Response and recovery of a subalpine stream following a catastrophic flood

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ABSTRACT

The July 15, 1982, Lawn Lake flood in Rocky Mountain National Park, Colorado, was caused by the failure of a 79-yr-old earthen dam. Peak discharges of the flood far exceeded naturally occurring flows, and it caused severe channel disturbance along most of Roaring River and some parts of Fall River. This study documents the geomorphic response of a 5-km reach of Fall River in the 5 yr following the flood. In 1983, the first year after the Lawn Lake flood, snowmelt flows were well above average. These high flows together with very high sediment yields from Roaring River resulted in significant geomorphic changes on reaches of Fall River downstream. During the 1983 snowmelt runoff, \(-15.5 \times 10^6\) kg of bedload sediment was eroded from the upper parts of the study area. These loads were at least 1,000 times higher than before the Lawn Lake flood. Most of this sediment was then deposited in a highly sinuous reach of Fall River in the lower part of the study area. This reach had not been much affected by the Lawn Lake flood, but sedimentation during the period of high flow in 1983 completely filled in the channel, resulting in the formation of a continuous 2.3-km-long depositional zone. In 1984, sediment yield from Roaring River declined dramatically, and this trend continued for the next 3 yr. By 1987, the bed-load sediment yield in the upper reaches of Fall River was only about \(0.4 \times 10^6\) kg/yr. The decline in sediment loads resulted in progressive erosion and recovery of the original channel of Fall River in the depositional zone reach. Recovery in the upstream part of the depositional zone was complete by 1985. Recovery in the downstream part of the depositional zone took longer because of the continued supply of sediment and because the sediment was mobile less of the time. As of 1987, about 80% of the material initially stored in the sedimentation zone had been eroded. Bed-load sediment yields at a sampling site 1 km downstream of the terminus of the depositional zone ranged from \(5 \times 10^6\) to \(11 \times 10^6\) kg/yr, but showed no significant decline over the 5-yr study period. The average rate of bed-load transport through this reach was at least 100 times greater than before the Lawn Lake flood, but few discernible channel changes resulted from the higher loads.

INTRODUCTION

Considerable effort has been spent trying to predict how alluvial rivers will respond to sudden changes in discharge (Q) and sediment load (Qs). These changes may result from disturbances within a watershed caused, for example, by floods, landslides, deforestation, or dams. When the normal water- and sediment-discharge regimes of a river are altered, two questions immediately arise: (1) how will the channel adjust its morphology and (2) how rapidly will these adjustments take place? To answer these questions, we could take the approach of using a physically based model of channel evolution, but to do this, we would need rather specific information about Q or Qs and how they change with time. Our ability to model or measure these processes at watershed scales is still very limited, so for the present, we must rely on direct observation and empiricism to understand how channels change in response to sudden disturbances.

Of various watershed-scale disturbances, floods have received perhaps the most attention from geomorphologists. Indeed, the literature contains many examples and descriptions of how floods have modified stream channels (see Baker and others, 1988, or Beven and Carling, 1989). From this work, it seems clear that the largest floods (meaning “large” in terms of discharge per unit drainage area or in relation to the mean annual flood) occur in small drainage basins in arid and semiarid regions (Costa, 1987). It is less clear why some floods are more effective in modifying stream channels than others. Wolman and Gerson (1978) suggested that geomorphic effectiveness was determined largely by climate, but topography, lithology, and vegetation are important as well. They further suggested that geomorphic effectiveness could be evaluated in terms of the length of time required for a landform to recover its prior condition. What constitutes “recovery” in the case of rivers remains somewhat ambiguous, but examples might include the reconstruction of floodplains (Hack and Goodlett, 1960; Schumm and Lichty, 1963) or the restoration of pre-flood channel width (Osterkamp and Costa, 1987), hydraulic geometry (Lisle, 1982), bed elevation (Kelsey, 1980), and sediment loads (Newson, 1980). When viewed in this context, recovery essentially involves the re-establishment of a quasi-equilibrium channel in response to changes in discharge and sediment load.

In this paper, I describe how two streams in Rocky Mountain National Park recovered from a dam-break flood. The flood occurred on July 15, 1982, when an earthen dam on Lawn Lake failed, unleashing a torrent to the valleys of Roaring River and Fall River (Fig. 1). Prior to this flood, Roaring River and Fall River were typical of alpine-subalpine streams in the Colorado Front Range: they had stable channels and carried low sediment loads (Caine, 1986). The Lawn Lake flood radically altered this scenario by causing extensive erosion along the entire length of Roaring River from Lawn Lake to its confluence with Fall River (Fig. 1). Where the two rivers join, the flood deposited a large alluvial fan, burying ~1 km of Fall River. The flood continued at much lower velocity through a low-gradient valley known as Horseshoe Park (Fig. 1). This 4-km reach, which owes its name to the tortuous meanders of Fall River, was left virtually unchanged because the flood was dispersed over much of the valley.
floor and its power was greatly reduced. At the downstream end of Horseshoe Park there are a series of late Pleistocene terminal moraines beyond which the gradient of Fall River increases. Here, the flood again gathered power, causing intermittent erosion and deposition for the next 8 km. The flood swept through the town of Estes Park and was finally stopped when it reached Lake Estes (Fig. 1). The flood caused $31 million in damage, and three lives were lost (Jarrett and Costa, 1986).

