Reprinted from

GEOMORPHOLOGY

Geomorphology 14 (1995) 199–210

Coarse-sediment distribution as evidence of an elevation limit for flash flooding, Bear Creek, Colorado

Michael M. Grimm [1,a], Ellen E. Wohl [a], Robert D. Jarrett [b]

[a] *Department of Earth Resources, Colorado State University, Ft. Collins, CO 80523, USA*

[b] *U.S. Geological Survey, Water Resources Division, Box 25046, MS 418, Denver Federal Center, Denver, CO 80225, USA*

ELSEVIER

ELSEVIER

Geomorphology 14 (1995) 199–210

GEOMORPHOLOGY

Coarse-sediment distribution as evidence of an elevation limit for flash flooding, Bear Creek, Colorado

Michael M. Grimm [1,a], Ellen E. Wohl [a], Robert D. Jarrett [b]

[a] Department of Earth Resources, Colorado State University, Ft. Collins, CO 80523, USA

[b] U.S. Geological Survey, Water Resources Division, Box 25046, MS 418, Denver Federal Center, Denver, CO 80225, USA

Received 10 October 1994; revised 18 February 1995; accepted 3 March 1995

Abstract

Bear Creek is a tributary of the South Platte River in central Colorado. The stream flows east from an elevation of 4348 m at the Continental Divide to the mountain front at 1670 m. It thus encompasses the 2300 m elevation limit for substantial rainfall flooding in the Colorado Front Range proposed by Jarrett. Maximum paleoflood discharges estimated from flood deposits at four sites along Bear Creek demonstrate a consistent decrease in unit discharge with increasing elevation and support the hypothesis of an upper elevation limit for rainfall floods. The unit discharge values were used to explain coarse-sediment distribution along Bear Creek. Measurements of coarse-grained channel sediment at 19 sites along the creek indicate a decrease in particle size in flood deposits with increasing elevation, as well as a decrease in the size of clasts introduced to the main channel along tributaries. These changes in grain size are hypothesized to reflect changes in the competence of channel transport as a result of snowmelt-dominated versus rainfall-dominated discharge regimes above and below 2100 m elevation. Calculations of flow competence versus entrainment thresholds for the deposits may support this interpretation. One of the geomorphic implications of the elevation limit on flash flooding is a reversal of the usual downstream-fining trend in coarse channel sediments.

1. Introduction

The effect of floods on channel morphology and sediment transport has been a subject of continuing interest since the pioneering studies of Wolman and Miller (1960) and Wolman and Gerson (1978). These studies noted the importance of floods on channel morphology in basins with high flow variability or resistant channel boundaries. Subsequent studies have attempted to quantify these relations (Gupta, 1975, 1988; Baker, 1984; Baker and Pickup, 1987; Miller, 1990; Wohl, 1992). This paper examines the effect of spatially disjunct rainfall-induced flash flooding on coarse-sediment deposition in a high-gradient channel of the Colorado Rocky Mountains.

The Colorado Rocky Mountains are representative of many mountainous areas of the world in that flood magnitude–frequency relations are difficult to estimate as a result of limited knowledge of flood hydrometeorology. Rocky Mountain flood hydrometeorology is poorly understood because of the complex meteorologic conditions in mountainous areas, the sparsity and short duration of precipitation and discharge records, and the mixed population of floods resulting from either rain-on-snow, snowmelt, or rainfall (Jarrett, 1987, 1990).

Of the three types of floods in the Colorado Rockies, intense rainfall floods are the most catastrophic. They

[1] During this research also employed at the U.S. Geological Survey. Present address: Floodplain Administrator, City of Ft. Collins, P.O. Box 580, Ft. Collins, CO 80522, USA.

SSDI 0169-555X(95)00037-2

are characterized by a rapid increase in river stage, over a few minutes to a few hours, followed by a rapid decline to preflood stages (Follansbee and Sawyer, 1948; McCain and Ebling, 1979). Overland and stream flow velocities are swift, enhancing erosion on hillslopes and in channels (Shroba et al., 1979; Jarrett and Costa, 1988) and creating boulder bars. Rain-on-snow floods seldom occur and are assumed to have a negligible influence on coarse sediment distribution at the study sites. Snowmelt-runoff floods have very broad hydrograph peaks and smaller magnitudes than rainfall floods. We assume that snowmelt flood peaks are less competent than rainfall floods to transport the coarse portion of available sediment at the study sites and that snowmelt floods are less likely to create boulder bars. This assumption is supported by observations of newly created boulder bars following rainfall floods (Costa, 1978, 1983; Shroba et al., 1979; Jarrett, 1987, 1990; Jarrett and Costa, 1988; Jarrett and Waythomas, 1995) and by the lack of such deposits following snowmelt floods.

Using U.S. Geological Survey gaged discharge records for the Rocky Mountains, Jarrett (1993) identified variable elevation limits to rainfall-produced flash flooding. The limit varies with latitude and distance from moisture sources in the Gulf of Mexico. It ranges from about 2350 m in New Mexico to 1650 m in Montana, and is 2300 m in Colorado (Jarrett and Costa, 1988; Jarrett, 1987, 1990). Flash flooding rarely occurs above this elevation limit, but the intense rainfalls below the limit produce frequent floods.

The disjunct spatial (downstream) distribution of rainfall-induced flash floods along channels in the Rocky Mountains facilitates a study of the geomorphic role of these floods by providing contiguous channel reaches affected and unaffected by rainfall floods. In this paper we characterize downstream trends in coarse sediment deposition along Bear Creek as an illustration of how flash floods affect coarse-sediment deposits in a mountain channel. The grain-size characteristics and spatial distribution of coarse sediments above and below the flash-flood elevation limit are distinctly different.

