland gage. Strengthening of correlations after 1978 indicates an increase in mobile bed material.

narrowed as it incised into cohesive sediment and that this narrowing proceeded independently of dam construction or the rapid channel incision that began in 1955.

Hydraulic Geometry. The hydraulic geometry data indicate that long-term changes in channel shape were gradual. Channel top-width, mean depth, and mean velocity are each significantly related to discharge as power functions (Table 3). Time series of the regression residuals (Fig. 14) show that width progressively decreased until stabilizing about 1970. Mean depths increased as widths were decreasing, until about 1970. The sudden decrease in depth may be due to the introduction of bed material in the early 1970s as indicated by cross sections and thalweg elevations. Long-term velocity changes are not pronounced, but an inverse relationship between depth and velocity residuals is strongly expressed for most of the record. This high-frequency inverse relationship between depth and velocity apparently reflects changes in small-scale bed configurations such as pools, dunes, and bars. Low-frequency changes, expressed by decreased width and increased depth over the period of record, reflect fundamental changes in channel cross-section shape.

Width/Depth Ratios. Bear River W/D ratios from 1930 to 1985 decrease at a negative power of discharge (Fig. 15a) due to containment of large floods by steep banks, and spreading out of low-flows over the relatively flat bed, especially before the thalweg trenched in 1978. The two largest outliers to this relationship between W/D and discharge occurred in 1931 in response to relatively small flows (35 and 25 m^3s^{-1}). The next largest outlier represents the 1955 flood (946 m^3s^{-1}), which topped the high terrace and spread out over adjacent fields.

Residuals from the W/D regression decrease progressively through time with no break in slope at 1955 or in the mid-1960s (Fig. 15b). This uniform rate of channel narrowing corroborates the trends revealed by hydraulic geometry and indicates that changes in overall channel shape did not respond directly to rapid incision of the thalweg. These observations display the importance of bank and bed sediment properties to channel morphology at this site. Channel cross-section shapes at the gage changed from wide and shallow to narrow and deep as the channel incised through the tailings into relatively cohesive alluvium below. Channel shape

TABLE 3. HYDRAULIC GEOMETRY POWER FUNCTIONS

Equation	Sample n	r^2	p*
$W = 35.6Q^{0.182}$	100	0.60	<.001
$D = 0.160Q^{0.417}$	100	0.84	<.001
$V = 0.175Q^{0.401}$	100	0.82	<.001

*p = probability of occurring with random data pairs.

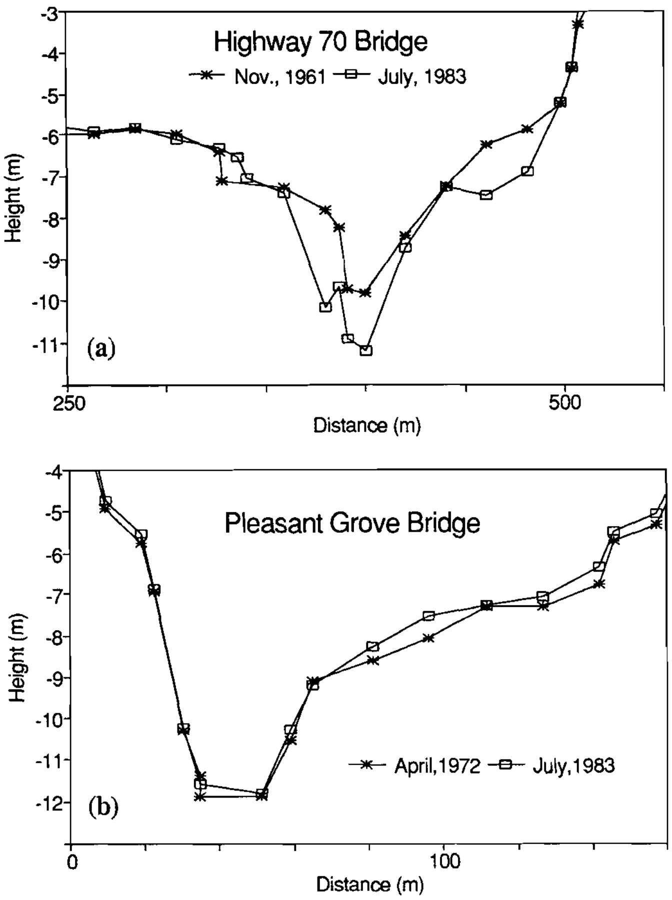


Figure 13. Resurveys of CalTrans highway bridge surveys agree with the observed periods of incision at Wheatland gage upstream. (a) The channel at the Highway 70 Bridge eroded 1.5 m from 1961 to 1983. (b) The Pleasant Grove Bridge section was stable from 1972 to 1983.

was evolving long before 1955, promoting greater depths and shear stresses of a given magnitude flood and encouraging bed erosion and failure of the channel pavement.

