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# Recent large-magnitude floods and their impact on valley-floor environments of northeastern Yellowstone

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**Abstract**

The Lamar River watershed of northeastern Yellowstone contains some of the most diverse and important habitat in the national park. Broad glacial valley floors feature grassland winter range for ungulates, riparian vegetation that provides food and cover for a variety of species, and alluvial channels that are requisite habitat for native fish. Rapid Neogene uplift and Quaternary climatic change have created a dynamic modern environment in which catastrophic processes exert a major influence on riverine–riparian ecosystems. Uplift and glacial erosion have generated high local relief and extensive cliffs of friable volcaniclastic bedrock. As a result, steep tributary basins produce voluminous runoff and sediment during intense precipitation and rapid snowmelt. Recent major floods on trunk streams deposited extensive overbank gravels that replaced loamy soils on flood plains and allowed conifers to colonize valley-floor meadows. Tree-ring dating identifies major floods in 1918, ca. 1873, and possibly ca. 1790. In 1996 and 1997, discharge during snowmelt runoff on Soda Butte Creek approached the 100-year flood estimated by regional techniques, with substantial local bank erosion and channel widening. Indirect estimates show that peak discharges in 1918 were approximately three times greater than in 1996, with similar duration and much greater flood plain impact. Nonetheless, 1918 peak discharge reconstructions fall well within the range of maximum recorded discharges in relation to basin area in the upper Yellowstone region. The ~1873 and 1918 floods produced lasting impacts on the channel form and flood plain of Soda Butte Creek. Channels may still be locally enlarged from flood erosion, and net downcutting has occurred in some reaches, leaving the pre-1790 flood plain abandoned as a terrace. Gravelly overbank deposits raise flood-plain surfaces above levels of frequent inundation and are well drained, therefore flood-plain soils are drier. Noncohesive gravels also reduce bank stability and may have persistent effects on channel form. Overall, floods are part of a suite of catastrophic geomorphic processes that exert a very strong influence on landscape patterns and valley-floor ecosystems in northeastern Yellowstone. 

*Keywords:* Floods; Flood plain; Gravel; Yellowstone National Park; Stream incision; Riparian

## 1. Introduction

In contrast to the rhyolite plateau that extends over much of Yellowstone National Park, NE Yellowstone and the Absaroka Range are characterized by steep topography and high relief of up to 1000 m, primarily on friable Eocene volcaniclastic rocks of the Absaroka Volcanic Supergroup (e.g., Prostka et al., 1975) (Figs. 1 and 2). As noted by Pierce and Morgan (1992), rapid Plio-Pleistocene uplift related to the Yellowstone "hotspot" thermal anomaly (e.g.,

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Smith and Braile, 1993; Humphreys et al., 2000) is probably the primary factor generating summit elevations over 3000 m and major relief on these nonresistant rocks. Extensive Quaternary glaciation was also promoted by the anomalously high elevations of the greater Yellowstone area (Pierce, 1979) and has played a major role in sculpting deep valleys with oversteepened sidewalls. The geologic evolution of NE Yellowstone produced a postglacial landscape that is highly susceptible to catastrophic geomorphic processes, including mass movements and flooding. The imprint of such processes on valley-floor environments is substantial, reinforcing the "geoecosystem" concept of strong linkages between physical and biological processes in greater Yellowstone. Here I focus on major flood events of the past few hundred years as an example of how fundamental geologic conditions control modern valley-floor and stream channel environments through geomorphic and hydrologic processes.

## 2. Stream systems of northeastern Yellowstone

The Lamar River drains a 1709-km$^2$ basin in the NE corner of Yellowstone, where the eastern border of the park follows the Lamar basin boundary (Fig. 1). Two large tributaries, Slough Creek and Soda

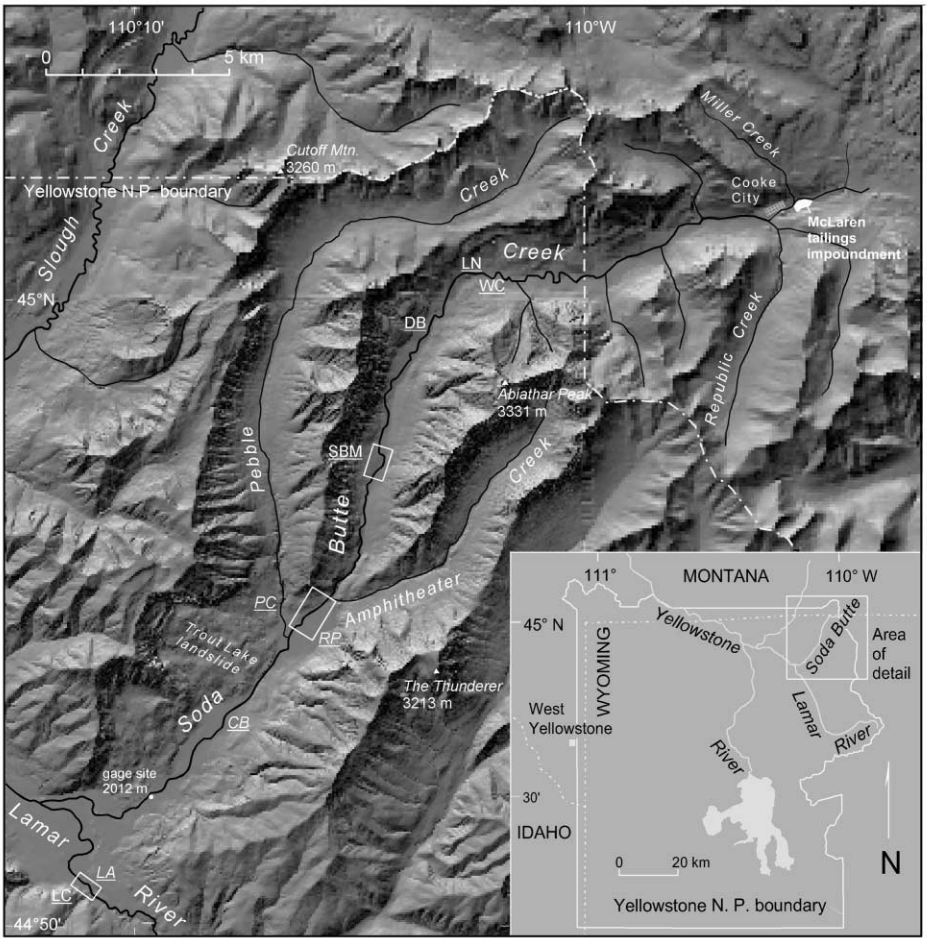



Fig. 1. Digital shaded relief map of the study area in the Lamar River basin, NE Yellowstone National Park area, showing elevations of selected peaks and the Soda Butte Creek stream gaging site. Dendrochronology sites are indicated by underlined letters, with discharge reconstruction sites in italics; boxes show areas mapped in Fig. 7.

Butte Creek, originate in the Absaroka and Beartooth Ranges NE of the park. These three trunk streams occupy broad glacial trough floors and resemble the wandering gravel-bed rivers described by Church (1983) and Nanson and Croke (1992) in deglaciated mountain regions (Fig. 2). Alluvial channels predominate, with shorter bedrock or bedrock-confined reaches. As is typical of mountain streams, upper reaches are strongly influenced by hillslope processes (e.g., Hack and Goodlett, 1960; Grant and Swanson, 1995). Broad glacial valley floors, however, allow large volumes of landslide and alluvial fan material to be stored along valley margins, buffering sediment input to trunk channels (Meyer, 1995; Meyer et al., 1995) (Fig. 2). Channels are confined to varying degrees by fan and landslide deposits, and flood-plain widths are thus highly variable. Channel patterns are also quite diverse. Sinuous single-thread channels exist locally in the lowest-gradient reaches (valley slope 0.002–0.004), particularly on Slough Creek, where glacial erosion has exposed resistant Archean granitic gneiss beneath the Paleozoic sedimentary cover and produced a stepped valley floor (Figs. 1 and 3). More commonly, trunk valleys show smoother long profiles with gradients of 0.005–0.013 in alluvial reaches. Braided channels with broad active bars are present, commonly below valley constrictions, and appear to be in part controlled by valley slope and local bed load sediment supply from steep, high-relief tributaries (e.g., Slough Creek below Cutoff Creek, and Soda Butte Creek below Republic Creek) (Fig. 3). Straighter channels exist where streams are confined by alluvial fans and/or rock avalanche deposits. In the reach between the Trout Lake landslide knickpoint and the gage site, parts of Soda Butte Creek show nearly