We begin by estimating maximum flood discharge at four sites along Bear Creek using paleostage indicators. Understanding of flood magnitude–frequency relations can be enhanced by using paleohydrologic data to reconstruct the characteristics of floods not included in systematic data. Paleoflood data can indicate probable upper limits for the largest floods that have occurred in a basin (Costa, 1983). Both the occurrence and absence of paleohydrologic indicators within a given channel reach indicate the spatial distribution of flooding in a basin. In channels of the Colorado Rocky Mountains, paleoflood indicators include flood boulder bars, alluvial fans, impact scars on trees, and slackwater deposits (Jarrett, 1990). These indicators are found between elevations of approximately 2300 m and 1670 m (the base of the mountain range).

We used peak flood discharges estimated from paleostage indicators at Bear Creek to explain coarse-sediment distribution along the creek. We hypothesize that values of D_{50} and D_{max} are inversely proportional to elevation within the drainage basin as a result of decreasing peak unit discharge with increasing elevation. The relation between unit discharge and elevation may be explained in terms of an upper elevation limit for rainfall-floods.

Based on the abundance of cobble- and boulder-sized clasts at all study sites, we assume that the availability of coarse sediment to the channels is not a limiting factor in the Bear Creek basin. Under that assumption, changes in sediment-size distribution between similarly sized tributary drainages can qualitatively indicate changes in flood peak magnitude and in stream power per unit area as a result of runoff regime (Grimm, 1993). For example, if the proposed 2300 m elevation limit for rainfall-dominated floods is correct, tributaries downstream from the limit should show substantially coarser particle size distributions than similarly sized tributaries above 2300 m. In addition, particle-size distributions upstream and downstream from tributary junctions below 2300 m should be substantially different.

Because these small tributary basins are entirely below 2300 m, they are often the source areas of large floods (Diebold, 1939; Follansbee and Sawyer, 1948). The larger peak unit discharge for these tributaries implies greater sediment-transport competence than most flows in the main channel. When a flood peak from a tributary enters the main channel, much of the coarse sediment is deposited as a bar, analogous to the debris fans described for the Colorado River in the Grand Canyon (Webb et al., 1989). In contrast, we hypothesize that a tributary at or upstream from 2300 m with approximately the same drainage area, but that is snowmelt dominated, should show similar particle-size distributions upstream and downstream from the tributary junction because snowmelt occurs more uniformly than intense rainfall–runoff floods and is unlikely to produce tributary unit peak discharges substantially larger than those in the main channel.

2. Field area

Bear Creek originates at the Continental Divide west of Denver, Colorado and flows 72 km to its confluence with the South Platte River (Fig. 1). The majority of the 680 km^2 basin lies in the Southern Rocky Mountains physiographic province, but the lower basin is in the Colorado Piedmont (Diebold, 1939). Elevation ranges from 4,348 m to 1,615 m. The mountainous portion of the basin is underlain by a complex assortment of Precambrian granites, schist, and gneiss (Smith, 1964; Scott, 1972; Sheridan et al., 1972; Sheridan and Marsh, 1976). The mountain–piedmont border is marked by N–S trending hogbacks formed of eastward-dipping sandstones, shales, and limestones of Carboniferous to Tertiary age.

Many faults and shear zones complicate the bedrock geology of the mountainous area, and stream courses often coincide with mapped faults. The canyon section of Bear Creek varies from a wide valley localized by faulting to a steep-sided, narrow canyon cut into granites and pegmatite. Channel gradient, wich ranges from 0.095 to 0.019, is related to bedrock resistance and valley width. The creek has built longitudinally discontinuous floodplains up to 150 m across in the wider reaches, and these reaches are characterized by fine sediment deposition overlying strath terraces. The underlying lithology also affects coarse sediment availability. Granitic terrain generally yields larger and more abundant boulders to erosive processes than does metamorphic terrain partly because the granite is more massive and the metamorphic rocks tend to be strongly foliated and closely jointed. Consequently, river reaches in granitic terrain may contain more boulders and be hydraulically rougher than reaches flowing across metamorphic rocks.

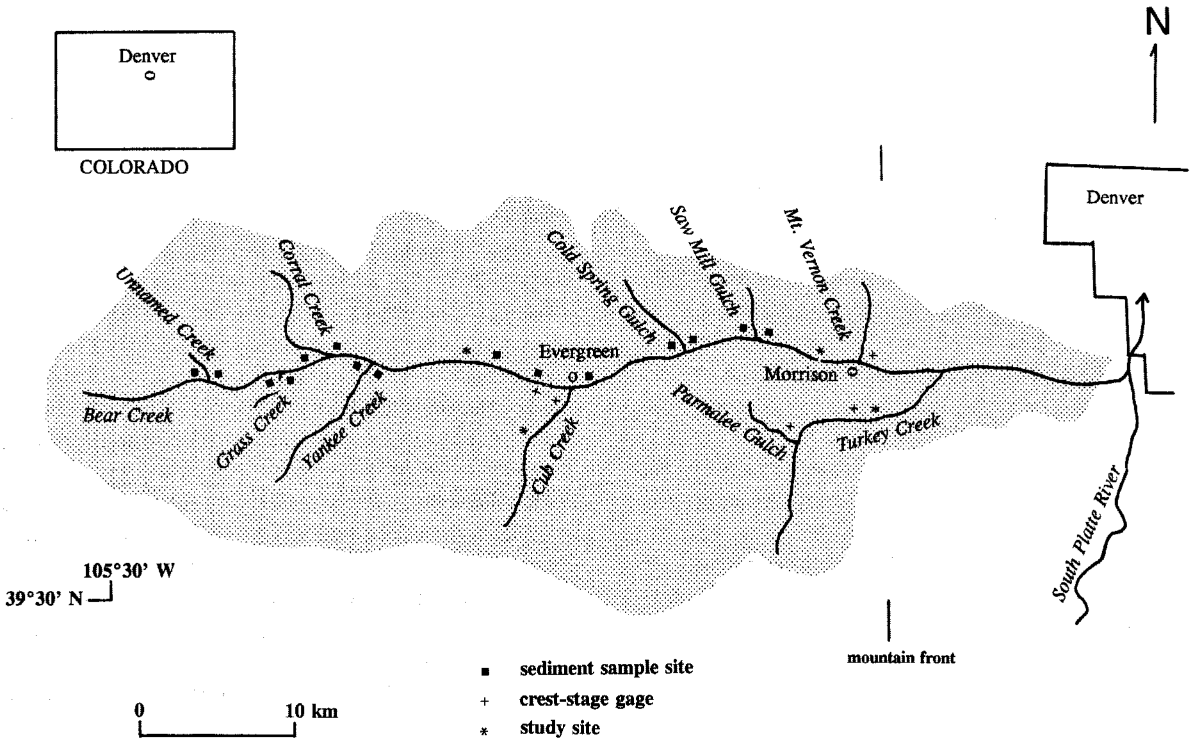

Fig. 1. Location map for the Bear Creek basin, Colorado. Drainage area of Bear Creek is shaded. The towns of Morrison and Evergreen are indicated by circles.