DISCUSSION

Controls on Bear River Channel Erosion

The timing of channel incision in the Bear River did not conform to either of the first two hypotheses: dam construction or gradual decline in sediment loads. Sediment detention behind major reservoirs generally lowers sediment concentrations and causes accelerated channel erosion below dams (Lane, 1934; Gessler, 1971) that typically extends hundreds of kilometers below dams, progresses downstream for up to 30 yr, and continues for more than 100 yr (Williams and Wolman, 1984). The construction of two dams in 1928 must have lowered sediment deliveries to the lower Bear River, because sediment production from reworked tailings remains very active upstream of the dams, and storage in steep mid-basin channels has been negligible (James, 1989). Nevertheless, channel cross sections, rating curves, and thalweg elevations from 1930 to 1985 do not show accelerated degradation rates following dam construction in 1928 or in the mid-1960s.

Gilbert (1917) presented an influential model of sediment transport based upon channel-bed incision in the Yuba River at Marysville and the Sacramento River at Sacramento. The Sacramento River channel bed returned to its premining base level by 1930 (Graves and Eliab, 1977). The Yuba River, which also incised down through thick tailings deposits, vigorously eroded prior to 1940 but has been relatively stable since that time (Adler, 1980) or certainly by 1950 (Graves and Eliab, 1977). Although down-valley sediment influxes were halted by dams, the record of incision in the lower Bear River is unlike the Sacramento or lower Yuba Rivers and indicates that the channel bed was maintained at a relatively high elevation until the 1970s (Fig. 10b). This observation calls into question the practice of using changes in channel-bed elevation as a surrogate measure of sediment loads (Gilbert, 1917; Graves and Eliab, 1977; compare with James, 1989).

Alluvial channel morphology responds to changes in water and sediment loads (Leopold and Maddock, 1953; Wolman and Miller, 1960) but is not entirely explained by these changes, because morphologic adjustments often vary between sections with similar loads (Colby, 1964, p. 25). Channel morphology also responds to the strength of bed and bank materials, which is influenced by particle size