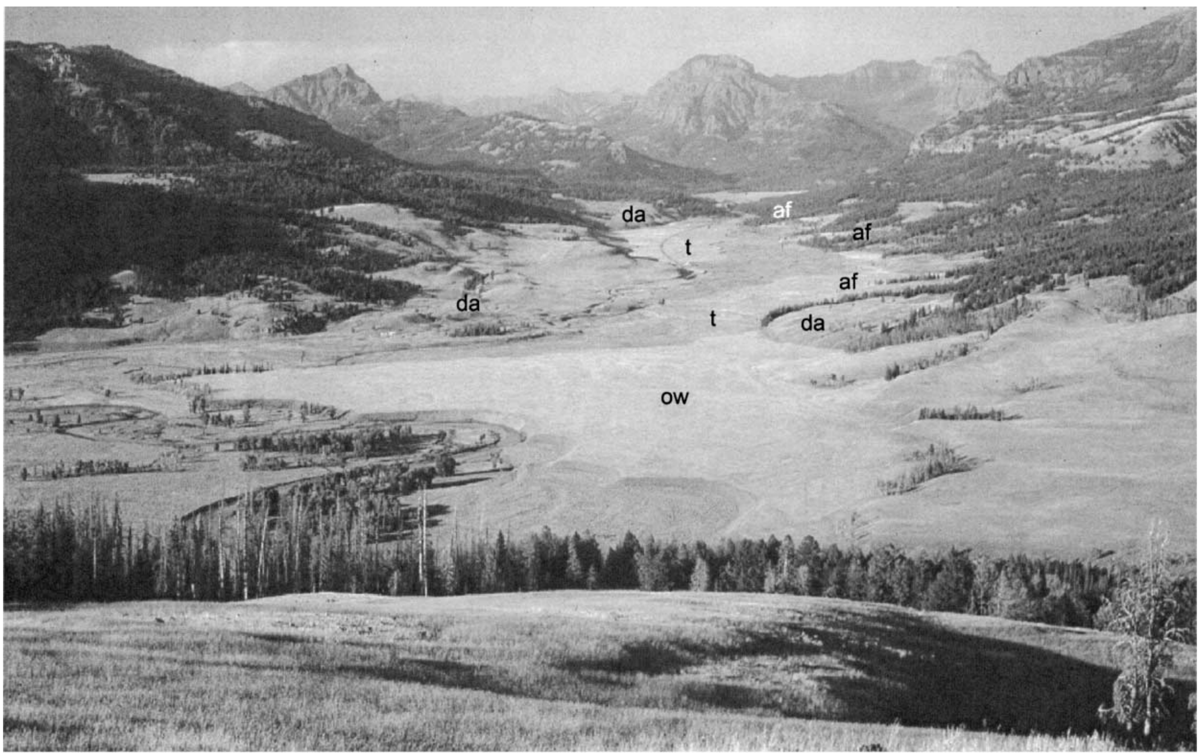



Fig. 2. Photograph looking NE up the Soda Butte Creek valley from the ridge near the SW corner of Fig. 1. The Lamar River flows across the near valley floor to the Soda Butte Creek confluence just off the left edge. A latest glacial outwash surface (ow) forms the prominent terrace in the middle ground; Holocene terraces (t) flank Soda Butte Creek (Meyer et al., 1995). Debris avalanche deposits (da) and alluvial fans (af) border the valley floor in this relatively broad section but more closely confine trunk streams farther up valley.

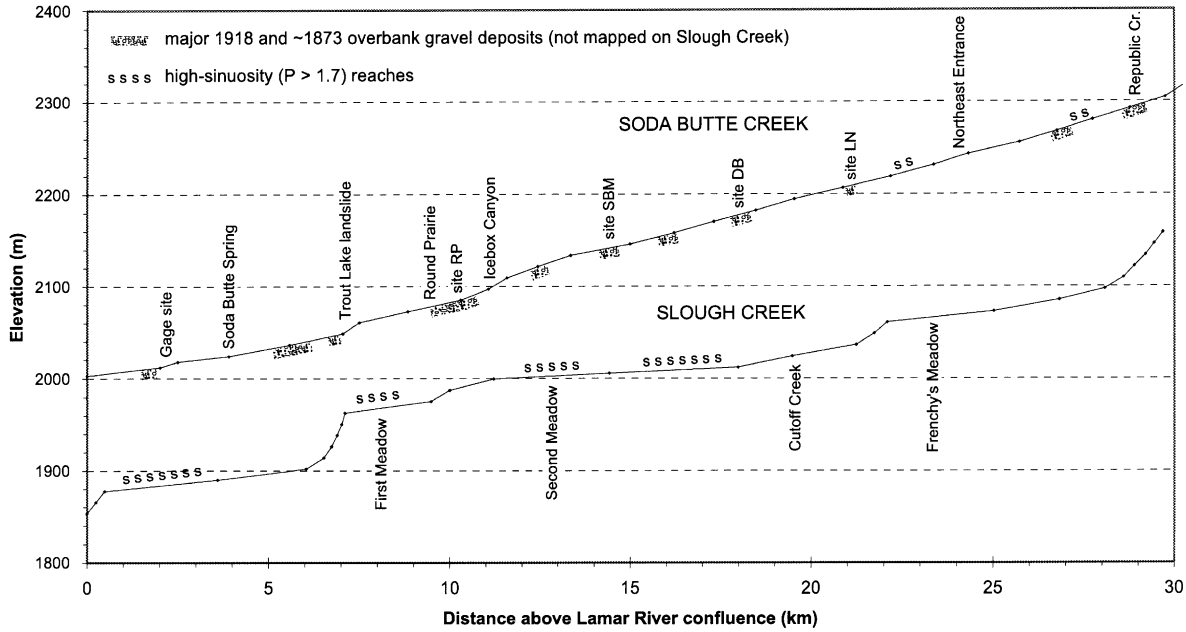



Fig. 3. Valley profiles for Soda Butte Creek (showing areas of major overbank gravel deposition during the 1918 and ~1873 floods) and Slough Creek. Steep reaches on Slough Creek are formed by granitic gneiss steps, except just above the Lamar confluence where the stream flows over a bouldery moraine. Less resistant volcaniclastic rocks result in a smoother Soda Butte profile. The Trout Lake landslide knickpoint was formed where slumping and lateral spread of the toe of a large late-Pleistocene debris-avalanche deposit constricted the creek ca. 2000 $^{14}$C years BP (Meyer, 1995).

anastomosing patterns, with sinuous channels separated by relatively stable grassy islands.

Although most of the Lamar River system has not been directly disturbed by human activities, metal mining and processing have been conducted in the Cooke City–New World mining district since around the time of park establishment in 1872 (Fig. 1). This activity in the Soda Butte Creek headwaters has produced geochemical and some hydrologic impacts on downstream flood-plain environments, including a tailings dam-break flood in 1950 (Meyer, 1993; Meyer and Watt, 1999; Stoughton and Marcus, 2000; Marcus et al., 2001). Also, some observers believe that park management policies since establishment of Yellowstone in 1872 have resulted in accelerated stream erosion through overbrowsing of riparian vegetation by elk (reviewed in Wagner et al., 1995; Huff and Varley, 1999), but data on channel changes are few. In addition, little research has assessed controls on present stream morphology in the park from a long-term systems perspective. In part to provide background for these issues, this paper explores the effects of natural floods on valley-floor environments and implications for ecosystem processes in NE Yellowstone.

### 2.1. Climatic and geomorphic controls on flood potential

Several characteristics of the climate, vegetation bedrock geology, and geomorphology in the Lamar River basin result in a high potential for flood generation and sediment production. On a regional scale, the strong topographic gradient from the Lamar Valley onto the Beartooth Plateau and Absaroka Range results in heavy winter snowfall in the headwaters, with mean annual precipitation of over 1000 mm at higher elevations (> 2800 m). Mountain slopes support dense mixed-conifer forests, but large expanses of bare bedrock are common above timberline (Fig. 2). In contrast, semiarid conditions exist in the broad lower valley floors at ca. 2000 m elevation, where mean annual precipitation is ~300–450 mm and much precipitation falls as rain in late spring through summer (Dirks and Martner, 1982; Whitlock and Bartlein, 1993). Sagebrush–grassland vegetation is dominant, with more open conifer stands on lower slopes.

The moderate elevation and steep slopes of most tributary drainages allow very rapid snowmelt runoff if unusually warm weather occurs in late spring and early summer. Localized summer convective storms occur throughout the basin, and rainfall intensity is probably enhanced by orographic lifting of the dominant airflow as storms track to the northeast. The potential for the largest floods on trunk streams of the Lamar River system most likely lies in prolonged, basin-wide heavy rain on melting snow during warm weather in late spring to early summer (Church, 1988; U.S. Geological Survey, 1991).

An important factor in flood potential in the Lamar basin is the large area of steep, exposed bedrock that generates runoff during snowmelt and rainstorms. Eocene andesitic volcaniclastic rocks of the Lamar River, Wapiti, and Mount Wallace Formations form upper valley walls and include abundant muddy laharic debris-flow sediments, with minor basaltic volcaniclastic rocks and lava flows (Prostka et al., 1975). These rocks are poorly indurated and weather rapidly, liberating voluminous mud-rich sediment as well as coarser clasts. The abundant fines are readily stripped from steep slopes and also provide muddy matrix for transport of larger clasts in debris flows. Extensive exposed bedrock results in low infiltration rates and high runoff generation. In alpine areas with more resistant bedrock, bouldery, high-permeability talus deposits help to absorb and retard runoff from cliffs above, but such talus is uncommon in NE Yellowstone, probably because of the muddy texture and high erodibility of the weathered volcaniclastic material. Although the bedrock itself provides much loose sediment, colluvium is typically thin except in interfluvial footslope areas and slopes are largely weathering-limited. Storage of colluvium in hollows and eventual en masse failure (e.g., Reneau and Dietrich, 1987) are relatively rare in the steep volcaniclastic terrane.

Along with copious runoff production, the high erodibility of bedrock in steep tributary basins has aided in the development of very dense drainage networks and a high flash-flood potential. Excluding the large basins of Amphitheater Creek and Pebble Creek (Fig. 1), drainage density for Soda Butte Creek tributary basins determined from 1:24,000-scale topographic maps averages 8.4 $km^{-1}$ and ranges up to 17.9 $km^{-1}$ (O'Hara, 1994). The steep, highly dissected terrain has a badlands-like character, but with much greater relief than is typical of more familiar shale badlands of the western United States. About 15% of tributary alluvial fans along the Soda Butte and Slough Creek valleys have deeply incised channels from erosion in flash floods (Fig. 4).

Fig. 4. Photograph of the east valley wall of Soda Butte Creek near Abiathar Peak (Fig. 1) showing a small steep tributary basin with extensive exposed volcaniclastic bedrock in the upper basin, and incised channels and recent debris flow-flash flood deposition on its alluvial fan. The fan channel near the road has been modified by maintenance activities.