The soils of the upper Bear Creek basin are shallow and patchy, with low infiltration rates, rapid runoff potential, and high susceptibility to erosion (Diebold, 1939; Shroba et al., 1979). Sheetflooding, rill erosion, and gullying provide a continuous supply of colluvium to the channels. Open montane forest of ponderosa pine, Douglas fir, and sparse undergrowth (Marr and Boyd, 1979) is supported by these soils.

Mean annual precipitation in the Bear Creek basin ranges from 760 mm at the upper elevations to 400 mm on the plains. Above 2300 m, the basin is dominated by snowmelt runoff, with rainfall–runoff dominating below 2300 m. From historical records, Bear Creek appears to be subject to more frequent cloudbursts than most South Platte River tributaries (Follansbee and Sawyer, 1948), although this may result from its proximity to Denver and increased attention to flooding. Streamflow, fed by mountain snowpack, is perennial, with periodic rises from accelerated snowmelt and intense rainstorms.

Discharge in Bear Creek has been gaged at Morrison since 1888, and crest-stage gages have been operated throughout the basin since 1978 (Fig. 1). Mean annual flow (for 77 years through 1992) at Morrison is 1.5 m^3 s^{-1}, with a recorded peak of 244 m^3 s^{-1} in 1896 (Grimm, 1993).

Flash floods are most common on Bear Creek between Evergreen (2135 m elevation) and Morrison (1770 m elevation) (Fig. 1). They are especially frequent on Mt. Vernon Creek and Cold Spring Gulch (Diebold, 1939). Twenty-five flash floods have occurred in the Bear Creek basin since 1876, causing 45 deaths and extensive property damage and leaving geomorphic evidence in the form of flood boulder bars and slackwater and overbank deposits (Grimm, 1993).

Flood deposits along Bear Creek are preserved (i) at sites of rapid energy dissipation, such as tributary junctions, abrupt decreases in channel gradient, and abrupt valley expansions, and (ii) downstream from cross-valley glacial moraines that serve as sediment sources. The coarse-grained flood sediments typically form longitudinal boulder bars that may be differentiated from debris-flow deposits on the basis of clast size and orientation, weathering characteristics, and bar morphology (Costa and Jarrett, 1981; Costa, 1984, 1988; Jarrett and Waythomas, 1995). Deposits from recent floods indicate that, in high-gradient channels, these bars closely approximate the water-surface elevation (Jarrett and Waythomas, 1995; Jarrett and Grimm, 1993, unpubl. data).

3. Methodology

We estimated the magnitude, frequency, and geomorphic effects of flooding at four sites in the Bear Creek basin (Fig. 1). In addition, we measured particle size distribution at 15 sites along Bear Creek. Criteria for selection of the four flood-evaluation sites were (i) relatively straight and uniform reaches that could be most accurately modeled with a step-backwater program, (ii) existence and degree of preservation of recent flood and paleoflood evidence, and (iii) location in the basin with respect to elevation and hypothesized distribution of flood-producing rainfall. Characteristics of all of the study sites are summarized in Table 1. At each of the four flood-evaluation sites, field methods included estimation of peak discharge, geochronologic examination, and measurements of coarse-sediments.

3.1. Discharge estimation

A laser theodolite was used to survey four to eight channel cross-sections at each of the four flood-evaluation sites (Table 1). Cross-sections were located to adequately characterize channel geometry for modeling step-backwater flow (Chow, 1959; Davidian, 1984).

For conditions of uniform flow, discharge is usually computed from the Manning equation that involves channel characteristics, water-surface elevations, and a roughness coefficient (Chow, 1959). Downstream changes in the water-surface profile in a uniform reach are accounted for as losses of energy caused by roughness elements in the channel bed. The Manning equation was developed for conditions of uniform flow in which the water-surface profile and energy gradient are parallel to the streambed and the area, hydraulic radius, and depth remain constant throughout the reach. In natural channels, however, uniform conditions rarely exist (Jarrett, 1984). We assume that the equation also is valid for nonuniform reaches if the energy gradient is modified to reflect only the losses that result from boundary friction (Dalrymple and Benson, 1968).

Step-backwater analysis was done using WSPRO, a program developed by the U.S. Geological Survey for

Table 1
Bear Creek basin study-site characteristics

Site[a]	Elevation (m)	Drainage area (km^2)	Channel gradient	Nearest USGS gage	Paleoflood evidence
Turkey Creek[b]	1768	130	0.025	1.6 km downstream USGS CSG[c] 06711000	2 flood boulder bars
Lower Bear Cr.[b]	1795	425	0.070	0.8 km upstream from USGS gage 06710500	2 flood boulder bars
Mt. Vernon Cr.	1865	25	0.033	USGS CSG 06710600 upstream from mouth	1 flood boulder bar
Saw Mill Gulch[a]	1981	6	0.090	–	1 flood boulder bar
Cold Spring Gulch[b]	2040	14	0.040	–	absent
Parmalee Gulch	2054	15	0.033	USGS CSG 06710900 upstream from mouth	absent
Lower Evergreen[b]	2121	200	0.021	just downstream from USGS CSG 06710350	absent
Upper Evergreen[b]	2170	170	0.019	2.4 km upstream from USGS CSG 06710350	absent
Above Evergreen[b]	2195	110	0.007	5.6 km upstream from USGS CSG 06710350	absent
Cub Creek[b]	2255	130	0.040	2.5 km upstream from USGS CSG 06710400	fine-grained overbank deposits
Upper Bear Cr.[b]	2268	249	0.007	–	1 flood gravel bar and fine-grained overbank deposits
Yankee Creek[b]	2298	23.5	0.028	–	absent
Corral Creek[b]	2300	80	0.018	–	absent
Grass Creek[b]	2530	6	0.024	–	absent
Unnamed Creek[b]	2705	2	0.148	–	absent