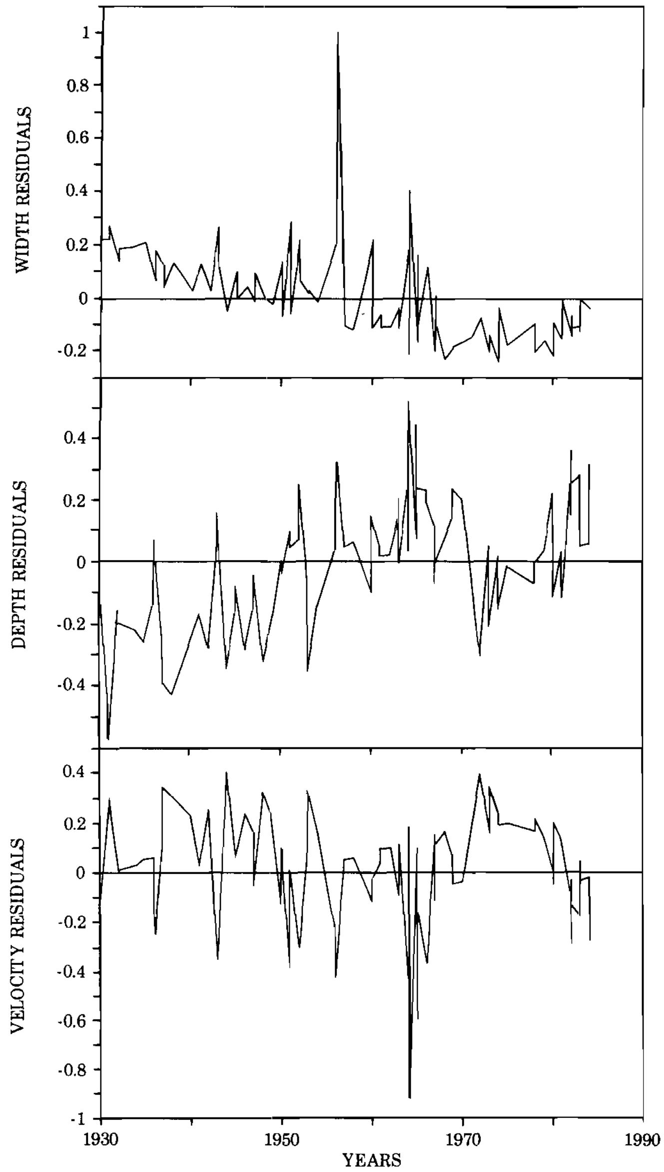


Figure 14. Hydraulic geometry power-function regression residuals at Wheatland gage. Progressive decreases in width and increases in depth reveal long-term channel narrowing and deepening.

loads due to the cessation of mining in 1884. Unlike the Yuba or Sacramento Rivers, however, the lower Bear River channel was stabilized from 1930 to 1955 by a resistant alluvial layer in the bed. Not until that unit was breached did sediment transport capacities have a substantial bearing on erosion rates. Apparently, bed stability was maximized in the pebbly clay unit. The clay cement enhanced resistance to particle entrainment, whereas the pebbles enhanced resistance to abrasion even when the saturated clay matrix became plastic.

Large Floods as Instigators of Episodic Response

The dominant discharge of large alluvial rivers in humid regions is generally a flood of moderate magnitude that occurs, on average, every few years (Wolman and Miller, 1960). This principle has been corroborated for large basins with humid climates by numerous studies (Jahns, 1947; Costa, 1974; Gupta and Fox, 1974; Dury, 1973, 1977; Harvey, 1977; Andrews, 1980) and is an important assumption to many paleohydrological methods based on channel morphology (Knox, 1985). The dominant discharge may be large and infrequent, however, where (1) flows are highly variable so that moderate reconstructive events are infrequent, or (2) channel boundaries are stable and (Schumm, 1960), shape and fabric of coarse bed material (Day, 1981), and cohesion of fine material (Hjülstrom, 1935; Committee on Sedimentation, 1968; Murray, 1977; Grissinger, 1982; Nanson and Hickin, 1986). Characteristics of boundary alluvium should not be inferred from modern sediment loads, however, if the bed and banks are composed of older alluvium that has been altered (for example, by pedogenesis or diagenesis) or retains a memory of different conditions in the past (for example, of former sediment loads or depositional environments).

This study documents the impact of paleostrata on morphologic evolution and incision rates of a channel avulsed by mine tailings. Channel incision in the lower Bear River was encouraged by (1) increased gradients due to channel aggradation and (2) decreased sediment

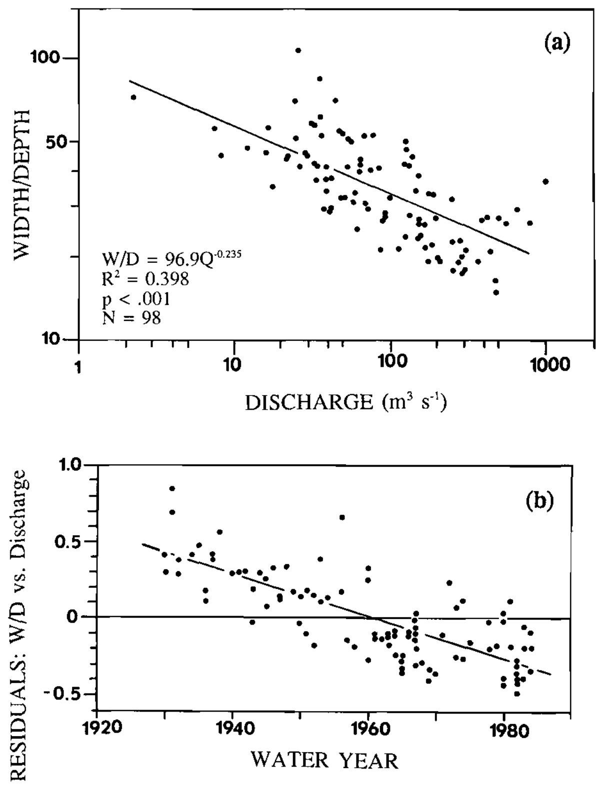


Figure 15. Width/depth ratios at the Wheatland gage. (a) W/D ratios (dimensionless) decrease with discharge due to confinement of large floods by steep banks. (b) Regression residuals (observed Ln W/D minus predicted Ln W/D) indicate that channel deepening and narrowing began prior to 1955 and was gradual in nature.