When I first visited this area in March 1983, it was clear that the ensuing spring snowmelt would cause erosion on the Roaring River fan, and this would greatly increase the sediment load of Fall River. Presumably, a new course would be established across the fan, and the meandering reaches below would have to adjust to the increased sediment load. Thus, I began a 5-yr field study to monitor changes in the morphology and sediment load of Fall River. The accessibility of Horseshoe Park allowed me to document these changes in some detail. The purpose of this paper is to summarize the results of this work, with a specific focus on processes of channel adjustment in a mountain stream subject to varying sediment supply. The Lawn Lake flood was not spawned by a natural hydrologic event, of course, but the sudden disturbance of otherwise stable stream systems, for example, due to landslides or forest fires, is not so unusual in mountain areas. It is in this context that the Lawn Lake flood and subsequent channel recovery are of broader interest.

STUDY AREA

Fall River and its major tributary, Roaring River, drain mountainous terrain within Rocky Mountain National Park (Fig. 1). The area is underlain by Precambrian crystalline rocks, which are mantled in many places by glacial deposits. Soils are thin and weakly developed. A forest cover is found up to an elevation of ~3,500 m, above which tundra and bare rock are common. Average annual precipitation in the area varies between 500 and 1,000 mm (Jarrett and Costa, 1986), most of which falls as snow between the months of October and May. The bulk of runoff occurs in the 3-mo period of spring and summer snowmelt from May through July. It is during this high-flow period that most sediment transport and channel change normally occurs. A point to emphasize here is that, in alpine basins of the Colorado Front Range, peak discharges generated by snowmelt are not "large" in terms of discharge per unit drainage area (Jarrett, 1990) or in terms of the mean annual flood (Pitlick, 1988). Flood-frequency curves for streams in this area have pronounced negative skew, and they become asymptotic above a return period of ~100 yr. This suggests an upper limit to the magnitude of snowmelt-generated floods in alpine-subalpine basins of Colorado, which is simply the result of there being a limit to the amount of energy available in the diurnal snowmelt cycle. Typically, the discharge of a 100-yr flood on an alpine stream in this area is not even twice the discharge of the 2-yr flood (Table 1). The effect of this hydrologic regimen is that Fall River experiences a relatively narrow range of peak discharges, although in some years, moderate flows persist for several weeks.

Roaring River and Fall River both flow within broad, glaciated valleys, and their channels are largely detached from hillslopes. Roaring River flows from Lawn Lake down a hanging valley with an average gradient of about 0.10. Above its confluence with Fall River, Roaring River has a drainage area of 33 km² (Fig. 1). Fall River drains similar terrain, but flows down the valley carved by the main glacier in the area. The average gradient of Fall River in Horseshoe Park is 0.003. Through most of this area, Fall River flows in a highly sinuous channel (average sinuosity in lower reaches is 2.5) cut in what appear to be glacial lake sediments. The drainage area of Fall River where it exits Horseshoe Park (Fig. 1) is 90 km².

Details of the Lawn Lake flood are described thoroughly by Jarrett and Costa (1986) and Blair (1987). A peak discharge estimated by the slope-area method of about 500 m³/s occurred just downstream of Lawn Lake. Above the Roaring River alluvial fan, the peak discharge of the flood was estimated to be 340 m³/s; Jarrett and Costa (1986) suggested this was about 30 times the 500-yr flood for Roaring River. The peak discharge on Fall River near the outlet of Horseshoe Park (Fig. 1) was estimated to be 200 m³/s, which is about eight times the 500-yr flood for this stream (Table 1). Two aspects of the Lawn Lake flood are important in the context of this
TABLE I. ESTIMATES OF PEAK DISCHARGE ON FALL RIVER AND UNREGULATED STREAMS NEARBY

<table>
<thead>
<tr>
<th>Station</th>
<th>Fall River at Estes Park</th>
<th>Big Thompson River at Estes Park</th>
<th>S. St. Vrain Creek near Ward</th>
<th>Middle Boulder Creek at Nederland</th>
</tr>
</thead>
<tbody>
<tr>
<td>USGS no.</td>
<td>(673209)</td>
<td>(6733000)</td>
<td>(6722000)</td>
<td>(6725000)</td>
</tr>
<tr>
<td>Drainage area (km²)</td>
<td>103</td>
<td>35</td>
<td>37</td>
<td>94</td>
</tr>
<tr>
<td>Years of record</td>
<td>22</td>
<td>44</td>
<td>24</td>
<td>77</td>
</tr>
</tbody>
</table>

Return period (yr) | Discharge (m³/s)
---|---
1.1 | 3.9 | 17.6 | 4.2 | 9.1 |
1.0 | 6.3 | 26.3 | 6.2 | 11.7 |
2.0 | 8.0 | 30.0 | 7.0 | 13.0 |
5.0 | 12.0 | 35.9 | 8.9 | 16.4 |
10 | 14.5 | 43.4 | 9.9 | 18.0 |
25 | 17.4 | 47.9 | 10.4 | 19.6 |
50 | 19.5 | 50.6 | 11.4 | 20.6 |
100 | 21.4 | 52.9 | 11.8 | 21.4 |
200 | 22.2 | 54.8 | 12.2 | 22.0 |
500 | 25.5 | 56.9 | 12.6 | 22.7 |

*Nonrecording crest gauge.