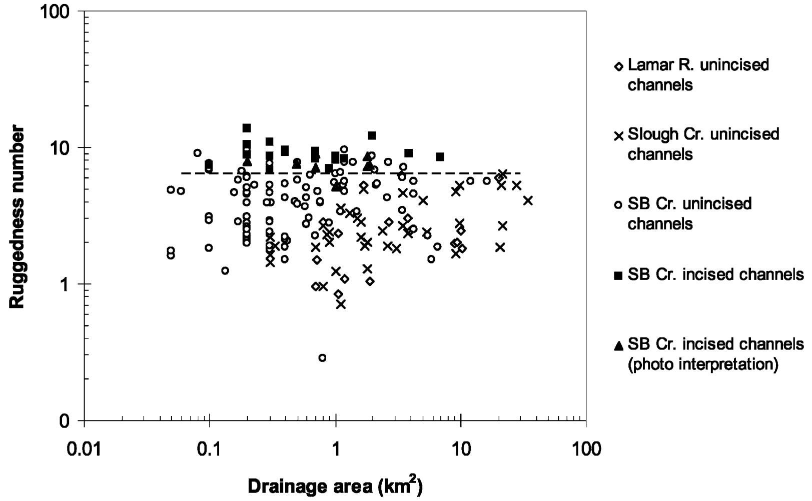



Fig. 5. Ruggedness number in relation to basin area for 181 small tributary basins in the Lamar River system, showing the threshold value of 6.5 (dashed line) for fan channel incision. Ruggedness number alone correctly predicts the state of the channel in 91% of cases.

Ruggedness number is a dimensionless parameter equal to drainage density times total relief (Patton and Baker, 1976). Nearly all basins of incised fans have a ruggedness number greater than 6.5, emphasizing the importance of basin morphometry in flash-flood generation (Fig. 5). Flash floods issue from these small, steep basins during intense summer convective storms (Meyer, 1995), and they are probably important in rapid runoff generation during rain-on-snow events as well.

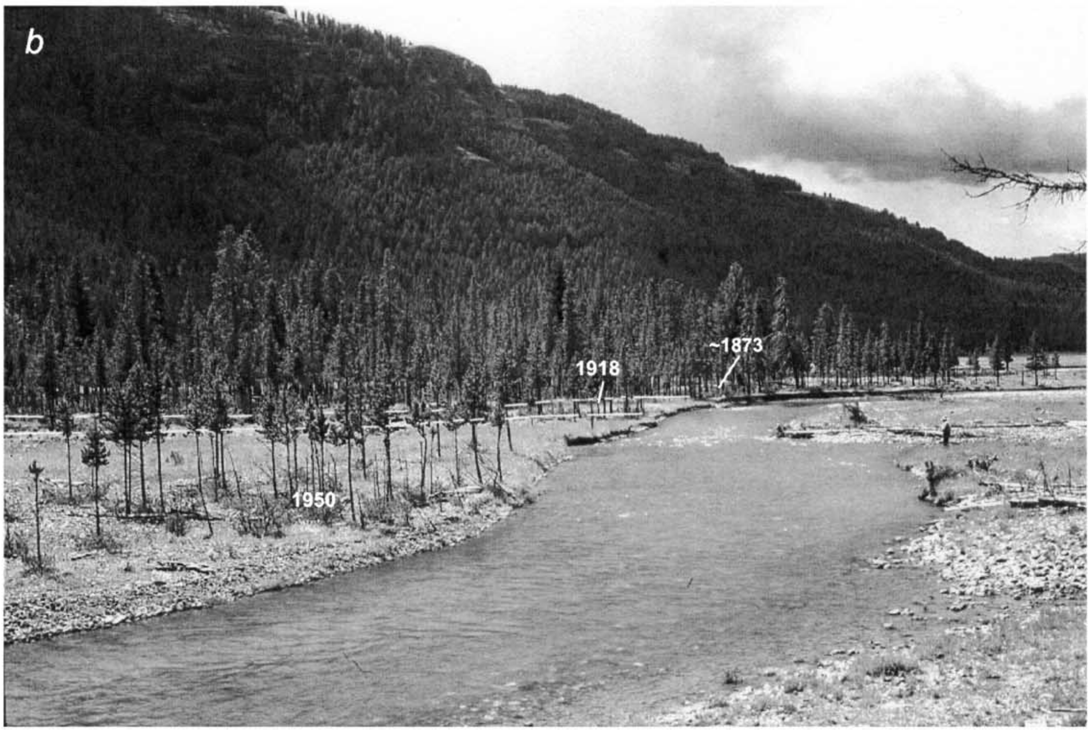
### 2.2. *Evidence for recent major floods on trunk streams*

Notable in the Lamar River and Soda Butte Creek drainages are broad sheets of gravel on higher flood-plain surfaces. These overbank gravels imply deposition by major floods with competence to move bed load materials across flood plains that normally experience only low-energy flow (Costa, 1974a; Ritter, 1975). Overbank gravel surfaces 1 m or more above bankfull stage are common, particularly inside large channel bends and at valley widenings. Bankfull stage on many streams has a recurrence interval of about 1.5 years; thus associated flood plains receive frequent fine-grained overbank sedimentation (Wolman and Leopold, 1957; Wolman and Miller, 1960). In this paper, bankfull stage is defined by this approximate recurrence interval for reference purposes. Although some flood-plain surfaces are inundated by discharges of $>1.5$-year frequency in the Lamar River system, especially in low-gradient reaches, many flood-plain surfaces lie well above the 1.5-year stage because of overbank gravel deposition.

At the head of Round Prairie below Icebox Canyon, nested flood plain levels with subtle relief

Fig. 6. (a) Photograph of the Soda Butte Creek flood plain in upper Round Prairie (upstream part of site RP, Fig. 7). View is to the SE. Note even-age stands of trees on overbank gravels with little grass cover and differences in stand density and height on gravel surfaces of different age. (b) View downstream from point shown in (a) of even-age stand of small lodgepole pines on left bank on a gravel bar reworked in the 1950 flood. Angler on right bank provides scale. Older stands on overbank gravels appear in middle ground. Note low-angle gravelly banks in most of reach and low-relief steps between inset flood-plain surfaces of different age that are reflected in the left bank height.


a
~1790
~1790
~1873
1918
1950
b


b
~1873
1918
1950

between surfaces are mantled by overbank gravel (Figs. 6 and 7). Conifers, primarily lodgepole pine (*Pinus contorta*), with minor Engelmann spruce (*Picea engelmannii*) and a few Douglas-fir (*Pseudotsuga menziesii*), appear to grow preferentially on the gravel instead of silty–sandy flood-plain sediments. These trees are of distinctly similar height on each level but vary markedly between levels, suggesting "even-age" stands that colonized the gravels after deposition. Similar relationships in several study

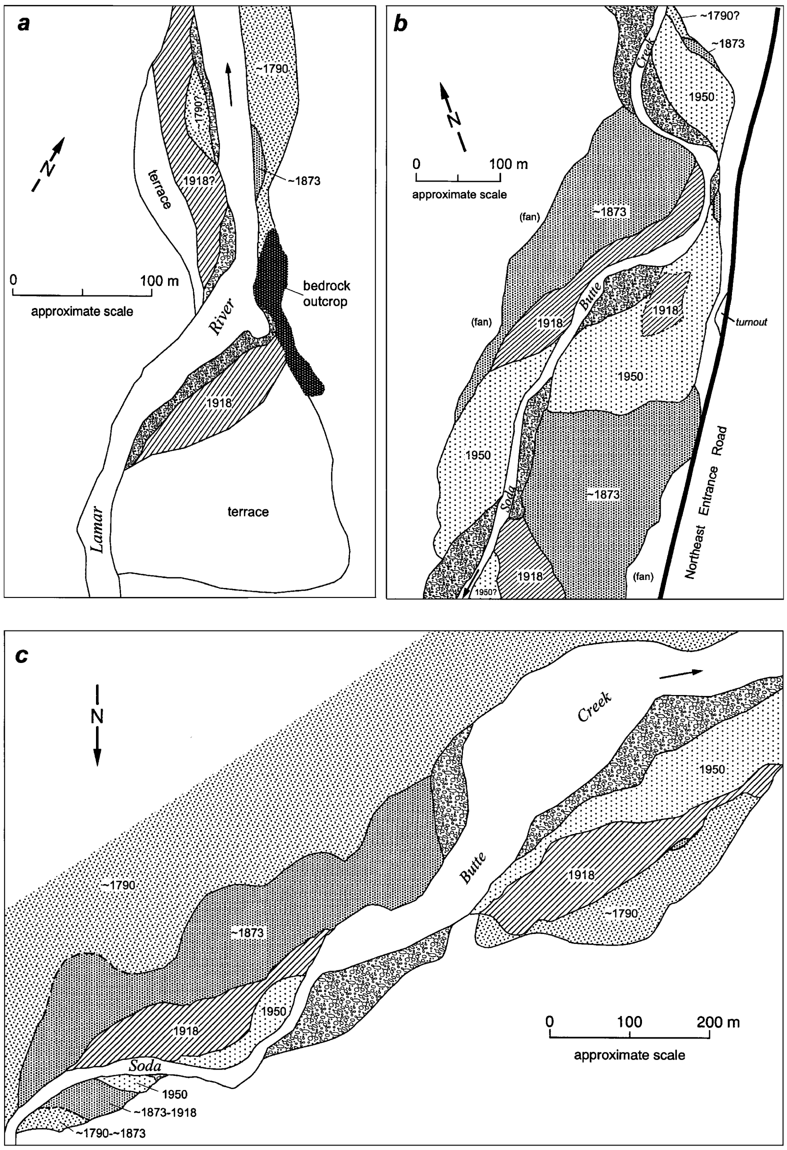



Fig. 7. Maps of flood-plain surfaces affected by major floods at (a) sites LA (NE bank) and LC (SW bank), Lamar River above the Soda Butte Creek confluence; (b) site SBM on middle Soda Butte Creek; and (c) site RP, Soda Butte Creek in upper Round Prairie. Some overbank flow in younger floods extended onto older flood surfaces, but primary erosional and depositional effects are within mapped limits.

reaches provide evidence of the successive major floods investigated herein.

## 3. Methods

Tree-ring methods allow both dating of flood events (e.g., Helley and LaMarche, 1973; Hupp, 1988) and investigation of how flooding controls vegetation patterns on valley floors (e.g. Everitt, 1995; Scott et al., 1996). Indirect techniques were used to estimate peak discharges for selected major floods to better understand the physical processes responsible for channel and flood plain impacts.