do not respond to moderate-magnitude floods (Wolman and Miller, 1960, p. 72; Schumm and Lichty, 1963; Stewart and LaMarche, 1967; Burkham, 1972; Baker, 1977; Wolman and Gerson, 1978; Knox, 1979; Lisle, 1981; Nolan and others, 1987). The role of vegetation, coarse sediment textures, and discharge variability on the stability of alluvial-channel boundaries, and thus on the dominant discharge, have been noted by several studies. Much less attention has been paid to the effects of stability imposed by bed and bank cohesion, although intermittent erosion has been associated with highly stratified, cohesive alluvium (Thorne and Lewin, 1979; Begin, 1981; Thorne and Tovey, 1981; Pizzuto, 1984). For example, an 18-cm-thick sand layer in the Rio Puerco, impregnated with about 50% clay in 1961, armored the channel bed and lowered sediment production until it was abruptly eroded by knickpoint retreat (Nordin, 1963).

Most studies of dominant discharge analyze sediment loads directly transported by flows of a specific magnitude (Wolman and Miller, 1960; Pickup and Warner, 1976; Andrews, 1980). This approach can underestimate the importance of large floods that trigger episodes of sediment production or channel change by destabilizing beds and banks and allowing subsequent small flows to transport sediment. Large flood events can instigate major geomorphic episodes, and their importance should not be overlooked simply because smaller events perform most of the work.

Exposure of cohesive materials in the lower Bear channel boundary stabilized the channel by decreasing the ability of a given shear stress to initiate erosion. The moderately large 1955 flood instigated an episode of sediment production and morphological change. It did not move extremely large sediment volumes or directly cause large changes to channel morphology, but it disrupted stability, allowing subsequent incision and high sediment production rates by relatively small flows that had not previously been competent to erode the channel margin.

Secular Changes in Channel Stability

Geomorphic stability can be disrupted not only by increasing applied stresses, but also by decreasing resistant forces (Schumm, 1973, 1977). Prior to the 1955 flood, progressive morphological changes by moderate-magnitude floods were creating conditions of inevitable failure of the channel pavement at the gage site. Channel-bed stability decreased through time as (1) channel incision thinned and weakened the paved layer and (2) applied bed shear stresses increased with channel narrowing and deepening of flows for a given discharge. Failure of the paved layer would ultimately have occurred even without a large flood event, due to the decreased differential between resistant and applied forces (Fig. 16). This model of channel response implies that progressive morphological changes may trigger sudden episodic changes. The difficulty lies in detecting such progressive changes, determining the destabilizing effects of these changes, and anticipating sudden extensive adjustments that they may portend.

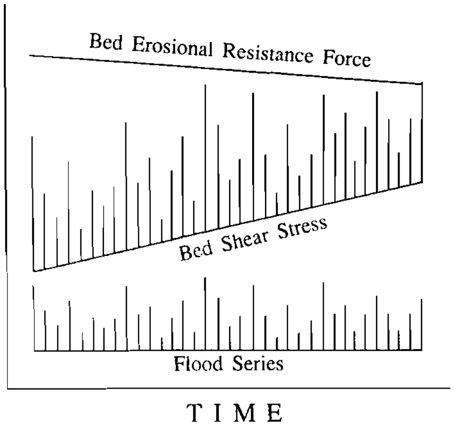


Figure 16. Conceptual model of increasing competence of small floods to substantially erode the Bear River channel bed in 1955. The strength of bed erosion resistance diminished only slightly as the pavement thinned. Destabilization was due primarily to progressive increases in bed shear stresses, due to flow deepening as the channel narrowed, added onto randomly varying flood-induced stresses. (Adapted from Schumm, 1973.)

CONCLUSIONS

Data from the Wheatland stream-flow gage on the lower Bear River document changes in channel cross sections, stage-discharge relationships, thalweg elevations, hydraulic geometry, and W/D ratios in response to floods and to stratified channel-boundary materials. Channel avulsion in the 1870s superposed the channel at a higher position on the pre-mining topographic surface. Erosional resistance of the older alluvium prevented a rapid return to pre-mining base levels. From 1930 to 1955, the channel was relatively stable. Channel incision began in 1955 and continued into the mid-1970s when bed elevations began to fluctuate, presumably in response to the introduction of gravel. Incision took place much later than in the Sacramento or Yuba Rivers, which also aggraded severely during the mining period but had stabilized by 1950.