[a]Flood-evaluation sites are set boldface.
[b]Clast sample site.
[c]USGS CSG: U.S. Geological Survey crest-stage gage.

estimating flow characteristics in rivers (Shearman, 1991). The step-backwater method evaluates energy losses between any two cross sections caused by non-uniform flow conditions (Davidian, 1984). Although open-channel flow is commonly unsteady and nonuniform, hydraulic step-backwater routines assume steady flow conditions between cross sections. To predict water-surface profiles associated with gradually varied flows, a necessary assumption is that the head loss at a section is the same for a uniform flow having the velocity and hydraulic radius of the section (O'Connor and Webb, 1988). The assumption of steadiness probably is justified because at peak flows the change in stage is minimal over short distances; thus, flow approximates steady conditions (Davidian, 1984; Jarrett, 1986).

Step-backwater programs are limited to modeling nondeformable boundaries, such as bedrock channels or channels with minimal erosion. Selection of an appropriate reach may be the most important part of the hydraulic analysis. Good high-water marks, uniform channel geometry, a uniform or slightly contracting reach, fully effective cross-sectional area, and sufficient length are the basic requirements for reach selection (Benson and Dalrymple, 1967; Davidian, 1984; Williams and Costa, 1988). Cross-section selection includes a minimum of three cross sections that represent the geometry of the reach.

Channel characteristics are obtained from cross-section surveys. The standard-step method is used to balance the Bernoulli (energy) equation between adjacent cross-sections (Chow, 1959). Water-surface elevations are determined for each cross-section at a specified discharge. A range of discharges is used to develop stage–discharge relations at cross-sections where paleostage indicators (PSI), such as the boulder and gravel bars of Bear Creek basin, have been identified. Paleodischarge estimates are then determined from the elevations of specific PSI, which are assumed to represent minimum flood stage.

A number of uncertainties affect paleodischarge estimates; therefore, an essential component of a paleoflood study is a sensitivity analysis of factors affecting the accuracy of discharge estimates (Jarrett and Malde, 1987; Jarrett and Waythomas, 1995). This type of analysis provides information used to determine a range of possible discharges and the most probable discharge and associated hydraulic conditions corresponding to the PSI.

A series of computational runs was made to test the sensitivity of the step-backwater analysis. In the various runs, different roughness coefficients, channel geometries, and gradients were considered. Contraction and expansion coefficients were not varied from the standard values of 0 for contracting reaches and 0.5 for expanding reaches because recent studies indicate that these coefficients have relatively little influence on discharge estimates (Jarrett and Malde, 1987; J.O. Shearman, U.S. Geological Survey, pers. commun., 1992). Values of Manning's coefficient were selected onsite based on guidelines for selecting n values in natural channels and floodplains (Barnes, 1967; Jarrett, 1985; Hicks and Mason, 1991) and then varied over a reasonable range of values to assess the effect on computed discharge.

Changes in channel geometry also were analyzed by running a scour-and-fill scenario for the Lower Bear Creek site. Because this site is confined by bedrock walls and channel bed, the thickness of the fill could be estimated relatively easily and removed from cross sections in the discharge computations.

3.2. Geochronology

Where possible, we sampled the flood deposits for organic materials suitable for radiocarbon analysis and used the characteristics of on-site vegetation to estimate minimum ages for floods. Radiocarbon samples consisted of allochthonous charcoal and wood that could have been reworked from older deposits. The samples thus provide a maximum limiting age for the associated flood sediments (Blong and Gillespie, 1978; Baker and Pickup, 1987). One radiocarbon sample was collected at each of the four primary sites.

Although tree growth is sparse on the flood bars in Bear Creek basin, we sampled one tree at Lower Bear Creek and one tree at Turkey Creek. These trees were sampled with an increment borer to determine the approximate tree age by counting annual growth rings (Phipps, 1985). The trees provide minimum limiting ages for the flood deposits (Costa, 1978; Hupp, 1988).

3.3. Coarse-sediment characteristics

The grain-size distribution of sediments coarser than 2 mm was assessed at 19 locations throughout the Bear Creek basin. Twelve of these locations were paired sites immediately upstream and downstream from tributary junctions with Bear Creek. The Wolman (1954) sampling technique was used to measure the b-axis diameter of 100 clasts at each site. A sampling site was a single depositional flood bar, where present, or an in-channel bar and channel-bed sediment where flood bars were absent. The sampling sites are noted in Table 1. Sites were chosen to minimize human or land-use effects on in-channel grain-size distributions. The presence of granitic clasts at all sampling sites indicated that lithologic control was not a limiting factor on grain size.

4. Results

4.1. Discharge estimation

Table 2 summarizes estimated paleoflood discharges, largest indirect discharges, and competence calculations for the four primary sites. The indirect discharges in Table 2 for the Upper Bear Creek, Cub Creek, and Turkey Creek sites were determined by U.S. Geological Survey personnel immediately after each flood, using the slope–area method, because the gage was destroyed by the flood (Follansbee and Sawyer, 1948; station records for USGS gages 06711000 and 06710400). (We were unable to locate the original station records for the 1896, 1933, 1934, and 1938 floods on Lower Bear Creek.) Slope–area estimates of discharge generally are too large for high-gradient (>0.002) streams (Jarrett, 1986; Quick, 1991). In addition, if a flood in the range of 240 $m^3 s^{-1}$ occurred at the Lower Bear Creek site in 1896, as indicated by the gage notes, flood deposits should verify this. Flood evidence, however, indicates a paleodischarge of