### 3.1. Tree-ring dating of floods

To obtain a minimum age for flood gravel deposits, between 5 and 85 (mean $n = 23$) of the taller and larger-diameter trees were cored on individual flood deposits to locate the oldest trees. Sample size depended on the deposit area and total number of trees, with the primary intent of leaving a very low probability that the oldest tree was left uncored. To estimate a calendar-year date for a flood, the lag time between the flood and the earliest date from trees on its deposits must be considered. This lag has two components: (i) most tree cores probably do not include the first few years of growth. Trees were cored as low as possible, usually $< 10$ cm above the root crown, to minimize this error. (ii) An ecesis period typically exists between flood gravel deposition and initial tree germination. Conifers colonize recently deglaciated terrain in about 5 to 60 years (Sigafoos and Hendricks, 1969; McCarthy and Luckman, 1993). Conifer establishment on gravelly flood-plain sediments probably occurs within the shorter end of this time range, but relevant data are limited. Helley and LaMarche (1973) noted that within 5 years or less, conifers had germinated on bars deposited in the 1964 floods in northern California. Trees on overbank gravels deposited by historical floods of known age were used to calibrate tree-ring ages. Along with identification of even-age stands of trees, inset relations, deposit texture, and thickness of overbank sediment cover allowed mapping of flood surfaces for events of different ages (Figs. 6 and 7).

### 3.2. Discharge reconstruction

Peak discharge reconstructions for the 1918 flood were made at site LA on the Lamar River above the Soda Butte Creek confluence, site RP on Soda Butte Creek in Round Prairie, and site PC on Pebble Creek in the canyon 1 km upstream of the Soda Butte confluence (Figs. 1 and 3). Discharge estimates were made by (i) the slope-area method using three cross-sections (Benson and Dalrymple, 1967), (ii) a simplified slope-area method that eliminates roughness by correlation with slope (Riggs, 1976), and (iii) average velocity estimated from maximum size of flood-transported clasts (Costa, 1983). Imbrication was used as evidence of flood transport of the largest clasts. Stage indicators included overbank gravels, erosional scarps, and flood depth estimated from maximum clast size and water surface slope (Costa, 1983). Because of deformable alluvial channels and the short length of stable reaches with suitable stage indicators, step-backwater modeling is unsuitable for discharge reconstruction at the study sites (O'Connor and Webb, 1988).

Because the primary source of potential error lies in alluvial channel change both during and following the 1918 flood, sites with relatively stable banks were chosen, mostly in straight reaches. Channels at these sites showed little change after the large 1996 and 1997 floods. The Soda Butte Creek site at the head of Round Prairie has boundaries entirely in alluvial deposits, but several trees growing directly on the banks predate 1918, indicating that the channel at this site has not widened significantly since their establishment (Fig. 6a). The Lamar River channel is bounded by bedrock on the west bank at the upper end of the discharge reconstruction reach (Fig. 7). The Pebble Creek site is in the canyon above the campground, 1 km upstream of the Soda Butte confluence. It has bedrock-confined banks; however, the floor of the channel is composed of boulder gravel that may have scoured during peak flow in this relatively steep reach. To represent uncertainty in discharge estimates, minimum and maximum values were calculated using the full range of possible stage heights, roughness, and slope; also, potential scour and fill were considered in channel cross-sections.

## 4. Results and discussion

### *4.1. Known historical floods in 1918 and 1950 and tree-ring calibration*

Anecdotal records document a major flood on the Lamar River system in June 1918 that is likely to have produced overbank gravel deposits. The Lamar River bridge on the Northeast Entrance road below Lamar Canyon was washed out on June 11 (National Park Service, 1918). This wooden bridge was about 1 km downstream of the current structure and had a deck roughly 2 m above bankfull level (Meagher and Houston, 1998, plate 55.1). At the Yellowstone River gage at Corwin Springs, MT, near the northern park boundary (Fig. 1), the second highest discharge (906 $m^3$ $s^{-1}$) since 1911 was recorded on June 14, 1918. This peak was only slightly lower than the June 1997 maximum of record (911 $m^3$ $s^{-1}$), and the 1918 peak stage was actually 0.18 m higher than in 1997. Stream gaging began in 1924 on the Lamar River and in 1989 on Soda Butte Creek, thus direct discharge measurements are not available for 1918.

Flood gravels with even-age stands closely postdating 1918 are common along Soda Butte Creek, its tributary Pebble Creek, and the upper Lamar River (Figs. 6 and 8). Ring counts at five sites give ages for the oldest trees that postdate the 1918 flood by an average of ~10 years (sites DB, SBM, LA, PC, and RP; Table 1) (Bingham and Meyer, 1994; Simpson and Meyer, 1995). Lag time reflects both the oldest growth rings not obtained in cores and ecesis, which may vary between sites with seed and water availability and other environmental factors. The Round Prairie (RP) site with the 17-year lag is particularly sparsely forested (Fig. 6), suggesting that coarse 1918 flood gravels at this site may hinder conifer establishment.

A 1950 flood was caused by failure of the tailings impoundment for the McLaren gold mine, located directly on Soda Butte Creek just above Cooke City (Meyer, 1993; Marcus et al., 2001). Although small spills were commonly reported, local residents recall a major dam break in June 1950 (Glidden, 1999). The specific causes of this event are uncertain but likely involve inflow from upper Soda Butte Creek that exceeded the culvert drain capacity. National Park Service records (Johnston, 1950) note that the

Table 1
Sample information for tree-ring sites

| Sites | | Flood dates | | | |
|---|---|---|---|---|---|
| | | 1950 | 1918 | ~1873 | ~1790 |
| WC | *n* | 20 | | | |
| | lag time | 7 | | | |
| LN | *n* | 18 | | | |
| | lag time | 4 | | | |
| DB | *n* | 4 | 13 | | |
| | lag time | 4 | 8 | | |
| SBM | *n* | 26 | 25 | 28 | |
| | lag time | 7 | 8 | 10 | |
| PC | *n* | | 12 | | |
| | lag time | | 10 | | |
| RP | *n* | 12 | 22 | 85 | 60 |
| | lag time | 9 | 17 | 9 | 10 |
| LA | *n* | | 17 | 5 | 11 |
| | lag time | | 10 | 9 | 14 |
| LC | *n* | | | | 18 |
| | lag time | | | | 38 |

*n* = number of trees cored; lag time = number of years between flood date and age of oldest ring obtained in all tree cores at site.

McLaren dam failed on June 23, 1950, because of heavy rain, but precipitation was recorded only on June 22 and June 24. Hourly data for June 24 show that almost 50 mm of precipitation fell on the Northeast Entrance of Yellowstone, but the daily total is listed as only 22 mm (U.S.D.C. Weather Bureau, 1950).

Peak discharge in upper reaches of Soda Butte Creek was as much as one order of magnitude greater than the 100-year flood; but the dam-break flood was brief, almost certainly $<1$ h duration, and the flood wave attenuated strongly downstream (Epstein and Meyer, 1997; Marcus et al., 2001). Thus, relatively little bed load transport is likely to have occurred. Dendrochronology identified a few side-attached channel gravel deposits that were either deposited in this flood or were older bars reworked in 1950 (Figs. 6b–8). Silty overbank tailings deposits, however, are widespread. Despite their acidic and metals-enriched character, they were rapidly colonized by lodgepole pines. The short ecesis period may occur because tailings deposits are toxic to grasses (Stoughton and Marcus, 2000), greatly reducing competition with lodgepole seedlings. Also, their silt loam texture provides for high water retention compared to gravel. The oldest rings from trees on two gravelly deposits postdate the 1950 flood by an average of 8 years (sites SBM and RP), whereas the oldest rings from tailings deposits postdate the flood by an average of 5 years (sites DB, LN, and WC) (Table 1). Overall, data from known-age floods indicate that for gravelly surfaces the typical lag time between a flood event and the oldest tree age is 8–10 years.

### 4.2. Tree-ring evidence for a flood ca. 1873

By adding a 9–10 year lag time to ring counts, tree-ring data from higher, older gravelly flood surfaces give evidence for a flood in the early 1870s at sites SBM and RP on Soda Butte Creek and site LA on the Lamar River (Simpson and Meyer, 1995) (Figs. 1 and 8; Table 1). Dendroclimatic studies in Yellowstone indicate that the spring of 1873 was an anomalously wet period in NE Yellowstone (Douglas and Stockton, 1975); thus, 1873 is used as a first approximation for the flood date. A ca. 1885 photograph of middle Soda Butte Creek shows a broad gravelly channel with a large active chute across a meander, as well as the currently active outer meander (Meagher and Houston, 1998, plate 62.1). The chute is currently inactive and reforested. These observations are consistent with the occurrence of a large flood shortly predating the photo. At site RP, broad overbank gravel sheets extending several hundred meters down valley were deposited where the ∼1873 flow exited the constriction caused by lower Icebox Canyon and the Amphitheater Creek fan toe (Figs. 6 and 7).

Amphitheater Creek was apparently not a dominant source of these gravels, as even-age stands of trees do not mark deposits on its fan. Although 1918 flood gravels overlap onto the ∼1873 flood surface just below the constriction at the head of Round Prairie, 1918 deposits are clearly inset below the ∼1873 surface a few hundred meters down valley (Fig. 6b), suggesting intervening channel incision. At site SBM, trees germinating after the ∼1873 flood colonized sand and gravel filling a large channel where the flood may have caused avulsion and channel abandonment. On the Lamar River at site LA, an early 1870s flood is suggested by limited tree-ring data from a small area inset just below the highest surface (Fig. 7).