Channel-bed incision rates in the lower Bear Basin were controlled largely by boundary materials. A stratum of pebbly alluvium cemented by pedogenic clay and iron prevented substantial incision for at least 25 yr. In 1955, a moderately large flood penetrated this stratum and instigated a period of steady bed incision that was maintained for about 20 yr by smaller floods. Decreased sediment influxes to the lower Bear caused by upstream dam closures in 1928 and in the mid-1960s had little apparent effect on the channel. Two channel surveys at bridges downstream of the gage support the interpretation that timing of channel incision is broadly synchronous throughout the lower Bear River.

The 1970s and early 1980s were marked by high variability in channel-bed elevations due to scour by floods and refilling by low flows. These bed fluctuations signal increased sediment loads that apparently coincide with the initiation of severe bank erosion upstream.

Hydraulic geometry and W/D ratios demonstrate progressive changes in channel shape since 1930 that were independent of the onset of rapid channel incision in 1955 or dam construction in the 1960s. Narrowing and deepening of cross sections began prior to penetration of the armored layer and continued at a constant rate as the channel incised through the tailings into the underlying pebbly clay alluvium. The progressive evolution of channel cross-section shapes ensured the ultimate failure of the resistant stratum, which resulted in a step-function change in channel-incision rates. Slow, progressive evolution of channel cross-section form thus led to an episodic response in erosion rates and base-level change.

ACKNOWLEDGMENTS

I am indebted to James C. Knox for use of laboratory equipment, Laurence B. James for field assistance, Francis J. Magilligan for timely statistical advice, Dave and Pat Beach for field accommodations, and CalTrans staff for bridge data. I am particularly thankful to the U.S. Geological Survey at Sacramento, including Thomas A. Hankins, Verrie F. Pierce, and Robert G. Simpson, for access to archival data and assistance. This paper benefitted considerably from reviews of early versions by Victor R. Baker, Harvey M. Kelsey, James C. Knox, Robert H. Meade, John C. Pitlick, and Garnett P. Williams. Although I take full responsibility for the interpretations and conclusions, which are my own, I thank these individuals for their invaluable comments.

REFERENCES CITED

Adler, L. L., 1980, Adjustment of the Yuba River, California, to the influx of hydraulic mining debris, 1849-1979 [M.S. thesis]: Los Angeles, California, Geography Department, University of California, Los Angeles.

Andrews, E. D., 1980, Effective and bankfull discharges of streams in the Yampa River Basin, Colorado and Wyoming: Journal of Hydrology, v. 46, p. 311-330.

Axelrod, D. I., 1962, Post-Pliocene uplift of the Sierra Nevada, California: Geological Society of America Bulletin, v. 73, p. 183-198.

Baker, V. R., 1977, Stream channel response to floods with examples from central Texas: Geological Society of America Bulletin, v. 88, p. 1057-1071.

Begin, Z. B., 1981, Stream curvature and bank erosion: A model based on the momentum equation: Journal of Geology, v. 89, p. 497-504.

Brush, L. M., Jr., 1961, Drainage basins, channels and flow characteristics of selected streams in central Pennsylvania: U.S. Geological Survey Professional Paper 282-F, p. 145-181.

Bryan, K., 1923, Geology and ground-water resources of Sacramento Valley, California: U.S. Geological Survey Water-Supply Paper 495.

Burkham, D. E., 1972, Channel changes of the Gila River in Safford Valley, Arizona 1846-1970: U.S. Geological Survey Professional Paper 655-G, Gila River Phreatophyte Project, 24 p.

Burnett, J. L., and Jennings, C. W., 1962, Geologic map of California, Chico sheet: California Division of Mines and Geology, scale 1:250,000.

Busacca, A. J., 1982, Geologic history and soil development, northeastern Sacramento Valley, California [Ph.D. dissert.]: Davis, California, University of California, Davis.

California Department of Water Resources, 1988, Dams within jurisdiction of the State of California: California Department of Water Resources Bulletin 17-88.

California Water Resources Board, 1955, Placer County investigation: California State Water Resources Board Bulletin 10a.

Christiansen, M. N., 1966, Late Cenozoic crustal movements in the Sierra Nevada of California: Geological Society of America Bulletin, v. 77, p. 162-181.