Table 2
Estimated paleoflood discharges and largest gaged discharges for the four primary study sites

Site (elev., m)	PSI[a]	Paleo-Q ($m^3 s^{-1}$)	Gaged Q ($m^3 s^{-1}$) and date	Unit Q[e] ($m^3 s^{-1} km^{-2}$)	Stream power (W/m^2)	Critical stream power (W/m^2)[f]
Turkey Creek (1768)	FB-1[b]	74	77 (1969)	0.57	1370	1365
	FB-2	74		0.57		
Lower Bear Creek (1795)	FB-1	113[c]	117 (1983);	0.27	5205	2026
	FB-2	(estimated to be inundated annually)	131 (1934);	–		
	FB-2		176 (1938); 230 (1933); 244 (1896)			
Cub Creek (2255)	SWD[d]	17	7 (1980)	0.13	207	196
Upper Bear Cr. (2268)	FB-1	17	13 (1980)	0.07	69	148

[a]PSI: Paleostage indicator.
[b]FB: Flood boulder bar.
[c]Preferred discharge resulting from sensitivity analysis.
[d]SWD: Slackwater deposit.
[e]Ratio based on paleodischarge and equal to estimated peak paleodischarge divided by drainage area.
[f]Critical stream power per unit area estimated to be necessary to entrain the D_{max}, after Costa (1983).

Table 3
Sensitivity analysis of hydraulic factors for the Lower Bear Creek site using step-backwater analysis

Run	Description[a]	Discharge ($m^3 s^{-1}$)
1	Cross sections as surveyed $n=0.08$, 0.07 $nd=0.3$, 1.5 m contraction/expansion coefficients = 0, 0.5 slope = 0.07, water-surface = PSI	85
2	Run 1 with $n=0.05$, 0.04	106
3	Run 1 with $n=0.03$ critical depth calculation	110
4	Fill removed from cross sections (avg. 0.6 m), $n=0.05$ slope = 0.07	113
5	Run 3 with $n=0.06$	104

[a] n is Manning's roughness coefficient, for main channel and overbank; nd indicates roughness varied with hydraulic depth; PSI is paleostage indicator.

approximately 113 $m^3 s^{-1}$, assuming that the top of the boulder bar represents the water-surface elevation. Table 2 clearly indicates the decrease in unit discharge with increasing elevation.

Comparisons of stream power per unit area in the channel thalweg at peak discharge, and critical stream power using Costa's relations developed for Front Range channels (Costa, 1983), indicates that critical stream power is exceeded at the two lower-elevation sites, but not at the highest elevation site (Table 2). Actual stream power is relatively low at the Turkey Creek site as a result of a relatively wide channel (45 m, as compared to 18 m at the other sites) and low gradient. Critical stream power is barely exceeded at the Cub Creek site, where a steep, confined channel produces high values of stream power. We consider these calculations to provide order-of-magnitude estimates because, as noted by Costa (1983), there are potentially large uncertainties involved in these calculations.

Table 4
Results of radiocarbon and dendrochronologic analyses in Bear Creek Basin

Sample site	Radiocarbon ages		Dendrochronologic age (yr)
	Reported age[a]	Calibrated age[b]	
Lower Bear Creek	8575±190 (GX-18271)	7546 B.C. (7890–7436)	22
Turkey Creek	2556±53 (GX-18272-AMS)	782 B.C. (799–564)	70
Cub Creek	6355±110 (GX-18273)	5277 B.C. (5428–5221)	
Upper Bear Creek	365±110 (GX-18275)	1490 A.D. 1605 A.D. (1437–1655) 1613 A.D.	

[a]Age in radiocarbon years before present, with lab number in parentheses. GX indicates Geochron Laboratory, AMS indicates accelerator mass spectrometer date.
[b]Radiocarbon age calibrated in calendric year B.C. or A.D., with age range in parentheses (Stuiver and Reimer, 1993).

A sensitivity analysis conducted at the Lower Bear Creek site evaluated the effect of channel fill or scour, channel gradient, and roughness coefficients on the water-surface profile. Estimated ranges for these variables were based on data from recent floods in the Colorado Rockies (McCain et al., 1979; Jarrett and Costa, 1986; Jarrett, 1987, 1990). The hydraulic factors and computed discharges are listed in Table 3.

Results of the sensitivity analysis indicate that two factors account for the greatest variability in the discharge estimates. These factors are uncertain flow-resistance coefficients and the water-surface elevation of the paleofloods. Changes in channel configuration since the flood (that is, the area of the flood bar in cross sections) did not substantially affect discharge estimates at this site because the bar is located on the inside

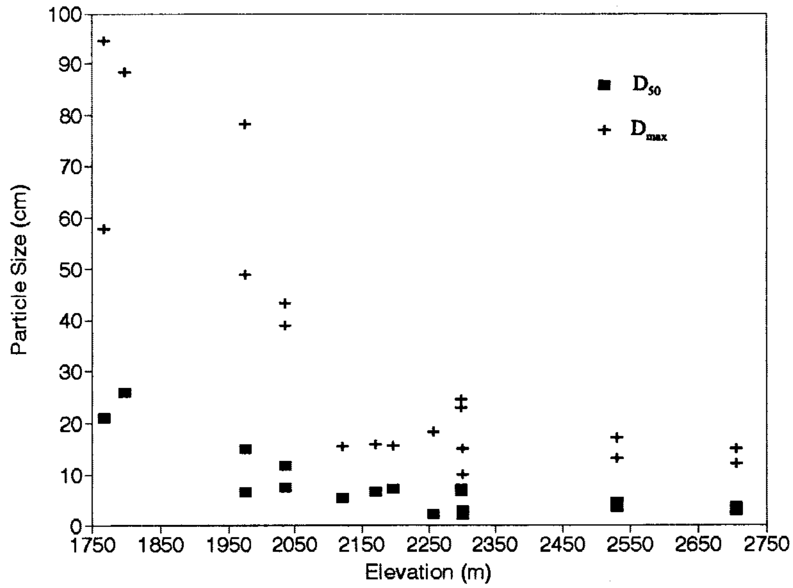

Fig. 2. Particle-size characteristics in relation to elevation. D_{50} indicates intermediate diameter of median clast size measured at each site; D_{max} indicates intermediate diameter of largest clast judged to be fluvially transported on the basis of proximity to the active channel, and clast rounding and imbrication.