### 4.3. Tree-ring evidence for a flood ca. 1790(?)

The highest flood-plain surface at site RP shows elongated stands of conifers that radiate from the head of Round Prairie, a pattern that appears to be controlled by underlying fluvial gravels deposited by divergent flood flows (Fig. 6a). A large sample shows that these trees germinated as early as ∼1800 (Fig. 8; Table 1); however, trees germinating after about 1830 are much more common. A small sample of the largest trees on slopes immediately adjacent to the flood plain yielded dates from 1750 to 1800. This indicates that the maximum age of the flood-plain trees was not set by forest fire because a stand-replacing fire on the flood plain would very likely have killed these adjacent trees. Also, the oldest trees on the highest flood-plain surface of the Lamar River date to ca. 1800 (Fig. 8), suggesting a common environmental control with tree growth on the highest surface in Round Prairie. The significance of the maximum tree age is less clear than for younger stands on flood gravels, as fewer trees approach the maximum observed age. I therefore tentatively identify a major flood ca. 1790.

In contrast to younger surfaces, the highest (∼1790 flood) surface at site RP has a thin but nearly

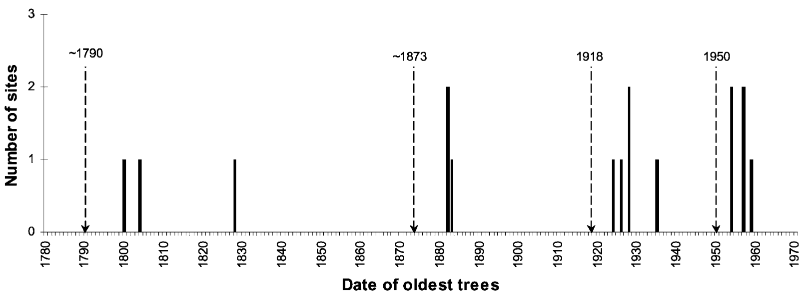



Fig. 8. Maximum ages of trees cored on flood deposits on Soda Butte Creek, Pebble Creek, and the Lamar River (Table 1).

continuous cover of fine overbank sediment. Although ~1873 and 1918 flood gravels are at the level of this surface at the head of Round Prairie (Figs. 6 and 7), a few hundred meters down valley the highest surface lies ~0.5 m above the ~1873 level, and it continues as a broad low terrace along the south side of the valley. Only a few probable corresponding remnants are present on the north side of Soda Butte Creek. This highest flood-plain surface in Round Prairie may correlate to the T4 terrace of the lower Soda Butte valley, which similarly lies a few decimeters above the highest flood bars of probable ~1873 and 1918 age (Meyer, 1995). Radiocarbon dates indicate that overbank deposition on the T4 terrace began ca. 1200 AD (Meyer et al., 1995). In sum, tree-ring data in Round Prairie suggest that downcutting from the T4 level to the present was initiated ca. 1790, possibly with a major flood event, and has continued with subsequent large floods.

### *4.4. Comparison of discharge, power, energetics, and geomorphic impact of floods*

Estimates of 1918 peak discharge at the three cross-section sites are Lamar River, 500 $m^3$ $s^{-1}$; Soda Butte Creek, 170 $m^3$ $s^{-1}$; and Pebble Creek, 64 $m^3$ $s^{-1}$, with uncertainty ranges of $\pm$10–50% (Table 2). The 1918 peak discharge on Soda Butte Creek at Round Prairie greatly exceeds the maximum flood of record (~49 $m^3$ $s^{-1}$ in June 1996, adjusted to the Round Prairie site from the downstream gage location using basin area and elevation in a regional flood frequency equation from Omang et al. (1986)). The much greater magnitude of the 1918 flood is supported by the observation that the 1996 peak stage either did not reach the level of 1918 overbank gravel deposits or barely spilled into lower channels on the 1918 surfaces. At the Round Prairie discharge reconstruction site, for example, maximum clast sizes in overbank gravels indicate a flow depth of ~0.5–1 m *over* the flood-plain surface in 1918 (Costa, 1983), whereas the 1996 discharge was contained entirely within the channel. Also, little overbank gravel deposition was observed along Soda Butte Creek in 1996 and 1997.

The 1918 peak discharge estimates can be evaluated in the context of stream gage records in the Yellowstone region. Log-Pearson type III flood frequency analysis yields a 100-year flood discharge ($Q_{100}$) of Soda Butte Creek for 66 $m^3$ $s^{-1}$ (adjusted

Table 2
Discharge and stream power estimates for the 1918 flood

| Discharge method | Estimate[a] | Lamar River | Soda Butte Creek | Pebble Creek | |
|---|---|---|---|---|---|
| | | LA | RP[b] | PC1 | PC2 |
| Slope-area ($m^3$ $s^{-1}$) | Min | 520 | 88 | 42 | 40 |
| | Mid | 590 | 170 | 59 | 68 |
| | Max | 650 | 290 | 76 | 99 |
| Simplified slope-area ($m^3$ $s^{-1}$) | Min | 430 | – | 42 | 31 |
| | Mid | 500 | – | 51 | 48 |
| | Max | 570 | – | 59 | 65 |
| Threshold velocity ($m^3$ $s^{-1}$) | Min | 420 | 140 | 68 | 57 |
| | Mid | 450 | 180 | 74 | 68 |
| | Max | 490 | 220 | 82 | 93 |
| Average[c] ($m^3$ $s^{-1}$) | Min | 450 | 110 | 55 | 46 |
| | Mid | 500 | 170 | 64 | 63 |
| | Max | 550 | 260 | 75 | 88 |
| Std. dev. of midrange means | | 64 | 10 | 10[d] | |
| Peak stream power (W $m^{-2}$) | | 330 | 280 | 2700 | |
| Drainage area ($km^2$) | | 752 | 148 | 67 | |

[a] Minimum, midrange, and maximum discharge reconstructions.
[b] Simplified slope-area method not used because of subdivided cross-section to account for overbank flow.
[c] Mean of results from two slope-area methods averaged with threshold-velocity result for each site.
[d] Standard deviation of six results from nearby reaches PC1 and PC2.

to the Round Prairie site), but the short 10-year record creates large uncertainty in this value. Regional flood frequency equations for the upper Yellowstone region (Omang et al., 1986) produce a lower $Q_{100}$ of 52 $m^3$ $s^{-1}$, but the gage records employed are also limited ($<52$ years; mean $=26$ years) and omit high discharges in the 1990s or before 1930. Despite these uncertainties, the 1918 peak discharge on Soda Butte Creek is about 2.5–3 times $Q_{100}$. Although the 1918 Lamar system peak discharges exceed $Q_{100}$ by factors of 2 to 4, they are well within the range of maximum recorded discharges in the region, and fall far below the envelope curve for maximum floods in the United States (Costa, 1987) (Fig. 9). The 1918 flood was thus of large but expectable magnitude, especially given the high runoff-generating potential of the Lamar River system. The relatively lower 1918 peak on Pebble Creek may be due to the lesser amount of steep exposed bedrock in this basin.

Stream power estimates provide a means to evaluate the potential for channel and flood plain changes during the 1918 flood (Table 2). Stream power per unit channel boundary area in W $m^{-2}$ is defined as:

$$\omega = \rho g Q S w^{-1} \tag{1}$$

where $\rho$ is the density of the fluid, $g$ is the acceleration of gravity, $Q$ is discharge, $S$ is energy slope (approximated by water surface slope), and $w$ is water-surface width. Although 1918 peak stream power for the Lamar River and Soda Butte Creek was not extreme (Baker and Costa, 1987), it was near the value of $\sim 300$ W $m^{-2}$ suggested by Magilligan (1992) as a threshold for major erosion in humid-region alluvial channels. Unit stream power in broader and lower-gradient reaches would be correspondingly lower. The stream power estimates are therefore consistent with field evidence suggesting local channel enlargement. The value at Pebble Creek is an order of magnitude greater than the other sites because of its steep, closely bedrock-confined channel. In the alluvial reach between the canyon site and Soda Butte Creek, unit stream power was in the range of 50–200 W $m^{-2}$. This was likely sufficient

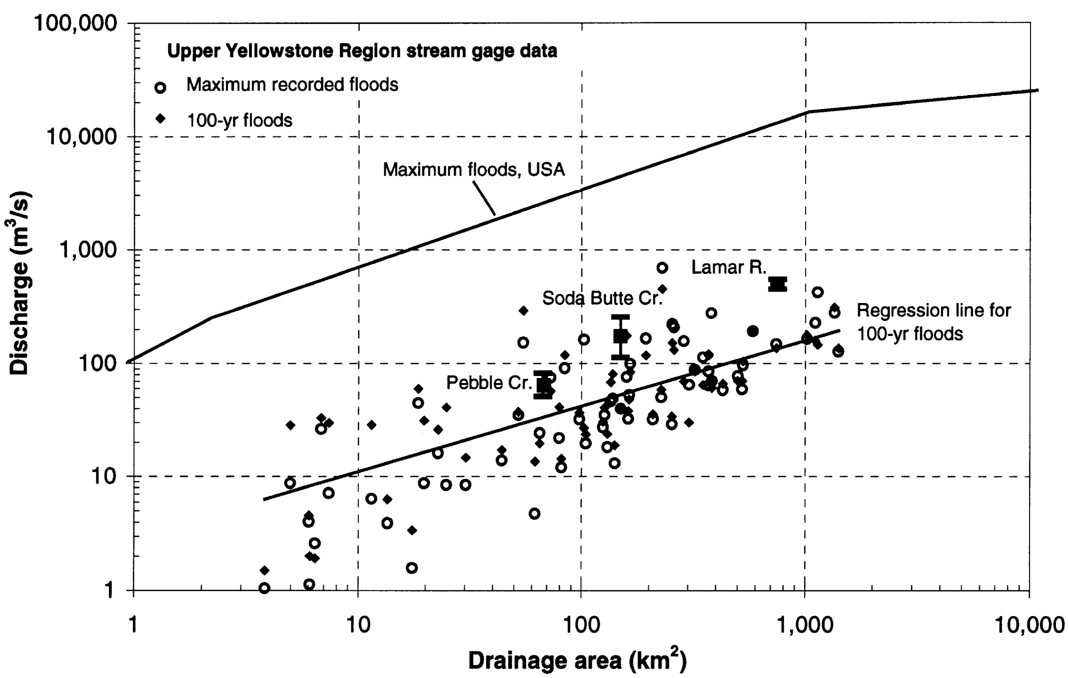



Fig. 9. Reconstructed peak discharges for the June 1918 flood in NE Yellowstone (points with error bars) compared to maximum gaged floods and estimated 100-year floods in the upper Yellowstone River–Central Mountain Region of Montana (Omang et al., 1986). Envelope curve is for maximum rainfall–runoff floods in the U.S. (Costa, 1987).

for major channel alteration, because even at much lower discharges in 1996 and 1997, bank erosion seriously impacted the campground along this reach.