Clark, L. D., 1976, Stratigraphy of the north half of the western Sierra Nevada metamorphic belt, California: U.S. Geological Survey Professional Paper 923.

Colby, B. R., 1964, Scour and fill in sand-bed streams: U.S. Geological Survey Professional Paper 462-D.

Committee on Sedimentation, 1968, Erosion of cohesive sediments, Report of the Task Committee on Erosion of Cohesive Materials: American Society of Civil Engineers, Journal of the Hydraulics Division, v. 94, HY4, p. 1017-1049.

Costa, J. E., 1974, Response and recovery of a Piedmont watershed from tropical storm Agnes, June 1972: Water Resources Research, v. 10, p. 106-112.

——1983, Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range: Geological Society of America Bulletin, v. 94, p. 986-1004.

Davis, G. H., and Olmsted, F. H., 1952, Geologic features and ground-water storage capacity of the Sutter-Yuba area, California: Sutter-Yuba Counties investigation, California Department of Water Resources Bulletin no. 6, Appendix B.

Davis, S. N., and Hall, F. R., 1959, Water quality of eastern Stanislaus and northern Merced Counties, California: Stanford University Publications, Geological Sciences 6.

Day, P. R., 1965, Particle fractionation and particle-size analysis, *in* Black, C. A., ed., Methods of soil analysis: Madison, Wisconsin, American Society of Agronomy.

Day, T. J., 1981, An experimental study of armouring and hydraulic properties of coarse bed material channels, *in* Erosion and sediment transport in Pacific rim steeplands: International Association of Hydrological Sciences Publication No. 132.

Durrell, C., 1959, Tertiary stratigraphy of the Blairsden quadrangle, Plumas County, California: California University, Department of Geological Sciences Bulletin, v. 34, p. 161-192.

Dury, G. H., 1973, Magnitude-frequency analysis and channel morphometry, *in* Morisawa, M., ed., Fluvial geomorphology: Binghampton, New York, State University of New York, p. 91-121.

——1977, Peak flows, low flows, and aspects of geomorphic dominance, *in* Gregory, K. J., ed., River channel changes: New York, Wiley & Sons, p. 61-74.

Fisk, H. N., 1944, Geological investigation of the alluvial valley of the lower Mississippi River: Vicksburg, Mississippi, Mississippi River Commission, 78 p.

Folk, R. L., and Ward, W. C., 1957, Brazos River bar: A study in the significance of grain size parameters: Journal of Sedimentary Petrology, v. 27, p. 2-26.

Gessler, J., 1971, Aggradation and degradation, *in* Shen, H., ed., River mechanics: Fort Collins, Colorado, Water Resources Publication, p. 8.1-8.24.

Gilbert, G. K., 1917, Hydraulic-mining debris in the Sierra Nevada: U.S. Geological Survey Professional Paper 105, 154 p.

Graves, W., and Eliab, P., 1977, Sediment study: Alternative delta water facilities; Peripheral Canal plan, California: California Department of Water Resources, Central Division, 120 p.

Grissinger, E. H., 1982, Bank erosion of cohesive materials, *in* Hey, R. D., Bathurst, J. C., and Thorne, C. R., eds., Gravel bed rivers: New York, J. Wiley & Sons, p. 273-288.

Gupta, A. H., and Fox, H. L., 1974, Effects of high magnitude floods on channel form: A case study from the Maryland Piedmont: Water Resources Research, v. 10, p. 499-509.

Hall, W. H., 1880, Report of the State Engineer to the Legislature of California, Session of 1880: Sacramento, California, California Printing Office.

Harvey, A. M., 1977, Event frequency in sediment production and channel change, *in* Gregory, K., ed., River channel changes: New York, Wiley Interscience Publications, p. 301-315.

Hey, R. D., 1978, Determinate hydraulic geometry of river channels: American Society of Civil Engineers, Journal of the Hydraulics Division, v. 105(HY4), p. 869-885.

Hey, R. D., and Thorne, C. R., 1986, Stable channels with mobile gravel beds: Journal of Hydraulic Engineering, v. 112(8), p. 671-689.

Hjülstrom, F., 1935, Studies of the morphological activity of rivers as illustrated by the River Fyris: Bulletin of the Geological Institutions of Uppsala, v. 25, p. 293-305.

Jahns, R. H., 1947, Geologic features of the Connecticut Valley, Massachusetts, as related to recent floods: U.S. Geological Survey Water-Supply Paper 996.