SUBALPINE STREAM FOLLOWING CATASTROPHIC FLOOD

The characteristics of the Roaring River alluvial fan, as described by Jarrett and Costa (1986) and Blair (1987), are as follows: the fan covers an area of 0.25 km², and it has a maximum thickness of 14 m and a volume of about 280,000 m³. Sediment in the fan was derived primarily from a lateral moraine immediately upstream that contains 60% boulders, 13% cobbles, 14% pebbles, 3% gravel, 10% sand, and <1% silt and clay. The fan was deposited in three phases, with each being represented by a distinctive lobe. The proximal and medial lobes contain the coarsest debris. The distal lobe contains sediment grading from cobbles at the apex to medium sand at the terminus. The abundance of sand in the distal lobe of the alluvial fan was particularly significant because, under normal snowmelt flows, Roaring River and Fall River could carry this size sediment rapidly to the undisturbed reaches. The Lawn Lake flood also left a network of distributary channels on the fan, most of which were small in comparison to the reaches of Fall River that were unchanged by the flood. It was clear, therefore, that with the onset of the 1983 snowmelt, these channels would enlarge and become unstable. It was not clear, however, what effect this instability would have on the meandering reach of Fall River downstream.

Study. First, the flood caused extensive erosion and sedimentation along Roaring River (Fig. 2a). In steeper reaches, the flood cut a channel as much as 15 m deep, and the unstable cut banks along the channel began collapsing immediately after the flood passed. In lower gradient reaches, the flood deposited a series of discontinuous terraces and alluvial fans, the most spectacular of which was deposited at the mouth of Roaring River (Fig. 2b). Thus, erosion caused by the flood made available a tremendous amount of loose sediment. Because the flood occurred in mid-July, however, well after the peak of the 1982 snowmelt, most of this sediment remained at the base of the eroded cliff faces awaiting the next year's snowmelt. Second, prior to the flood, Fall River flowed within a highly sinuous channel bounded by well-vegetated banks and a cobble bed. The channel was thus quite resistant to erosion. Using data from cross sections surveyed in Horseshoe Park by Jarrett and Costa (1986) and my own field surveys, I estimate that the shear stress in the channel of Fall River during the Lawn Lake flood was less than two times the normal bankfull shear stress and probably not more than three times the critical shear stress, that is, the stress needed to initiate bed-material movement. Comparison of postflood observations with ground photographs and bed material samples taken in Horseshoe Park in 1975 by M. P. Mosley (1983, written communication) clearly indicated that, except for the area covered by the alluvial fan, the reach of Fall River in Horseshoe Park was not significantly modified by the Lawn Lake flood.

Figure 2. (a) Eroded channel of Roaring River, looking upstream. Channel is about 6 m wide, and cut bank in left of photo is about 5 m high. (b) Roaring River alluvial fan. Roaring River enters from the left and joins Fall River just downstream from the outlet of the lake. The lake formed when the drainage of Fall River was partially blocked by the alluvial fan.
METHODS

In March 1983, I began setting a network of 80 channel cross sections on Fall River between the mouth of Roaring River and the downstream end of Horseshoe Park (Fig. 3). These cross sections were surveyed at least once each year, and many were surveyed several times per year. Point-count samples of the bed material were repeated at selected cross sections.

Next, I established stream-gauging and sediment-sampling stations. The upstream station, FR-1, was established in late May 1983 in a straight reach about 400 m downstream of the alluvial fan (Fig. 3). At this locality, Fall River flows within a well-vegetated channel having a bankfull width of 8.5 m and a slope of 0.0032. The channel here was stable for the entire study period, except in June 1983, when bed-load transport rates were high and dunes were migrating over the bed. The downstream station, FR-2, was established in July 1983 in the very sinuous reach where Fall River exits Horseshoe Park (Fig. 3). At this point, the channel has a bankfull width of 7 m and a slope of 0.0015. This section remained stable except for minor fluctuations in bed elevation due to dunes. Flow and sediment discharge measurements were made at these sites almost daily during the annual 4- to 6-wk period of snowmelt runoff.

The majority of sediment-discharge measurements include samples of only bed load and not suspended load. The emphasis here was on sampling bed load because, after 1983, suspended sediment concentrations in Fall River were typically < 100 mg/l. Bed load was measured with a hand-held Helley-Smith sampler having a 7.6-cm nozzle. A bed-load sample includes all of the bed load measured at 14 to 18 points spaced equally across the channel. All bed-load samples were dried, weighed, and sieved. In all, more than 500 bed-load samples were taken in Horseshoe Park. Included as part of this data set are the location, date, time, discharge, mean flow velocity, flow depth, bed-load transport rate, and grain-size distribution for each bed-load measurement.¹

Finally, a water-stage recorder and staff gauge were installed on the U.S. Highway 34 bridge where it crosses Fall River (Fig. 3). Current-meter measurements were made often to establish discharge-gauge height rating curves for both sediment-sampling sites; these rating curves were adjusted periodically. Instantaneous peak discharges and mean daily discharges of Fall River in Horseshoe Park were determined from the strip-chart record and later compared with data from a crest gauge on Fall River near Estes Park (USGS station number 673250) and a continuously recording gauge on the Big Thompson River, to which Fall River is a tributary (USGS station number 673300). I considered the latter record the more useful one because it was

¹GSA Data Repository item 9316, four tables, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301.
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longer (Table 1) and because, in addition to peak flows, it includes a continuous record of daily flows. On the basis of these comparisons, I found that the discharge of Fall River in Horseshoe Park was consistently 29% of the discharge of the Big Thompson River in Estes Park. Not coincidentally, the drainage area of Fall River in Horseshoe Park is 29% of the drainage area of the Big Thompson River in Estes Park. I thus adopted the convention of taking 29% of the discharge of the Big Thompson River in Estes Park to get the discharge of Fall River in Horseshoe Park for the period prior to my study. From this proxy flow record, I then constructed flood-frequency and flow-duration curves for Fall River in Horseshoe Park, which allowed me to place the flows observed from 1983 to 1987 in a long-term context.