Although likely initiated by heavy rainfall, the June 1950 flood peak in mainstem Soda Butte Creek was produced almost entirely by the tailings dam failure. Indirect discharge estimates declined rapidly downstream from the dam, opposite of the expected pattern for a rainfall–runoff flood, and no coincident flood peak was recorded on the Lamar River. Despite peak discharges in upper Soda Butte Creek as much as one order of magnitude greater than regional 100-year flood estimates (Omang et al., 1986), little flood plain or channel modification is discernible between 1949 and 1954 air photos (Epstein and Meyer, 1997; Meyer and Watt, 1999). The lack of erosion is attributable to the short duration of the 1950 flood; reconstructed hydrographs indicate that the entire event lasted $<1$ h, even with attenuation at downstream locations. Discharge and stream power likely exceeded threshold values for erosion for only a matter of minutes; and total energy expended was small, $<3\times10^5$ J m$^{-2}$ (cf. Costa and O'Connor, 1995).

In contrast to the 1950 dam-break flood, the 1996 and 1997 snowmelt runoff floods each expended 5 to $7\times10^7$ J m$^{-2}$ over a 16-day period at site CB on lower Soda Butte Creek (Fig. 1), nearly three orders of magnitude greater energy than the 1950 dam-break flood. Bank erosion was only locally significant, but lateral bank retreat of up to 23 m occurred during the 1997 flood on Soda Butte Creek, with one major meander cutoff. Air photo analysis revealed that bankfull width increased significantly at four of five study reaches between 1987 and 1997, with most change likely occurring in 1996 and 1997 (Mowry, 1998). The 1918 flood hydrograph at the Yellowstone River–Corwin Springs gage shows that discharge was maintained near the peak flow for 5–6 days. Because 1918 peak discharges were also approximately three times greater than in 1996, the 1918 flood had the potential for substantially more geomorphic work, with channel enlargement and overbank gravel deposition the expectable result.

The 1918 flood in the upper Yellowstone River system was generated by snowmelt combined with rainfall runoff (U.S. Geological Survey, 1991), and high flows were maintained at the Corwin Springs gage from about June 10 to 24. Temperatures in the region were abnormally high in mid-June 1918 (U.S.D.A. Weather Bureau, 1918). At Mammoth in NW Yellowstone (1900 m elevation; Fig. 1), maximum temperatures rose to 28–31°C from June 9–14 and fell only slightly until June 24. Minimum temperatures remained between 9°C and 13°C over the period; thus, freezing levels were probably $>3000$ m elevation, promoting rapid snowmelt. In addition, about 69 mm of precipitation fell at Mammoth from June 10–24. Regionally, heavy thunderstorms were noted during the flood period, especially in eastern Idaho and NW Wyoming (U.S.D.A. Weather Bureau, 1918). Although rainfall amounts were moderate at Mammoth, orographic effects probably produced much greater convective-storm rainfall in the Lamar River headwaters, adding to snowmelt to generate the 1918 flood. Because most of the Yellowstone River basin above the Lamar River confluence is controlled by the substantial reservoir effect of Yellowstone Lake, floods in the Lamar system are not necessarily matched by equally large unit-area discharges at the Corwin Springs gage (Fig. 1).

Probable subsequent channel modifications prevent accurate reconstruction of peak discharge for the $\sim1873$ flood. The relative extent of overbank gravel deposition on Soda Butte Creek during the $\sim1873$ event suggest that its magnitude was significantly greater than the 1918 flood (Figs. 6a and 7). Its peak discharge is unlikely to have exceeded that of 1918 by more than a factor of two, and a discharge of up to this magnitude would still be within the expectable range given regional flood frequency relations (Bingham and Meyer, 1994; Omang et al., 1986) (Fig. 9). On the basis of overbank gravel extent in Round Prairie, the $\sim1790$ flood also appears to have been larger than 1918 (Fig. 6a). Although the 1996 and 1997 floods caused local channel change, clearly no event since 1918 has produced peak flows and flood plain impact comparable to the earlier floods.

### *4.5. Changing flood magnitudes and climate*

Studies of flood history over the mid- to late Holocene emphasize that large floods tend to cluster in time, and that relatively small changes in climate may produce major changes in flood magnitude (Ely et al., 1993; Knox, 1993). Nonstationary flood magnitudes create a fundamental problem for flood frequency statistics because frequency values are strongly dependent on the magnitude of events over the period of observation. Flood probabilities calculated from gage records may poorly represent flood regimes in past or future hydroclimatic conditions (Klemeš, 1989). Flood frequency estimates in the Lamar system clearly depend on the interval of observation, especially if the large floods identified herein are included. The ~1873 and 1918 floods in Yellowstone are broadly associated with the global transition toward warmer conditions at the end of the Little Ice Age, ca. 1900 (e.g., Grove, 1987; Kreutz et al., 1997). Trends of increasing temperature and decreasing winter precipitation characterize Yellowstone instrumental climate records from 1895–1990 (Balling et al., 1992a,b). At face value, these trends imply a general decrease in snowpacks, thus smaller snowmelt–runoff floods. Unseasonably high temperatures during the early snowmelt period, however, appear to be a key factor in flood generation that may be enhanced with a warming climate. Also, despite recent trends, heavy snowpacks did accumulate in the 1996 and 1997 water years. If snowmelt had been accompanied by prolonged high temperatures or heavy rainfall, flooding could easily have matched or exceeded the 1918 magnitude.

Terraces along lower Soda Butte Creek preserve some evidence of Holocene flood processes (Meyer, 1995; Meyer et al., 1995). Formation of the T4 terrace tread was essentially coincident with the Little Ice Age. Although several lines of evidence suggest a cooler period in the Yellowstone region from about 1300 to 1900 AD (e.g., Meyer et al., 1992, 1995; Millspaugh and Whitlock, 1995; Hadly, 1996), little detailed paleoclimatic information is presently available. The relatively narrow T4 flood plain suggests that channel lateral migration and stream power were relatively moderate. Abandonment of the T4 flood plain after ca. 1790 may have resulted from a period of increased flood magnitude represented by the events documented herein. These floods shortly preceded or were coincident with the early phase of warming out of the Little Ice Age, a dramatic climatic trend over the last century that is apparent both globally (e.g., Crowley, 2000) and in Yellowstone instrumental climate records (Balling et al., 1992a,b). Similarly, downcutting from the T3 to T4 terrace levels also occurred during a transition toward warmer climate; radiocarbon dating places incision around 1200–1150 $^{14}$C years BP (ca. 800–850 AD) with the onset of the Medieval Warm Period. Gravelly flood bars on a low, scoured T3 tread below the Soda Butte Creek gaging station suggest that T3–T4 incision was associated with major floods. Similarly, gravelly bars intermediate in height between the $T_D$ and $T_E$ terraces on the Lamar River below site LA are consistent with flood-related incision sometime between 2600 and 2000 $^{14}$C years BP, but only limited evidence is preserved in each case (Meyer, 1995; Meyer et al., 1995).

Links between climatic change and flood magnitude in Yellowstone remain tentative, but an apparent concentration of large-magnitude floods in the last few hundred years has also been observed in other western U.S. river systems, particularly in the late 1800s and early 1900s (e.g., Ely et al., 1993). This broad-scale similarity in flood history suggests that climatic conditions play an important role in changing flood magnitudes.

### *4.6. Flood-plain change and implications for valley-floor ecosystems*

Some of the most significant and persistent ecological effects of large floods since ~1790 stem from the conversion of flood-plain surfaces from loamy overbank deposits to gravel. Little evidence exists to indicate pre-existing stands of conifers in areas of overbank gravel deposition. Conifers would likely have been abundant in these areas if they were previously mantled by gravel. Because overbank flow in many of these areas was relatively shallow and divergent, flood processes such as scouring and impacts by floating debris were probably insufficient to completely uproot and remove large conifers. If older stands were present, trees predating ~1790 should survive at least in lower-energy parts of the flood plain. Trees of this age exist only on terraces or slopes above the flood-plain study sites. On 1918 overbank gravel deposits, minor driftwood is preserved that was probably left by that flood, but very few old uprooted trees are present. I infer that a change from loamy flood-plain soils with high moisture retention to well-drained flood gravel allowed rapid colonization of many former grass and sedge meadows by conifers. Little herbaceous cover exists on the xeric soil surfaces under the conifers.

Enlargement of alluvial channels is a typical consequence of major floods, especially where channel gradient is relatively high and abundant coarse bed load sediment is transported (e.g., Costa, 1974b; Kochel, 1988). Persistent enlargement of some reaches of Soda Butte Creek is suggested by the observation that many flood-plain surfaces inundated by the ∼1873 and 1918 floods received little or no overbank flow during the 1996 flood, as a discharge of 1996 relative magnitude would rise well over the banks of typical alluvial channels. Data are lacking, however, to directly document changes in channel dimensions since 1918. A persistent local increase in channel capacity could stem from raising of the flood-plain surface by aggradation of overbank gravel or by lowering of the channel bed by incision, as well as through channel widening by bank erosion. Probably, these have all occurred locally, with their importance in a given reach dependent on slope, net bed material flux, and erodibility of the channel margins. In 1996, the most consistently inundated flood plains were in reaches with lower channel slopes in the lower parts of broad valleys, e.g., lower Round Prairie, well below areas where overbank gravel was deposited. This suggests that bank height increase from overbank gravel deposition is a primary factor in increased channel capacity, but low-gradient reaches are less likely to have undergone major width increase or incision in floods, and post-flood narrowing by bar deposition should also proceed more rapidly (Kochel, 1988).