James, L. A., 1988, Historical transport and storage of hydraulic mining sediment in the Bear River, California: [Ph.D. dissert.]: Madison, Wisconsin, Departments of Geology and Geography, University of Wisconsin, Madison.

——1989, Sustained storage and transport of hydraulic gold mining sediment in the Bear River, California: Annals of the Association of American Geographers, v. 79, p. 570-592.

——1990, Late Quaternary stream capture, northern Sierra Nevada: Association of American Geographers, Annual Meeting, Program and Abstracts.

Kelley, R., 1959, Gold vs. grain: The hydraulic mining controversy in California's Sacramento Valley: Glendale, California, Arthur Clark Company.

Keyes vs. Little York *et al.*, 1878, Keyes vs. Little York Gold Washing Company *et al.*: California District Court, 10th District, Sutter County.

Knighton, A. D., 1974, Variation in width-discharge relation and some implications for hydraulic geometry: Geological Society of America Bulletin, v. 85, p. 1069-1076.

Knox, J. C., 1979, Geomorphic evidence of frequent and extreme floods, *in* Henry, W. P., ed., Improved hydrologic forecasting, why and how: New York, American Society of Civil Engineers, p. 220-238.

——1985, Responses of floods to Holocene climatic change in the upper Mississippi Valley: Quaternary Research, v. 23, p. 287-300.

Kochel, R. C., and Baker, V. R., 1982, Paleoflood hydrology: Science, v. 215, p. 353-361.

Lane, E. W., 1934, Retrogression of levels in riverbeds below dams: Engineering News Record, v. 112, p. 836-838.

——1937, Stable channels in erodible material: American Society of Civil Engineers Transactions, v. 63, p. 123-142.

Lane, E. W., and Borland, W. M., 1954, River-bed scour during floods: American Society of Civil Engineers Transactions, v. 119, p. 1069-1079.

Leopold, L. B., and Maddock, T., Jr., 1953, The hydraulic geometry of stream channels and some physiographic implications: U.S. Geological Survey Professional Paper 252.

Leopold, L. B., Wolman, M. G., and Miller, J., 1964, Fluvial geomorphology: San Francisco, W. H. Freeman and Company, 522 p.

Lindgren, W., 1911, Tertiary gravels of the Sierra Nevada of California: U.S. Geological Survey Professional Paper 73, 226 p.

Lisle, T. E., 1981, The recovery of aggraded stream channels at gauging stations in northern California and southern Oregon, *in* Erosion and sediment transport in Pacific rim steeplands: International Association of Hydrological Sciences, Publication No. 132, p. 189-209.

Lorens, P. J., 1978, Evaluation of ground water resources: Sacramento Valley: California Department of Water Resources Bulletin 118-6.

Mackin, J. H., 1948, Concept of the graded river: Geological Society of America Bulletin, v. 59, p. 463-512.

Marchand, D. E., and Allwardt, A., 1981, Late Cenozoic stratigraphic units, northeastern San Joaquin Valley, California: U.S. Geological Survey Bulletin 1470, 70 p.

Mendell, Col. G. H., 1880, Mining debris in Sacramento River: House Document 69, 46th Congress, 2nd Session, 11 p.

——1881, Protection of the navigable waters of California from injury from the debris of mines: House Document 76, 46th Congress, 3rd Session.

——1882, Report upon a project to protect the navigable waters of California from the effects of hydraulic mining: House Document 98, 47th Congress, 1st Session, 110 p.

Murray, W. A., 1977, Erodibility of coarse sand-clayey silt mixtures: American Society of Civil Engineers, Journal of the Hydraulics Division, v. 103(HY10), p. 1222-1227.

Nanson, G. C., and Hickin, E. J., 1986, A statistical analysis of bank erosion and channel migration in western Canada: Geological Society of America Bulletin, v. 97, p. 497-504.

Nolan, K. M., Lisle, T. E., and Kelsey, H. M., 1987, Bankfull discharge and sediment transport in northwestern California, *in* Bechta, R. L., ed., Erosion and sedimentation in the Pacific rim: International Association of Hydrological Sciences Publication No. 165, p. 439-450.

Nordin, C. F., Jr., 1963, A preliminary study of sediment transport parameters, Rio Puerco near Bernardo, New Mexico: U.S. Geological Survey Professional Paper 462-C, p. 1-21.

Pickup, G., and Warner, R. F., 1976, Effects of hydrologic regime on magnitude and frequency of dominant discharge: Journal of Hydrology, v. 29, p. 51-75.