RESULTS

Hydrology

Snowmelt discharges on Fall River ranged from well above average in 1983 to well below average in 1987 (Fig. 4). Peak and mean daily discharges in 1983, the year immediately after the Lawn Lake flood, were higher than in any other year of this study (Fig. 4). The peak discharge in 1983 of 11.9 m³/s was equivalent to a 5-yr flood (Table 1). Perhaps as important was the fact that mean daily discharges of 7 m³/s (~1.5-yr flood) were exceeded for more than 3 wk (Fig. 4). The volume of runoff produced from June 14 to July 13, 1983, ranks as the highest 30-d mean discharge in 39 yr of record. Runoff throughout the region was high in 1983 because of heavy winter snowfalls associated with an El Nino event. The years 1984 through 1986 were more or less average in terms of peak discharges and total runoff volumes. In 1987, the peak discharge and total runoff were somewhat below average (Fig. 4). Fall River, like other streams draining alpine basins in the Front Range, is not likely to experience flows much greater than those observed in this 5-yr period. As will be shown, the process of geomorphic recovery on Fall River appears to become less sensitive with time to variations in the hydrologic regimen.

Channel Changes in Upper Horseshoe Park

The pronounced channel changes that occurred within Horseshoe Park owe their history in part to the particular sequencing of runoff events that followed the Lawn Lake flood. High flows in spring of 1983, together with the tremendous influx of sediment from Roaring River, immediately caused channels on the alluvial fan to widen and become unstable. Distributary channels on the fan frequently avulsed, and new courses were carved across the surface. The unconsolidated finer-grained flood deposits were continually reworked as a result of these avulsions, and the sand-size fraction of the fan sediment was transported rapidly downstream.

Sediment yields from Roaring River declined significantly in 1984. By the end of the 1984 snowmelt, most of the steep gully walls along Roaring River had stabilized, and much of the loose sediment near the channel had been removed. In addition, nearly all of the surface of the distal alluvial-fan lobe had been reworked, leaving a lag of gravel-size sediment. The channel of Fall River within the alluvial-fan reach remained stable from this time on.

It would be overstating the case to say that in the alluvial-fan reach Fall River has "recovered," because the newly formed channel is very different from the pre-flood channel. From 1984 on, however, channel changes in this reach were minor. Sediment supply from Roaring River remained low, but it did not cease altogether, and Fall River continued to transport small but measurable amounts of bed load out of the alluvial reach. This condition, where the channel is stable yet capable of transporting the load supplied from upstream, could be interpreted as a state of equilibrium. Indeed, many self-formed gravel-bed rivers appear to have mobile beds but stable banks (Andrews, 1984). Parker (1979) pointed out the paradox of this condition, but he resolved it by showing that the distribution of shear stress across the bed of a rectangular channel is not uniform. Parker (1979) found that the local boundary shear stress, as given by the depth-slope product, would always be greater in the center of the channel than near the banks. As a result, at bankfull flow, the shear stress (τr) in the center of the channel can be above the threshold for motion (τrγ) and sediment transport can occur, while near the banks, τr < τrγ, no sediment transport occurs, and the banks remain stable. The river thus adjusts its width (and its depth) to the point where it is just able to transport the supplied load at the bankfull discharge.

The channel of Fall River on the alluvial fan is entirely self-formed in sediment deposited by the Lawn Lake flood, and thus cross sections surveyed in this reach offer a test of Parker's theory (Fig. 5). At the time of the first survey (May 1983), the channel was 7 m wide, the reach slope (s) was 0.0068, and the median grain size of the bed material was 16 mm. By June 1985, the channel had widened to 11 m (Fig. 5), and D50 had coarsened to 32 mm, while the slope did not change. The critical shear stress (and hence critical flow depth, dγ) for 32-mm bed material particles can be estimated using the Shields parameter,

$$\tau_r^* = \frac{\gamma d_c \gamma_s}{(\gamma_0 - \gamma) d}$$ (1)

where γ and γs are the specific weight of water and sediment, respectively. Taking τr^* = 0.06 (the appropriate value for D50) and rearranging, the solution of equation 1 indicates that D50 begins to move at dγ = 0.5 m. This flow depth corresponds exactly to the observed bankfull stage (Fig. 5). At this stage, the discharge is 9 m³/s, which is equivalent to the mean annual flood. Thus, within 2 yr of the Lawn Lake flood, Fall River had formed a channel wherein the bed material just begins to move at the bankfull discharge.