Table 5
Particle sizes upstream and downstream from six major tributary junctions on Bear Creek

Sample site	Particle size[a] (cm)			Elevation (m)	Sorting = $(D_{84}-D_{16})/D_{50}$
	D_{16}	D_{50}	D_{84}		
Saw Mill Gulch (u)	3.0	6.4	24.4	1981	3.3
Saw Mill Gulch (d)	4.6	14.9	35.4		2.1
Cold Spring Gulch (u)	3.7	7.3	23.5	2040	2.7
Cold Spring Gulch (d)	5.5	11.6	21.3		1.4
Yankee Creek (u)	3.7	6.7	11.3	2298	1.1
Yankee Creek (d)	4.0	6.7	12.8		1.3
Corral Creek (u)	1.6	2.8	4.4	2300	1.0
Corral Creek (d)	1.1	1.9	3.1		1.0
Grass Creek (u)	1.6	3.4	6.0	2530	1.3
Grass Creek (d)	2.6	4.4	7.0		1.0
Unnamed Creek (u)	2.3	3.6	5.5	2705	0.9
Unnamed Creek (d)	1.8	2.7	4.4		1.0

[a] D_{16}, D_{50}, and D_{84} represent particle size for which 16, 50, and 84% of the distribution, respectively, is finer.
[b] (u) means upstream, (d) downstream.

of a channel bend in an area of relatively ineffective flow. In addition, the area of the bar is relatively small compared to the total cross-sectional area. We believe that the sensitivity analysis for the Lower Bear Creek site is representative of the magnitude of uncertainty in discharge estimation for each of the four study sites. Therefore, we believe that the trend of decreasing unit discharge with increasing elevation (Table 2) is accurate.

4.2. Geochronology

Results of the radiocarbon and dendrochronologic analyses are summarized in Table 4. These results very loosely bracket the ages of the flood deposits. We believe the radiocarbon ages to be more representative of the age of the flood deposits than the dendrochronologic ages at these sites. Although repeated timber harvest throughout the basin has resulted in removal of older trees growing on flood deposits, the age of trees present provide a minimum estimate for the age of the flood deposit. The radiocarbon ages indicate that the gaged or historically estimated discharges at the study sites have not been exceeded during a period several times the length of the historical record.

4.3. Coarse-sediment characteristics

Particle-size distributions for the 19 sites in the Bear Creek basin indicate a substantial decrease in the mean diameter (D_{50}) from 26 cm at the lower elevations to 2 cm in the upper basin, and a corresponding decrease in the size of the largest fluvially transported clast (Fig. 2). At tributary junctions below 2300 m, however, particle sizes are locally controlled by the introduction of coarse material from small, flood-producing tributary basins. Comparison of particle-size distributions immediately upstream and downstream from tributaries above and below 2300 m (Table 5) indicates substantial differences between paired sites below 2300 m, but very similar distributions above 2300 m. Particle-size sorting also decreases below 2300 m. We were limited to six tributary junctions by the need to avoid sites potentially disturbed by human activities.

5. Discussion and conclusions

The existence of an elevation limit on flash flooding in the Colorado Rocky Mountains is supported by discharge estimates from flood sediments in the Bear Creek basin. Peak unit discharge in the basin decreases systematically as elevation increases (Table 2). The geomorphic implications of this relation may be assessed from coarse-sediment deposits. Flood boulder bars are absent along the channels at elevations above about 2100 m (Table 1). The D_{50} and maximum clast size of coarse channel sediments decrease steadily up to this elevation (Fig. 2). The large differences in particle-size distributions upstream and downstream from Cold Spring and Saw Mill Gulches (a *t*-test indicated that the D_{50} values were significantly different upstream and downstream of these tributaries) suggest that these tributaries are subject to flash floods capable of transporting coarse-grained sediments (Table 5). The particle size distributions upstream and downstream from the four tributaries above 2100 m, in contrast, are essentially identical, indicating that these higher elevation basins are not subject to rainfall-induced flash floods.

In addition, the degree of sorting [$(D_{84}-D_{16})/D_{50}$] increases at the high-elevation sites relative to the Saw Mill and Cold Spring Gulch sites. The channel bed sediment upstream from 2100 m is better sorted than the bed sediment downstream. The effect of the tributaries on particle-size distributions at the sites below 2100 m is to introduce coarse-grained material in a variety of particle sizes that substantially change the distributions downstream from the tributary junction. Upstream from 2100 m, the effect of the tributaries is to introduce material of about the same size distribution as that in the main channel. The discrepancy between our estimated elevation limit of 2100 m and Jarrett's (1993) limit of 2,300 m may be insignificant in view of the precision and scarcity of available data. Jarrett's selection of 2300 m was based on stream flow gaging station data covering the late 1880s through 1988 from 935 gages in Colorado. Prior to the work described here, no systematic field studies assessed the variability of the elevation-limit among individual basins and independent of gage locations. The discrepancy could also be caused by basin characteristics such as aspect, which would affect the response of flood-producing storms and introduce minor variations in the elevation limit along the Colorado Front Range.

Drainage basin area is an important factor in determining the amount of runoff that is generated, all other factors being equal. Corral Creek has the largest drainage area of the six tributary basins (Table 1) and, therefore, the largest runoff per unit area potential. Corral Creek, however, has the smallest D_{50} of the six tributaries compared. We interpret this to indicate that flood hydrometeorology (in this case, the predominance of snowmelt rather than rainfall floods in Corral Creek) is a more important control than drainage area on coarse-sediment characteristics.