The causes of decline of riparian tall-willow communities in NE Yellowstone since the late 1800s have been the subject of much debate (Huff and Varley, 1999). Chadde and Kay (1991) invoked elk browsing as the sole cause, but other factors are likely involved, including increasing drought that produced lower stream flows and generally milder winters (Balling et al., 1992a,b), with greater elk survival and access to winter range. Notable declines in willows occurred during the prolonged 1930s drought (Singer et al., 1994). In addition, the combination of gravelly deposits and local channel incision produced by large floods has created drier conditions less suitable for willows on large areas of flood plain.

Recovery time for alluvial channel morphology, flood plain character, and riparian vegetation after major flood events varies greatly with initial impact, climate, and event sequencing (Wolman and Gerson, 1978; Beven, 1981). On lower-gradient streams in humid regions, well-vegetated flood plains and banks tend to limit channel modification. Recovery of channels to the pre-flood condition in humid systems may be very rapid ($<1$ to 10 years) even after major floods because frequent flows rebuild bars and fill widened channels, and vegetation is quickly re-established (Hack and Goodlett, 1960; Moss and Kochel, 1978; Kochel, 1988). Channel reaches with greater transport competence may remain enlarged for longer periods, however (Costa, 1974b). In sparsely vegetated semi-arid systems, lower resistance may allow major flood plain and channel change with large-magnitude floods. Infrequent channel-forming flows and slow vegetation growth may also retard recovery for long periods (Wolman and Gerson, 1978). The NE Yellowstone environment generally falls between these humid and semi-arid cases, but equally important in determining recovery time is the high stream power and energetic gravel transport during floods on Soda Butte Creek and the Lamar River (e.g., Kochel, 1988). Major floods replace more densely vegetated and cohesive fine-grained banks with drier, loose overbank gravels that retard recovery of grasses and willows. Persistent bank instability results, and channel widening may occur in subsequent more moderate floods. In a more extreme example of changing flood plain states, Nanson (1986) documented that confined flood plains along some high-energy Australian streams experience catastrophic stripping of fine overbank sediments in major floods with recurrence intervals of hundreds to thousands of years. Slow replacement of the loamy cover occurs by vertical accretion over intervening periods.

Our observations indicate that major flood-induced modifications to alluvial channels and flood plains of the mountainous Lamar River system may persist locally over periods $>100$ years, which prompts several questions: in the future, will these flood plains recover from flood effects and evolve under a regime of frequent small to moderate floods and gradual channel migration, producing flood-plain surfaces largely mantled by fine overbank sediments as in classic models (Wolman and Leopold, 1957; Wolman and Miller, 1960)? Some terrace treads along lower Soda Butte Creek preserve broad flood plains with single-thread channel patterns and a thick cover of fine sediments (e.g., the T3 terrace), suggesting formation dominated by such a regime (Meyer, 1995; Meyer et al., 1995). Or, will major flood disturbance continue to occur with magnitude and frequency as in the past few hundred years, thus maintaining texturally and topographically diverse flood plains? Even the T3 terrace locally preserves gravelly surfaces as evidence of major floods. The terrace sequence of lower Soda Butte Creek shows that vertical stability of channels has been maintained for only several hundred years between episodes of incision in the middle to late Holocene. Thus, an additional possibility is that the modern flood plains represent a transitional state; lower Soda Butte Creek may still be in the process of downcutting from the T4 terrace level to attain a new grade, with major floods as the primary mechanism of threshold excess and channel incision (Bull, 1988). Regardless of the precise answers to these questions, it is likely that over 100-year time scales floods will produce a dynamic and changeable valley-floor environment in NE Yellowstone.

## 5. Conclusions

(i) Major floods in 1918, ca. 1873, and probably ca. 1790 in the Lamar River system were of much greater magnitude than those observed in more recent gage records. Substantially greater flood plain effects occurred in the 1918 and earlier floods through prolonged high stream power. The 1918 flood produced peak discharges on Soda Butte Creek about 2.5–3 times greater than the maximum gaged discharge in 1996. Although the 1950 tailings dambreak flood produced extreme discharge in upper Soda Butte Creek, its short duration resulted in little geomorphic impact. Substantial flood-plain pollution, however, stemmed from deposition of metal-contaminated overbank sediments.

(ii) The 1918 and earlier floods have had lasting ecological effects on many parts of Soda Butte Creek and the Lamar River through: (a) scouring of loamy flood-plain soils; (b) deposition of overbank gravels and local channel incision that raise flood-plain surfaces well above the 1.5-year stage, isolating the flood plain from frequent inundation and fine sediment deposition, and resulting in drier soil conditions; and (c) possible persistent channel widening, especially below confined reaches. Conifers typically colonize xeric overbank gravels within about 10 years, replacing former meadow and willow communities. Where loose gravels replace cohesive and densely vegetated banks, bank stability also decreases.

(iii) Large-magnitude floods in NE Yellowstone may be more common in some climatic intervals. Limited evidence suggests a link between increased flood magnitude and a warming climate. Centennial-length periods of unusually large floods may also be responsible for the episodic downcutting of stream systems over postglacial time (Meyer et al., 1995).

(iv) Flood plains of the Lamar River system are characterized by a significantly greater diversity of processes, both in time and space, than classic models of flood-plain development via gradual channel migration and frequent fine overbank deposition would imply (Wolman and Leopold, 1957; Wolman and Miller, 1960). In NE Yellowstone, episodic large-magnitude floods on trunk streams are but one of several climate-related disturbance regimes; others include forest fires and flash-flood to debris-flow events (Meyer, 1995; Meyer et al., 1995). Although a high flood potential is clearly not unique to this dynamic mountain environment, the geologic setting of the Yellowstone geoecosystem exerts a fundamental control on flood processes. Valley-floor ecosystems in NE Yellowstone have been subjected to rapid changes in the physical environment over postglacial time. With the strong possibility of rapid anthropogenic climatic change in the near future, these ecosystems will likely be impacted by changing hydrological processes of uncertain but potentially dramatic character.

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## References

Baker, V.R., Costa, J.E., 1987. Flood power. In: Mayer, L., Nash, D. (Eds.), Catastrophic Flooding. Binghamton Symposium in Geomorphology, vol. 18, Allen and Unwin, Boston, pp. 1–21.

Balling Jr., R.C., Meyer, G.A., Wells, S.G., 1992a. Climate change in Yellowstone National Park: is the drought-related risk of wildfires increasing? Climatic Change 22, 34–35.

Balling Jr., R.C., Meyer, G.A., Wells, S.G., 1992b. Relation of surface climate and burned area in Yellowstone National Park. Agricultural and Forest Meteorology 60, 285–293.

Benson, M.A., Dalrymple, T., 1967. General field and office procedures for indirect discharge measurements. Techniques of Water-Resources Investigations, U.S. Geological Survey, Book 3, Chap. A1, pp. 1–30.

Beven, K., 1981. The effect of ordering on the geomorphic effectiveness of hydrologic events. IAHS-AISH Publication 132, 510–526.

Bingham, M.K., Meyer, G.A., 1994. Flood history since 1800 in relation to late Holocene climate change and stream incision, Yellowstone National Park. Geological Society of America Abstracts with Programs 26 (6), 5.

Bull, W.B., 1988. Floods; degradation and aggradation. In: Baker, V.R., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. Wiley, New York, pp. 157–165.

Chadde, S.W., Kay, C.E., 1991. Tall-willow communities on Yellowstone's northern range: a test of the "natural-regulation" paradigm. In: Keiter, R.R., Boyce, M.S. (Eds.), The Greater Yellowstone Ecosystem: Redefining America's Wilderness Heritage. Yale Univ. Press, New Haven, CT, pp. 231–262.

Church, M., 1983. Patterns of instability in a wandering gravel bed channel. In: Collinson, J.D., Lewin, J. (Eds.), Modern and Ancient Fluvial Systems. International Association of Sedimentologists Special Publication, vol. 6, Blackwell Scientific, Oxford, UK, pp. 169–180.

Church, M., 1988. Floods in cold climates. In: Baker, V.R., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. Wiley, New York, pp. 205–229.

Costa, J.E., 1974a. Stratigraphic, morphologic and pedologic evidence of large floods in humid environments. Geology 2, 301–303.

Costa, J.E., 1974b. Response and recovery of a Piedmont watershed from Tropical Storm Agnes, June 1972. Water Resources Research 10, 106–112.

Costa, J.E., 1983. Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range. Geological Society of America Bulletin 94, 986–1004.

Costa, J.E., 1987. A comparison of the largest rainfall–runoff floods in the United States, with those of the People's Republic of China. Journal of Hydrology 96, 101–115.

Costa, J.E., O'Connor, J.E., 1995. Geomorphically effective floods. In: Costa, J.E., Miller, A.J., Potter, K.W., Wilcock, P.R. (Eds.), Natural and Anthropogenic Influences in Fluvial Geomorphology. American Geophysical Union, Geophysical Monograph, vol. 89, pp. 45–56, Washington, DC.

Crowley, T.J., 2000. Causes of climatic change over the past 1000 years. Science 289, 270–277.

Dirks, R.A., Martner, B.E., 1982. The Climate of Yellowstone and Grand Teton National Parks. USDI-National Park Service Occasional Paper 6, Mammoth, WY, 26 pp.

Douglas, A.V., Stockton, C.W., 1975. Long-term reconstruction of seasonal temperature and precipitation in the Yellowstone National Park region using dendroclimatic techniques. Report to U.S. National Park Service-Yellowstone, Laboratory of Tree-Ring Research, University of Arizona, Tucson, 86 pp.

Ely, L.L., Enzel, Y., Baker, V.R., Cayan, D., 1993. A 5000-year record of extreme floods and climate change in the southwestern United States. Science 262, 410–412.

Epstein, J.L., Meyer, G.A., 1997. Hydraulic aspects of the McLaren mine tailings dam-break flood on Soda Butte Creek, Yellowstone National Park. Geological Society of America Abstracts with Programs 29 (1), 43.