Changes in Bed Material

The sequence of channel evolution described above was accompanied by distinct changes in the bed material and sediment load of Fall River. Prior to the Lawn Lake flood, the bed material of Fall River was gravel-sized, but the flood left deposits in the distal lobe of the alluvial fan (cross sections 17 and 18, Fig. 2) that were much finer. This sediment was easily moved, and not surprisingly, the bed load sampled at FR-1 at the peak of the 1983 snowmelt (6-13-83) had nearly the same grain size distribution as the surface deposits on the alluvial fan (Fig. 6a). The bed material coarsened over the next 2 yr to form a gravel armor (Fig. 6a), which could be moved by flows that occur several days per year. At these flows, movement of particles is probably sporadic and the armor remains largely intact, protecting the bulk of sediment underneath from being eroded. Still, some of the sediment below is eroded and transported as bed load. The bed load sampled at FR-1 in June 1985 had virtually the same size distribution as the subsurface bed material sampled on the alluvial fan in July 1985 (Fig. 6b). This again implies that, as early as 1985, Fall River was reaching a near-equilibrium state, because if the subsurface bed material was the primary source of bed load at FR-1, and the channel was not degrading, then there must be a balance between the load supplied from upstream and the load transported off the fan.
Figure 4. Mean daily discharges on Fall River for periods between May 1 and August 31 from 1983 through 1987. The single dark circle marks the instantaneous peak discharge for the year. The horizontal bars correspond to discharges of the 5- and 2-yr floods.
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Figure 5. Cross section of Fall River in the upstream part of the alluvial fan showing changes in morphology at three different times. The flow level indicated is that required to move the median grain size of the bed material as it existed in June 1987.

Sediment Transport, FR-1

The description of postflood geomorphic response of Fall River in upper Horseshoe Park concludes here with a discussion of trends in bed-load transport at the FR-1 sampling station. When plotted as a semicontinuous time series, these data provide a remarkably consistent picture of Fall River's response to the Lawn Lake flood (Fig. 7). Bed-load transport rates reached their peak in June 1983 (Fig. 7a) when flows were highest and erosion was occurring upstream. During this period, as much as 750 Mg (Mg = 10^3 kg) of bed load was transported per day at FR-1. From mid-June 1983 on, bed-load transport rates declined steadily (Fig. 7a). The overall trend of this decline is log-linear (note that the ordinate scale is logarithmic), implying an exponential decay in bed-load transport. By late June 1987, when the last measurements were made, Fall River was carrying only 3 Mg of bed load per day, which is nearly three orders of magnitude lower than the rates measured in 1983. Superimposed on the log-linear decline in bed-load transport are annual fluctuations corresponding to the rise and fall of the snowmelt hydrograph. A time series of average boundary shear stress at FR-1 shows a slight decreasing trend (Fig. 7b), but this trend is not commensurate with the overall decline in bed-load transport rates. On the basis of these data, it appears that the long-term decline in bed-load transport in the upper reaches of Fall River was governed mostly by sediment supply rather than by sediment transport capacity.

A similar time series of median grain size ($D_{50}$) shows a slight coarsening of the bed load (Fig. 7c). At the start of the 1983 snowmelt, the bed load at FR-1 was medium sand, derived primarily from the distal alluvial fan. Subsequently, the bed load at FR-1 coarsened, but still much of it was sand-sized. As discussed above, sediment stored beneath the armor layer in the alluvial-fan reach was the source for this bed load. The quasi-steady trend in $D_{50}$ of the bed load (Fig. 7c) suggests that there was continued movement of sediment in the armor layer, but as the declining trend in bed-load transport rates (Fig. 7a) indicates, decreasingly less of the bed was mobile.

Annual sediment yields of Fall River in upper Horseshoe Park were estimated by integrating the area under the bed-load time-sc--

Figure 6. Grain-size distributions of bed material in the distal reach of the alluvial fan and bed load at FR-1. (a) Comparison of bed material prior to the Lawn Lake flood with a sample obtained at cross section 17 at the start of the 1983 snowmelt, and bed load at FR-1 at the peak of snowmelt. The pre-flood grain-size distribution is based on samples taken in Horseshoe Park in 1975 by M. P. Mosley (1983, written commun.). (b) Comparison of surface and subsurface bed material sampled at cross section 18 in 1985, and bed load at FR-1.

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Figure 7. Trends in (a) bed-load transport rate, (b) average boundary shear stress, and (c) median grain size at the upstream sediment sampling site, FR-1. The time series includes only measurements made each year during the 60-d period from June 1 to July 30.

The alluvial fan the equivalent of 28,000 yr of bed load. These comparisons are of little significance, but they illustrate how it takes a truly extraordinary event like the Lawn Lake flood to push an alpine fluvial system much beyond its normally very stable condition.

Channel Changes in Lower Horseshoe Park

A complex pattern of deposition and erosion occurred in the reaches of Fall River downstream of the U.S. Highway 34 bridge. Recall that the Lawn Lake flood did not significantly disturb Fall River here, and when cross sections were first surveyed at the start of the 1983 field season, the channel existed in essentially the same condition as prior to the flood. That situation changed in 1983 as sediment from Roaring River and the alluvial fan was transported to the more sinuous reaches in lower Horseshoe Park. Apparently, the increase in roughness and decrease in slope associated with higher sinuosity reduced sediment transport capacity to the point where much of the load was deposited and the channel aggraded to the floodplain (Fig. 8a). Ultimately, this formed a continuous zone of deposition extending from approximately the highway bridge to a point 2,300 m downstream. The loss of channel capacity in the depositional zone forced much of the flow overbank in a network of anabranch channels. The depositional zone terminated abruptly in a single cut-off meander located 1,000 m above the FR-2 gauge (Fig. 8b). This cutoff appeared to act like a broad-crested weir that caused a backwater effect, slowing the movement of sediment through the cutoff and inducing deposition upstream.