Pilon, P. J., Condie, R., and Harvey, K. D., 1985, Consolidated frequency analysis package, CFA, User manual for version 1 - DEC PRO series: Ottawa, Ontario, Water Resources Branch, Inland Waters Directorate, Environment Canada.

Piper, A. M., Gale, H. S., Thomas, H. E., and Robinson, T., 1939, Geology and ground water hydrology of the Mokelumne area, California: U.S. Geological Survey Water-Supply Paper 780.

Pizzuto, J. E., 1984, Bank erodibility of sand-bed streams: Earth Surface Processes and Landforms, v. 9, p. 113-124.

Rantz, S. E., and others, 1982, Measurement and computation of streamflow: Volume I. Measurement of stage and discharge: U.S. Geological Survey Water-Supply Paper 2175.

Richards, K. S., 1976, Complex width-discharge relations in natural river sections: Geological Society of America Bulletin, v. 87, p. 199-206.

Schumm, S. A., 1960, The effect of sediment type on the shape and stratification of some modern fluvial deposits: American Journal of Science, v. 258, p. 177-184.

——1963, Sinuosity of alluvial rivers on the Great Plains: Geological Society of America Bulletin, v. 74, p. 1089-1100.

——1973, Geomorphic thresholds and complex response of drainage systems, *in* Morisawa, M., ed., Fluvial geomorphology: Binghampton, New York, State University of New York, p. 299-310.

——1977, The fluvial system: New York, J. Wiley & Sons, 338 p.

Schumm, S. A., and Lichty, R. W., 1963, Channel widening and floodplain construction along Cimarron River in southwestern Kansas: U.S. Geological Survey Professional Paper 352-D, p. 71-88.

Schweickert, R., and Cowan, D., 1975, Early Mesozoic tectonic evolution of the western Sierra Nevada, California: Geological Society of America Bulletin, v. 86, p. 1329-1336.

Stewart, J. H., and LaMarche, V. C., 1967, Erosion and deposition in the flood of December 1964 on Coffee Creek, Trinity County, California: U.S. Geological Survey Professional Paper 422-K.

Thorne, C. R., and Lewin, J., 1979, Bank processes, bed material movement and planform development in a meandering river, *in* Rhodes, D. D., and Williams, G. P., eds., Adjustments of the fluvial system: Dubuque, Iowa, Kendal Hunt, p. 117-137.

Thorne, C. R., and Tovey, N. K., 1981, Stability of composite riverbank: Earth Surface Processes and Landforms, v. 6, p. 469-484.

Von Geldern, O., 1891, Report of Otto Von Geldern, Assistant Engineer, on an examination of damaged lands in the Sacramento Valley: House Document 1, Part 2, 52nd Congress, 1st Session: Appendix VV, p. 3088-3113.

Water Resources Council, 1981, Guidelines for determining flood flow frequency: Water Resources Council Bulletin 17B, Hydrology Committee.

Wentworth, C. K., 1922, A scale of grade and class terms for clastic sediments: Journal of Geology, v. 30, p. 377-392.

Whitney, J. D., 1880, The auriferous gravels of the Sierra Nevada: Harvard University Museum of Comparative Zoology Memoir 6, 569 p.

Williams, G. P., 1978, Hydraulic geometry of river cross sections—Theory of minimum variance: U.S. Geological Survey Professional Paper 1929, 47 p.

Williams, G. P., and Wolman, M. G., 1984, Downstream effects of dams on alluvial rivers: U.S. Geological Survey Professional Paper 1286.

Williams, J. R., 1982, Mathematical modeling of watershed sediment yield, *in* Singh, V. P., ed., Components of hydrologic cycle: International Symposium on Runoff Modelling; May, 1981, Mississippi State University, Proceedings, p. 590.

Wolman, M. G., 1959, Factors influencing erosion of a cohesive river bank: American Journal of Science, v. 257, p. 204-216.

Wolman, M. G., and Gerson, R., 1978, Relative scales of time and effectiveness of climate in watershed geomorphology: Earth Surface Processes, v. 3, p. 189-208.

Wolman, M. G., and Miller, J. P., 1960, Magnitude and frequency of forces in geomorphic processes: Journal of Geology, v. 68, p. 54-74.

Manuscript Received by the Society April 11, 1990
Revised Manuscript Received September 11, 1990
Manuscript Accepted September 18, 1990