One of the geomorphic implications of the elevation limit on flash flooding is, therefore, a reversal of the usual downstream-fining trend (Knighton, 1984) in coarse channel sediments. Downstream changes in bed-material characteristics have been attributed primarily to abrasion (Schumm and Stevens, 1973) and sorting (Bradley et al., 1972), although the importance of lithologic controls on weathering (Knighton, 1984), and of sediment supply processes at tributary junctions (Knighton, 1980; Troutman, 1980) have been noted. To our knowledge, however, such changes have not previously been explained in terms of changes in unit discharge associated with flood hydrometeorology. On Bear Creek the D_{50} of channel sediments and the size of the largest fluvially transported clast increase downstream to the mountain front, and the sorting decreases, probably because of an increased transport capacity associated with intense rainfalls. This relation is also influenced by the junction of tributaries below 2100 m that are subject to flash floods.

Although the data presented in this paper are limited to a single basin, they illustrate the importance of flood hydrometeorology in controlling both flood magnitude and coarse-sediment distribution. Differences in peak unit discharge and grain sizes above and below about 2100 m elevation imply differences in sediment transport, channel morphology, and appropriate engineering response. In regions where flood hydrometeorology varies over short distances as a result of aspect or elevation, the geomorphic effects of floods may also vary, altering the usual downstream trends observed in basins with consistent flood hydrometeorology.

Acknowledgements

This research formed a part of Michael Grimm's M.S. thesis, which was supported by the U.S. Geological Survey. We thank Marsha Hilmes for field assistance, Christopher Waythomas and John Elliott for helpful discussions, and Jim Bennett, Gar Williams, and two anonymous reviewers for thorough reviews.

References

Baker, V.R., 1984. Flood sedimentation in bedrock fluvial systems. In: E.H. Koster and R.J. Steel (Editors), Sedimentology of Gravels and Conglomerates. Can. Soc. Petrol. Geol. Mem., 10: 87–98.

Baker, V.R. and Pickup, G., 1987. Flood geomorphology of the Katherine Gorge, Northern Territory, Australia. Geol. Soc. Am. Bull., 98: 635–646.

Barnes, H.H., 1967. Roughness characteristics of natural channels. U.S. Geol. Surv. Water-Supply Pap., 1849: 213 pp.

Benson, M.A. and Dalrymple, T., 1967. General field and office procedures for indirect discharge measurements. U.S. Geological Survey, Techniques of Water-Resources Investigations, Book 3, ch. A1, 30 pp.

Blong, R.J. and Gillespie, R., 1978. Fluvially transported charcoal gives erroneous C-14 ages for recent deposits. Nature, 271: 739–741.

Bradley, W.C., Fahnestock, R.K. and Rowekamp, E.T., 1972. Coarse sediment transport by flood flows on the Knik River, Alaska. Geol. Soc. Am. Bull., 83: 1261–1284.

Chow, V.T., 1959. Open Channel Hydraulics. McGraw-Hill, New York, 680 pp.

Costa, J.E., 1978. Holocene stratigraphy in flood-frequency analysis. Water Resour. Res., 14: 626–632.

Costa, J.E., 1983. Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range. Geol. Soc. Am. Bull., 94: 986–1004.

Costa, J.E., 1984. Physical geomorphology of debris flows. In: J.E Costa and P.J. Fisher (Editors), Developments and Applications of Geomorphology. Springer, New York, pp. 268–317.

Costa, J.E., 1988. Rheologic, geomorphic and sedimentologic differentiation of water floods, hyperconcentrated flows and debris flows. In: V.R. Baker, R.C. Kochel and P.C. Patton (Editors), Flood Geomorphology. Wiley, New York, pp. 113–122.

Costa, J.E. and Jarrett, R.D., 1981. Debris flows in small mountain stream channels of Colorado and their hydrologic applications. Bull. Assoc. Eng. Geol., 18: 309–322.

Dalrymple, T. and Benson, M.A., 1968. Measurement of peak discharge by the slope–area method. U.S. Geological Survey Techniques of Water-Resources Investigations, Book 3, Ch. A2, 12 pp.

Davidian, J., 1984. Computation of water-surface profiles in open channels. U.S. Geological Survey Techniques in Water-Resources Investigations, Book 3, Ch. A14, 48 pp.

Diebold, C.H., 1939. Floods in the Bear Creek watershed, Colorado. U.S. Dept. of Agriculture Rocky Mountain Forest and Range Experiment Station, 40 pp.

Follansbee, R. and Sawyer, L.R., 1948. Floods in Colorado. U.S. Geol. Surv. Water-Supply Pap., 997: 151 pp.

Grimm, M.M., 1993. Paleoflood history and geomorphology of Bear Creek basin, Colorado. M.S. thesis, Colorado State University, Ft. Collins, Colorado, 126 pp.

Gupta, A., 1975. Stream characteristics in eastern Jamaica, an environment of seasonal flow and large floods. Am. J. Sci., 275: 825–847.

Gupta, A., 1988. Large floods as geomorphic events in humid tropics. In: V.R. Baker, R.C. Kochel and P.C. Patton (Editors), Flood Geomorphology. Wiley, New York, pp. 301–315.

Hicks, D.M. and Mason, P.D., 1991. Roughness Characteristics of New Zealand Rivers. Water Resources Survey, Wellington, New Zealand, 329 pp.

Hupp, C.R., 1988. Plant ecological aspects of flood geomorphology and paleoflood history. In: V.R. Baker, R.C. Kochel and P.C. Patton (Editors), Flood Geomorphology. Wiley, New York, pp. 335–356.

Jarrett, R.D., 1984. Hydraulics of high gradient streams. J. Hydraul. Div. Am. Soc. Civ. Eng., 110: 1519–1539.

Jarrett, R.D., 1985. Determination of roughness coefficients for streams in Colorado. U.S. Geol. Surv. Water-Resources Investigations Report 85-4004, 54 pp.

Jarrett, R.D., 1986. Evaluation of the slope–area method of computing peak discharge. In: S. Subitzky (Editor), Selected papers in the hydrological sciences 1986. U.S. Geol. Surv. Water-Supply Pap., 2310: 13–26.

Jarrett, R.D., 1987. Flood hydrology of foothill and mountain streams in Colorado. PhD dissertation, Colorado State University, Ft. Collins, CO, 239 pp., unpubl.