Surveys of cross sections in the reach downstream of the highway bridge illustrate the rather dramatic changes that occurred here (Fig. 9). Cross section 46, which is located 900 m downstream of the bridge, was first surveyed on June 7, 1983, when the channel existed in essentially the same condition.

| TABLE 2. ANNUAL SEDIMENT YIELDS AT FALL RIVER GAUGING STATIONS |
|------------------------|-----------------|
| **Year**               | **FR-1**        | **FR-2**        |
| 1983                   | 15,500          | 9,800           |
| 1984                   | 5,000           | 6,900           |
| 1985                   | 3,500           | 3,500           |
| 1986                   | 1,500           | 8,700           |
| 1987                   | 400             | 5,900           |
| **Total**              | **26,400**      | **30,400**      |

Note: values are in megagrams (10^9 kg) per year.
*Too few measurements made to accurately determine sediment yield.*
Figure 8. September 1983 aerial photographs of Fall River: (a) the downstream end of the depositional zone showing selected cross sections; (b) the reach downstream of the depositional zone showing selected cross sections and the FR-2 sediment sampling station.
as before the Lawn Lake flood (Fig. 9a). By July 20, 1983, the channel had aggraded to the level of the floodplain (Fig. 9a). Subsequent measurements at this site revealed that aggradation was short lived. A survey in mid-July 1984 (Fig. 9a) showed that most of the sediment in the channel had been eroded, and Fall River had nearly recovered its pre-1983 morphology. From then on, channel changes in this reach were minor.

Cross sections in the downstream end of the depositional zone were not surveyed until 1984, but it was clear from looking at the extent of overbank deposits in the field and on 1983 aerial photographs (Fig. 8a) that deposition had been ubiquitous here as well. The 1984 survey of cross section 54 shows Fall River in its aggraded state (Fig. 9b). By 1986, the channel had begun to recover its original configuration, but some aggradation was still evident in the survey of June 1987 (Fig. 9b). Thus, in contrast to the upstream reaches, the downstream reaches of Fall River had not fully recovered after 5 yr from the high sediment loads of 1983.

Recovery was most rapid in the upstream part of the depositional zone because of the drop in sediment loads from the alluvial-fan reach, and because sediment in the channel could be easily transported. The bed material in this reach in July 1983 was 80% sand and 20% gravel, almost identical to the bed load measured at FR-1 in June 1983 (Fig. 10a). Calculation of the critical shear stress using Equation 1 and specific values for this reach of \( s = 0.0035 \) and \( D_{10} = 2.0 \text{ mm} \) indicates that sediment at cross section 46 could be moved by a flow 6 cm deep. A discharge of this sort would be exceeded about 50% of the time. By July 1984, most of the sand was removed, and a gravel armor with a size distribution approaching that of the pre-1983 bed material had formed (Fig. 10a). Armor layers were slower to form in the lower part of the depositional zone. Once in place, the armor lay-

Figure 10. Grain-size distributions of bed material in the depositional zone. (a) Comparison of bed material prior to the Lawn Lake flood (M. P. Mosley, 1983, written commun.), bed load at FR-1 at the peak of the 1983 snowmelt, and bed material at cross section 43 in July 1983, July 1984, and May 1987. (b) Comparison of bed material prior to the Lawn Lake flood, and surface bed material at cross section 54 in July 1983, September 1985, and May 1987.
ers slowed erosion considerably, but not to the point where it ceased altogether. The bed material at cross section 54 on May 22, 1987, was predominantly gravel (Fig. 10b), which could be moved under moderate flows. Calculation of the critical shear stress using Equation 1 and specific values of $n = 0.0025$ and $D_0 = 10$ mm, indicates the bed material at cross section 54 would move at a depth of 0.4 m or a discharge corresponding to 4.5 m³/s. A flow of this magnitude would be exceeded about 18 dyr, which is often enough to cause persistent erosion.

The enigma of the depositional zone was that it terminated so abruptly that aggradation was never very significant in highly sinuous reach immediately downstream (Fig. 8b). After the experience of the 1983 snowmelt, I expected that in 1984 sediment would move en masse into this reach. This proved not to be the case. Repeated surveys of cross sections in this reach showed that there was a small amount of aggradation in 1984 (Fig. 11), but it was clear that sediment did not move through as a wave. The same conditions prevailed through 1987. Obviously, channel capacity in this reach was sufficient to convey most of the supplied load, which was modulated upstream by the backwater effect of the cut-off meander.