Everitt, B.L., 1995. Hydrologic factors in the regeneration of Fremont Cottonwood along the Fremont River, Utah. In: Costa, J.E., Miller, A.J., Potter, K.W., Wilcock, P.R. (Eds.), Natural and Anthropogenic Influences in Fluvial Geomorphology. American Geophysical Union, Geophysical Monograph, vol. 89, pp. 197–208, Washington, DC.

Glidden, R., 1999. McLaren tailings site background report. Beartooth Alliance McLaren Tailings Committee, unpublished report, Cooke City, MT, 19 pp.

Grant, G.E., Swanson, F.J., 1995. Morphology and processes of valley floors in mountain streams, western Cascades, Oregon. In: Costa, J.E., Miller, A.J., Potter, K.W., Wilcock, P.R. (Eds.), Natural and Anthropogenic Influences in Fluvial Geomorphology. American Geophysical Union, Geophysical Monograph, vol. 89, pp. 83–101, Washington, DC.

Grove, J.M., 1987. The Little Ice Age. Cambridge Univ. Press, Cambridge, UK, 498 pp.

Hack, J.T., Goodlett, J.C., 1960. Geomorphology and Forest Ecology of a Mountain Region in the Central Appalachians. U.S. Geological Survey Professional Paper 347, 66 pp.

Hadly, E.A., 1996. Influence of late-Holocene climate on northern Rocky Mountain mammals. Quaternary Research 46, 298–310.

Helley, E.J., LaMarche, V.C., Jr., 1973. Historic Flood Information for Northern California Streams from Geological and Botanical Evidence. U.S. Geological Survey Professional Paper 485-E, 16 pp.

Huff, D.E., Varley, J.D., 1999. Natural regulation in Yellowstone National Park's northern range. Ecological Applications 9, 17–29.

Humphreys, E.D., Dueker, K.G., Schutt, D.L., Smith, R.B., 2000. Beneath Yellowstone: evaluating plume and nonplume models using teleseismic images of the upper mantle. GSA Today 10 (12), 1–7.

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

Johnston, M.S., 1950. Pollution of Soda Butte Creek. Memorandum for the Chief Ranger, June 27, 1950, Yellowstone National Park archives, Mammoth, WY.

Klemeš, V., 1989. The improbable probabilities of extreme floods and droughts. In: Starosolszky, Ö., Melder, O.M. (Eds.), The Hydrology of Disasters. James and James, London, pp. 43–51.

Knox, J.C., 1993. Large increases in flood magnitude in response to modest changes in climate. Nature 361, 430–432.

Kochel, R.C., 1988. Geomorphic impact of large floods: review and new perspectives on magnitude and frequency. In: Baker, V.R., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. Wiley, New York, pp. 169–187.

Kreutz, K.J., Mayewski, P.A., Meeker, L.D., Twickler, M.S., Whitlow, S.I., Pittalwala, I.I., 1997. Bipolar changes in atmospheric circulation during the Little Ice Age. Science 277, 1294–1296.

Magilligan, F.J., 1992. Thresholds and the spatial variability of flood power during extreme floods. Geomorphology 5, 373–390.

Marcus, W.A., Meyer, G.A., Nimmo, D.R., 2001. Geomorphic control of persistent mine impacts in a Yellowstone Park stream and implications for the recovery of fluvial systems. Geology 29, 355–358.

McCarthy, D.P., Luckman, B.H., 1993. Estimating ecesis for tree-ring dating of moraines: a comparative study from the Canadian Cordillera. Arctic and Alpine Research 25, 63–68.

Meagher, M., Houston, D.B., 1998. Yellowstone and the Biology of Time. University of Oklahoma Press, Norman, 287 pp.

Meyer, G.A., 1993. A polluted flash flood and its consequences. Yellowstone Science 2 (1), 2–6.

Meyer, G.A. (Ed.), Late Pleistocene–Holocene Evolution of the Northeastern Yellowstone Landscape. Friends of the Pleistocene Rocky Mountain Cell Guidebook, Middlebury College, VT, 55 pp.

Meyer, G.A., Watt, P.M., 1999. Hydrological controls on the geomorphic and ecological impacts of a tailings dam-break flood on Soda Butte Creek, Montana and Yellowstone National Park. Geological Society of America Abstracts with Programs 31 (7), 253.

Meyer, G.A., Wells, S.G., Balling Jr., R.C., Jull, A.J.T., 1992. Response of alluvial systems to fire and climate change in Yellowstone National Park. Nature 357, 147–150.

Meyer, G.A., Wells, S.G., Jull, A.J.T., 1995. Fire and alluvial chronology in Yellowstone National Park: climatic and intrinsic controls on Holocene geomorphic processes. Geological Society of American Bulletin 107, 1211–1230.

Millspaugh, S.H., Whitlock, C., 1995. A 750-year fire history based on lake sediment records in central Yellowstone National Park. The Holocene 5, 283–292.

Moss, J.H., Kochel, R.C., 1978. Unexpected geomorphic effects of the Hurricane Agnes storm and flood, Conestoga drainage basin, south-eastern Pennsylvania. Journal of Geology 86, 1–11.

Mowry, A.D., 1998. Vegetation and channel changes on Soda Butte Creek, Yellowstone National Park. Unpublished Bachelors Thesis, Geology Department, Middlebury College, VT, 61 pp.

Nanson, G.C., 1986. Episodes of vertical accretion and catastrophic stripping: a model of disequilibrium flood-plain development. Geological Society of America Bulletin 97, 1467–1475.

Nanson, G.C., Croke, J.C., 1992. A genetic classification of floodplains. Geomorphology 4, 459–486.

National Park Service, 1918. Conditions for the month of June. Superintendent's Monthly Report, Yellowstone National Park archives, Mammoth, WY.

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

O'Hara, P.M., 1994. Morphometric analysis of flood potential and channel erosion in small drainage basins of northeastern Yellowstone National Park. Unpublished Bachelors Thesis, Geology Department, Middlebury College, VT, 46 pp.

Omang, R.J., Parrett, C., Hull, J.A., 1986. Methods for Estimating Magnitude and Frequency of Floods in Montana Based on Data Through 1983. USGS Water-Resources Investigations Report 86-4027, 85 pp.

Patton, P.C., Baker, V.R., 1976. Morphometry and floods in small drainage basins subject to diverse hydrogeomorphic controls. Water Resources Research 12, 941–952.

Pierce, K.L., 1979. History and Dynamics of Glaciation in the Northern Yellowstone National Park Area. U.S. Geological Survey Professional Paper 729-F, 90 pp.

Pierce, K.L., Morgan, L.A., 1992. The track of the Yellowstone hotspot: volcanism, faulting and uplift. In: Link, P.K., Kuntz, M.A., Platt, L.B. (Eds.), Regional Geology of Eastern Idaho and Western Wyoming. Geological Society of America Memoir, vol. 179, pp. 1–53.

Prostka, H.J., Ruppel, E.T., Christiansen, R.L., 1975. Geologic map of the Abiathar Peak Quadrangle, Yellowstone National Park, WY. U.S. Geological Survey Geological Quadrangle Map GQ-1244, scale 1:62,500, 1 sheet.

Reneau, S.L., Dietrich, W.E., 1987. The importance of hollows in debris-flow studies; Examples from Marin County, California.

In: Costa, J.E., Wieczorek, G.F. (Eds.), Debris Flows/Avalanches. Geological Society of America Reviews in Engineering Geology, vol. 7, pp. 165–180.

Riggs, H.C., 1976. A simplified slope-area method for estimating flood discharges in natural channels. Journal of Research of the U.S. Geological Survey 4 (3), 285–291.

Ritter, D.F., 1975. Stratigraphic implications of coarse-grained gravel deposited as overbank sediment, southern Illinois. Journal of Geology 83, 645–650.

Scott, M.L., Friedman, J.M., Auble, G.T., 1996. Fluvial process and the establishment of bottomland trees. Geomorphology 14, 327–339.

Sigafoos, R.S., Hendricks, E.L., 1969. The Time Interval Between Stabilization of Glacial Deposits and Establishment of Tree Seedlings. U.S. Geological Survey Professional Paper 650-B, pp. 89–93.

Simpson, E.D., Meyer, G.A., 1995. Changes in flood magnitudes and processes in northeastern Yellowstone Park. Geological Society of America Abstracts with Programs 27 (4), 55.

Singer, F.J., Mack, L.C., Cates, R.C., 1994. Ungulate herbivory of willows on Yellowstone's northern winter range. Journal of Range Management 47, 435–443.

Smith, R.B., Braile, L.W., 1993. Topographic signature, space–time evolution, and physical properties of the Yellowstone–Snake River Plain volcanic system: the Yellowstone hotspot. In: Snoke, A.W., Steidtmann, J.R., Roberts, S.M. (Eds.), Geology of Wyoming. Geological Survey of Wyoming Memoir, vol. 5, pp. 695–754.

Stoughton, J.A., Marcus, W.A., 2000. Persistent impacts of trace metals from mining on floodplain grass communities along Soda Butte Creek, Yellowstone National Park. Environmental Management 25, 305–320.

U.S.D.A. Weather Bureau, 1918. Climatological data for the United States by sections. Washington, DC, U.S. Department of Agriculture 5, (4).

U.S.D.C. Weather Bureau, 1950. Climatological data, Montana. Kansas City, U.S. Department of Commerce 53 (6).

U.S. Geological Survey, 1991. National Water Summary 1988–1989—Hydrologic Events and Floods and Droughts. U.S. Geological Survey Water-Supply Paper 2375, 591 pp.

Wagner, F.H., Foresta, R., Gill, R.B., McCullough, D.R., Pelton, M., Porter, W.F., Salwasser, H., 1995. Wildlife Policies in the U.S. National Parks. Island Press, Washington, DC, 242 pp.

Whitlock, C., Bartlein, P.J., 1993. Spatial variations of Holocene climatic change in the Yellowstone region. Quaternary Research 39, 231–238.

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

Wolman, M.G., Leopold, L.B., 1957. River Floodplains; Some Observations on Their Formation. U.S. Geological Survey Professional Paper 282-C, pp. 87–109.

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