Jarrett, R.D., 1990. Paleohydrologic techniques used to define the spatial occurrence of floods. Geomorphology, 3: 181–195.

Jarrett, R.D., 1993. Flood elevation limits in the Rocky Mountains. In: C.Y. Kuo (Editor), Engineering Hydrology. American Society of Civil Engineering, pp. 180–185.

Jarrett, R.D. and Costa, J.E., 1986. Hydrology, geomorphology, and dam-break modeling of the July 15, 1982, Lawn Lake Dam and Cascade Lake Dam failures, Larimer County, Colorado. U.S. Geol. Surv. Prof. Pap. 1369: 78 pp.

Jarrett, R.D. and Costa, J.E., 1988. Evaluation of the flood hydrology in the Colorado Front Range using precipitation, streamflow, and paleoflood data. U.S. Geological Survey Water-Resources Investigations Report 87-4117, 37 pp.

Jarrett, R.D. and Malde, H.E., 1987. Paleodischarge of the late Pleistocene Bonneville Flood, Snake River, Idaho, computed from new evidence. Geol. Soc. Am. Bull., 99: 127–134.

Jarrett, R.D. and Waythomas, C.F., 1995. Paleoflood hydrology and fluvial geomorphology of Arthurs Rock Gulch near Fort Collins, Colorado. U.S. Geological Survey Water Resources Investigations Report 94-4032, in press.

Knighton, D., 1984. Fluvial Forms and Processes. Edward Arnold, London, 218 pp.

Knighton, A.D., 1980. Longitudinal changes in size and sorting of stream-bed material in four English rivers. Geol. Soc. Am. Bull., 91: 55–62.

Marr, J.W. and Boyd, W.S., 1979. Vegetation in the greater Denver area, Front Range Urban Corridor, Colorado. U.S. Geological Survey Miscellaneous Investigations Series, map I-856-I.

McCain, J.F. and Ebling, J.L., 1979. A plan for the study of flood hydrology of foothill streams in Colorado. U.S. Geol. Surv. Open-File Rep., 79-1276: 29 pp.

McCain, J.F., Hoxit, L.R., Maddox, R.A., Chappell, C.F. and Caracena, F., 1979. Storm and flood of July 31 – August 1, 1976, in the Big Thompson River and Cache la Poudre River Basins, Larimer and Weld Counties, Colorado. Part A, Meteorology and hydrology in Big Thompson River and Cache la Poudre River Basins. U.S. Geol. Surv. Prof. Pap., 1115: 1–85.

Miller, A.J., 1990. Fluvial response to debris associated with mass wasting during extreme floods. Geology, 18: 599–602.

O'Connor, J.E. and Webb, R.H., 1988. Hydraulic modeling for paleoflood analysis. In: V.R. Baker, R.C. Kochel, and P.C. Patton (Editors), Flood Geomorphology. Wiley, New York, pp. 393–402.

Phipps, R.L., 1985. Collecting, preparing, crossdating, and measuring tree increment cores. U.S. Geol. Surv. Water-Resour. Invest. Rep., 85-4148: 48 pp.

Quick, M.C., 1991. Reliability of flood discharge estimates. Can. J. Civ. Eng., 18: 624–630.

Schumm, S.A. and Stevens, M.A., 1973. Abrasion in place: a mechanism for rounding and size reduction of coarse sediments in rivers. Geology, 1: 37–40.

Scott, G.R., 1972. Geologic map of the Morrison Quadrangle. U.S. Geological Survey Map I-790-A.

Shearman, J.O., 1991. User's manual for WSPRO — A computer model for water surface profile computations. Federal Highway Administration, Report No. FHWA-IP-89-027, 177 pp.

Sheridan, D.M. and Marsh, S.P., 1976. Geologic map of the Squaw Pass Quadrangle, Clear Creek, Jefferson County, Colorado. U.S. Geological Survey Map I-786.

Sheridan, D.M., Reed, J.C. and Bryant, B., 1972. Geologic map of the Evergreen Quadrangle, Jefferson County, Colorado. U.S. Geological Survey Map I-786.

Shroba, R.R., Schmidt, P.W., Crosby, E.J., Hansen, W.R. and Soule, J.M., 1979. Storm and flood of July 31 – August 1, 1976, in the Big Thompson River and Cache la Poudre River Basins, Larimer and Weld Counties, Colorado. Part B, Geologic and geomorphic effects in the Big Thompson Canyon area, Larimer County. U.S. Geol. Surv. Prof. Pap., 1115: 86–152.

Smith, J.H., 1964. Geology of the sedimentary rocks of Morrison Quadrangle, Colorado. U.S. Geological Survey Miscellaneous Investigations Map I-248.

Stuiver, M. and Reimer, P.J., 1993. Extended ^{14}C data base Calib 3.0 ^{14}C age calibration program. Radiocarbon, 35: 215–230.

Troutman, B.M., 1980. A stochastic model for particle sorting and related phenomena. Water Resour. Res., 16: 65–76.

Webb, R.H., Pringle, P.T. and Rink, G.R., 1989. Debris flows from tributaries of the Colorado River, Grand Canyon National Park, Arizona. U.S. Geol. Surv. Prof. Pap., 1492: 39 pp.

Williams, G.P. and Costa, J.E., 1988. Geomorphic measurements after a flood. In: V.R. Baker, R.C. Kochel and P.C. Patton (Editors), Flood Geomorphology. Wiley, New York, pp. 65–77.

Wohl, E.E., 1992. Bedrock benches and boulder bars: Floods in the Burdekin Gorge of Australia. Geol. Soc. Am. Bull., 104: 770–778.

Wolman, M.G., 1954. A method of sampling coarse river-bed material. Trans. Am. Geophys. Union, 35: 951–956.

Wolman, M.G. and Gerson, R., 1978. Relative scales of time and effectiveness of climate in watershed geomorphology. Earth Surf. Process., 3: 189–208.

Wolman, M.G. and Miller, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. J. Geol., 68: 54–74.