To summarize these data, changes in bed elevation at cross sections from the upper end of the depositional zone to FR-2 were calculated for different time periods (Fig. 12). Changes in bed elevation are with reference to a common datum defined by the average bed elevation in June 1983; points above the dashed line indicate aggradation, and points below the line indicate degradation. Bankfull elevations at each cross section are also shown. Figure 12 shows that by July 1983, the entire length of channel from cross section 40 to 57 had aggraded by about 1.0 m (Fig. 12). The "nose" of the depositional zone is clearly seen between cross sections 57 and 60; in this area, about 0.5 m of aggradation is evident. From cross section 60 to FR-1, there was only ~0.3 m of aggradation, and there was little loss of channel capacity (Fig. 12). By the end of the 1983 snowmelt, about 25,000 m³ of sediment was stored in lower Horseshoe Park, most of it in the reach between the highway bridge and the cut-off meander. At a bulk density of 1.5 Mg/m³, this volume equates to a mass of about $37.5 \times 10^8$ Mg, which is more than twice the bed load measured at FR-1 in 1983 ($15.5 \times 10^8$ Mg). Assuming my estimates of volume and bulk density are not far off, the discrepancy in sediment yield either reflects the fact that some of the sediment in the depositional zone was transported as suspended load, which was not sampled systematically at FR-1, or that the 1983 series of bed-load measurements at FR-1 is incomplete.

By 1984, much sediment had been eroded from the upstream end of the depositional zone, but aggradation is still evident in the downstream reaches (Fig. 12). By 1987, only...
3,000 m$^3$ of sediment (~12% of the initial volume) remained stored within this reach, and the profile of average bed elevations followed a relatively smooth but slightly over-steepened trend (Fig. 12).

Sediment Transport, FR-2

The cut-off meander that formed the downstream terminus of the depositional zone slowed sediment movement, but did not stop it altogether. A time series of bed-load transport at the FR-2 gauge shows that sediment transport in the sinuous reaches in lower Horseshoe Park was relatively steady (Fig. 13a). Each year there are rises and falls in bed-load transport related to snowmelt, but overall, the trends in unit bed-load transport rate (Fig. 13a), boundary shear stress (Fig. 13b), and median grain size (Fig. 13c) are consistent from year to year. Integrating the area under the bed-load transport curve, I estimate that over 4 yr Fall River transported \( \sim 30 \times 10^3 \) Mg of bed load past the FR-2 gauge. This number exceeds the 5-yr total at FR-1 by ~15%, which isn't likely considering some sediment remains in storage between the two sites. Again, the discrepancy is probably in the estimate of bed-load sediment yield at FR-1 in 1983, which I suggested above was perhaps 100% low. Assuming a bulk density of 1.5 Mg/m$^3$, the 4-yr total of bed-load sediment yield at FR-2 is equivalent to a volume of 20,000 m$^3$, which is 80% of the volume initially stored in the depositional zone. From cross-section measurements, I estimated that 88% of the sediment had been removed from the depositional zone, so these two estimates of sediment yield are in close agreement with each other. In any event, transport rates at FR-2 were never as high as they were at FR-1 in 1983, although moderate rates persisted for a longer time. When compared to data from the upstream site, the time series of bed-load measurements at FR-2 indicates little in the way of recovery. Together these two time series illustrate how the processes of recovery may be temporally disjointed over relatively short distances.

**DISCUSSION**

Although the Lawn Lake flood was not a natural event, the results of this study provide the basis for a general discussion of floods and their impacts on streams. Questions of scale and transferability will undoubtedly come to mind, but there are both parallels and contrasts between my study and other studies of the fluvial response to large floods.

**Geomorphic Effects of Large Floods**

The amount of erosion and deposition caused directly by the Lawn Lake flood was strongly influenced by the local topography, most of which is a relic of the last glaciation. Steeper reaches of Roaring River were scoured to bedrock, gentler reaches were sites of deposition, and wide, low-gradient reaches of Fall River were affected little. Abrupt changes in gradient and valley width are common in mountain streams, regardless of whether or not the area has been glaciated, and this makes it difficult to generalize about the role of catastrophic floods in small drainage basins. Baker and Costa (1987) showed that floods achieve the greatest power (the product of shear stress and velocity) in basins of 10–50 km$^2$, but it does not necessarily follow that floods on small, high-gradient streams are geomorphically the most effective. Small streams typically have coarse bed materials and banks that are difficult to erode. If thresholds for bed or bank erosion are greatly exceeded, however, floods can have catastrophic impacts, as witnessed on Cimarron River (Schumm and Lichy, 1963), Coffee Creek (Stewart and LaMarche, 1967), Plum Creek (Osterkamp and Costa, 1987), and Roaring River (this study). On the other hand, if sediment transport thresholds are only slightly exceeded, erosion and deposition by floods tend to be spotty, as observed on streams in Connecticut (Wolman and Eiler, 1956), in Wales (Newson, 1980), and on Fall River in Horseshoe Park (this study). In a study of the geomorphic effects of historic floods in basins of 250–2,500 km$^2$ in the central Appalachians, Miller (1990) found that stream power was a poor predictor of geomorphic change. Miller (1990) concluded that local site characteristics, such as channel alignment or valley width, have as much to do with morphologic change as does the discharge or power of the flood. The observation
that the Lawn Lake flood had such different geomorphic effects on Roaring River and Fall River supports this conclusion.

Role of Sediment Supply and Hydrology

It is common for sediment loads to initially increase as the result of disturbance caused by a flood, and then to decline as hillslopes and channels become more stable. The rate of recovery thus depends on the rate of change in sediment supply. If the channel continues to be overloaded with coarse sediment, widening will persist (Schumm, 1969); if the channel is starved of sediment, narrowing will result (Williams and Wolman, 1984). The time it takes for a new equilibrium to be established appears to be quite variable, but it clearly depends on the rate that processes within the watershed continue to supply sediment, and on the types of flows that can move sediment.

Both of these factors played an important role in Fall River’s response to the Lawn Lake flood. Fall River was quick to re-establish a stable channel on the alluvial fan largely because sediment yield of Roaring River decreased rapidly. Most of the sediment in the Roaring River valley is of glacial origin, and it could not be moved by normal snowmelt flows. The inherent stability of this watershed perhaps more than anything dictated the pattern of recovery on Fall River. This suggests that watershed factors, such as relief, geology, vegetation, or land use, are as important in governing rates of recovery as climate because these factors control sediment supply.

Hydrology still plays an important role in the recovery process. The snowmelt runoff regime of the Colorado Front Range is characterized by moderate flows that persist for some time. Under this hydrologic regime, flows greater than two times the mean annual flood are improbable, and it is unlikely that erosion on Roaring River or the alluvial fan will be re-initiated. This does not mean that morphologic adjustments on Fall River will cease altogether, because there will continue to be small but measurable amounts of sediment transport.

Spatial and Temporal Discontinuity

The different reaches of Fall River were never in the same phase of adjustment. During the period when erosion was occurring in the alluvial-fan reach, deposition was occurring in lower Horseshoe Park. Subsequently, the same upstream to downstream progressive erosion was observed in the depositional-zone reach. Downstream-progressive trends in erosion and sediment transport appear to be common in rivers overloaded with coarse sediment. Gilbert (1917) was the first to identify such trends, but similar patterns of sediment movement were observed on rivers in northern California following a large flood in 1964 (Kelsey, 1980; Madej, 1992). I would not characterize sediment transport on Fall River as “wave-like,” but my observations of erosion and deposition are not unlike what has occurred on braided rivers in New Zealand (Griffiths, 1979; Beschta, 1983) and New Guinea (Pickup and others, 1983).

Another trend revealed by the data from FR-1 is one of an exponential decline in sediment loads. Similar trends have been reported in studies of rill erosion on hillslopes near Mount St. Helens (Collins and Dunne, 1986), bed-load sediment transport following floods in Wales (Newson, 1980), and changes in channel-sorted sediment following timber harvesting in low-order basins (Pitlick, 1992). That the upper reaches of Fall River returned to near-background sediment loads within 5 yr of the Lawn Lake flood is remarkable considering the magnitude of the event. Contrast this with the bed-load data from FR-2, which showed little in the way of a trend toward recovery, and the continuity in time is lost over a very short distance of channel.

CONCLUSIONS

The Lawn Lake flood attained peak discharges far outside the range of naturally occurring flows. Channel changes caused directly by this flood, however, were quite variable. Intermittent erosion and deposition occurred along the entire length of Roaring River from Lawn Lake to its confluence with Fall River, deposition of a massive alluvial fan occurred where the two rivers join, and virtually no channel changes occurred in the 4-km reach of Fall River in Horseshoe Park. The variable patterns of erosion and deposition caused by the flood were due to changes in shear stress and sediment-transport capacity associated with variations in average gradient and valley width.

Geomorphic changes on Fall River under subsequent snowmelt discharges provide a sharp contrast between the direct and indirect effects of a catastrophic flood. In 1983, the first year after the Lawn Lake flood, $\sim 15.5 \times 10^3$ Mg of bed load was transported from the upper part of the study area; this represents the equivalent of perhaps 300 yr of sediment yield from an undisturbed watershed. From 1984 on, sediment supply from Roaring River decreased, and the channel of Fall River on the alluvial fan stabilized. The trend toward rapid recovery in the upper part of the study area is represented also by the exponential decay in bed-load measurements at a site downstream of the alluvial fan. By 1987, the sediment yield of Fall River was $0.4 \times 10^3$ Mg/yr. This was only 3% of the amount of sediment carried in 1983, but probably close to pre-flood levels. The implication is that within 5 yr the upper reaches of Fall River had nearly recovered from the effects of a flood that in every sense was catastrophic. Considering the scale of the initial disturbance, this was unexpected, but it can be explained qualitatively by the inherent stability of the Roaring River and Fall River watersheds.

In the reach of Fall River in lower Horseshoe Park, the indirect effects of the Lawn Lake flood were far more significant than the direct effects. The flood produced few channel changes in this reach, but during a month- to two-month period of high flow and high sediment supply in 1983, the channel aggraded to the level of the floodplain. This produced a continuous 2,000-m-long zone of deposition. Aggradation in the upstream reaches of the depositional zone was short-lived, and Fall River quickly recovered its pre-existing channel. Recovery of channel morphology in the upstream reaches of the depositional zone was rapid because of the reduction in sediment yield from the alluvial-fan reach and because the sediment in the channel could be moved frequently. Recovery in the downstream reaches of the depositional zone was slower because there was continued supply of sediment from eroding reaches upstream and because there was time for gravel-armor layers to form on the stream bed. Sediment in these armor layers could still be moved by moderate flows, however, and by 1987, 80% to 90% of the material that had been eroded from the depositional zone. A 4-yr time series of bed-load measurements indicates that sediment eroded from the depositional zone was transported steadily through the sinuous reaches of Fall River in lower Horseshoe Park. Although much higher than natural sediment transport rates, the increase in sediment load within this reach caused little channel change, mostly because here Fall River is bounded by very well vegetated banks. The complex history of erosion and deposition in this gravel-bed river illustrates that floods in small drainage basins may produce highly variable geomorphic responses.
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