

Terrace aggradation during the 1978 flood on Powder River, Montana, USA

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Abstract

Flood processes no longer actively increase the planform area of terraces. Instead, lateral erosion decreases the area. However, infrequent extreme floods continue episodic aggradation of terrace surfaces. We quantify this type of evolution of terraces by an extreme flood in May 1978 on Powder River in southeastern Montana. Within an 89-km study reach of the river, we (1) determine a sediment budget for each geomorphic feature, (2) interpret the stratigraphy of the newly deposited sediment, and (3) discuss the essential role of vegetation in the depositional processes.

Peak flood discharge was about $930 \text{ m}^3 \text{ s}^{-1}$, which lasted about eight days. During this time, the flood transported 8.2 million tons of sediment into and 4.5 million tons out of the study reach. The masses of sediment transferred between features or eroded from one feature and redeposited on the same feature exceeded the mass transported out of the reach. The flood inundated the floodplain and some of the remnants of two terraces along the river. Lateral erosion decreased the planform area of the lower of the two terraces ($\sim 2.7 \text{ m}$ above the riverbed) by 3.2% and that of the higher terrace ($\sim 3.5 \text{ m}$ above the riverbed) by 4.1%. However, overbank aggradation, on average, raised the lower terrace by 0.16 m and the higher terrace by 0.063 m.

Vegetation controlled the type, thickness, and stratigraphy of the aggradation on terrace surfaces. Two characteristic overbank deposits were common: coarsening-upward sequences and lee dunes. Grass caused the deposition of the coarsening-upward sequences, which had 0.02 to 0.07 m of mud at the base, and in some cases, the deposits coarsened upwards to coarse sand on the top. Lee dunes, composed of fine and very fine sand, were deposited in the wake zone downstream from the trees. The characteristic morphology of the dunes can be used to estimate some flood variables such as suspended-sediment particle size, minimum depth, and critical shear velocity. Information about depositional processes during extreme floods is rare, and therefore, the results from this study aid in interpreting the record of terrace stratigraphy along other rivers.

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Keywords: Terrace; Aggradation; Terrace evolution; Coarsening-upward sequences; Extreme floods

1. Introduction

Fluvial terraces often are defined as abandoned floodplains that are no longer active. The concepts of abandonment and activity need clarification. Floodplains are active fluvial surfaces that are expanding laterally in their planform dimensions (Wolman and Leopold, 1957) as well as accreting in their vertical dimension (Nanson and Young, 1981; Nanson, 1986; Moody et al., 1999). Terraces are formed when the river ceases to increase the planform area of the floodplain by accretion and incises below the level of (i.e., abandons) the floodplain because of tectonic uplift, base-level lowering, or complex response

(Schumm and Lichty, 1965; Baker et al., 1988; Bull, 1990). These processes leave paired terrace surfaces, which are continuous and lie at discrete elevations above the river. The number of surfaces depends on the relative rate of lowering or uplift (Veldkamp and van Dijke, 2000). Because the total surface area between the edge of the valley is finite for a given unit distance downvalley, the continued planform accretion of the floodplain requires that the planform area of some remnants of the terraces must decrease with time. Thus, the river whittles away at the terrace scarp removing near-vertical slices of sediment to be replaced by the planform accretion of the adjacent floodplain.

While active planform accretion of a terrace surface by annual floods eventually stops, or even reverses, the vertical accretion or aggradation on a terrace continues. Thus, terrace surfaces are not truly abandoned. They continue to be revisited, but only infrequently by extreme floods (Hirschboeck, 1988; Klemeš,

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1989; Ely, 1997). Extreme floods sometimes result from an infrequent combination of conditions such as rain-on-snow (Harr, 1981; Church, 1988), or sometimes they are floods that are initiated by the lowering of some threshold (Schumm, 1979). This often produces a complex response (Schumm and Parker, 1973; Schumm, 1991) related to some natural or anthropogenic disturbance such as fire (Meyer et al., 1995; Moody and Martin, 2001), tectonic activity (Schumm et al., 2000), landslides (Shimazu and Oguchi, 1996), overgrazing (Baker et al., 1988; World Meteorological Organization, 1999; Hereford, 2002), hydraulic mining (Gilbert, 1917), or climate change (Hereford, 1984; Redmond, 2002). Extreme floods can cause net erosion (Nansen, 1986) or net deposition on the terrace surface.

Deposition of new sediment on terrace surfaces is the result of overbank flow processes, which are quantitatively less well understood than similar in-channel flow processes of sediment transport and deposition (Leopold et al., 1964; Dietrich and Smith, 1984; Smith and McLean, 1984; Nelson and Smith, 1989; Pizzuto, 1994). This is particularly true of overbank flows during extreme floods that inundate terraces. Even during annual floods, little time is available to adequately measure flow velocities over these vegetated surfaces. The effects of plants on flow depend on the non-dimensional vegetation density parameter $\alpha = hD_s/\lambda^2$ where h (m) is the water depth, D_s (m) is the average stem diameter, and λ (m) is the average stem spacing (Kean and Smith, 2004; Smith, 2004; Griffin et al., 2005; Smith, 2006). While few direct observations have been made of sediment transport processes during extreme floods, net depositional processes can be inferred from the spatial distributions and particle-size characteristics of sediment deposits. Sediment is deposited when the boundary shear velocity is less than the critical shear velocity for initiation of motion. Rates of deposition and mean sediment sizes are expected to decrease with increasing distance from the channel as transport energy decreases (Allen, 1985; Pizzuto, 1987). Flood deposits also are expected to fine upward (Allen, 1970), but coarsening-upward deposits have been documented after floods (Costa, 1974; Fisk, 1974; Knox, 1987; Iseya, 1989).

This paper does not focus on the formation of terraces, but rather on their evolution. We quantify the processes of continued aggradation of existing terraces by an extreme flood and extend earlier work by providing an evolutionary perspective on the aging processes of terrace development. We present (1) a sediment budget (mass of sediment eroded and deposited) for each alluvial surface, (2) the stratigraphy of the newly deposited sediment, and (3) measurements of the essential role of vegetation in the depositional processes. Information on processes during extreme floods is rare enough to be valuable in interpreting the geologic record of other floodplains. Aggradation from extreme floods, in some terraces, may constitute the majority of deposits in a stratigraphic record.

2. Background

2.1. River character

Powder River is a northward-flowing, meandering river with its headwaters in the Bighorn Mountains (elevation ~4000 m)

of NE Wyoming, USA. Much of the river flows through prairie and badland country of northeastern Wyoming and southeastern Montana (Northern Great Plains Province) with mean annual precipitation ranging from 300 to 400 mm y^{-1} (Hembree et al., 1952). Powder River drains an area of 34,706 km², and discharges an average of 12.7 m³ s⁻¹ into the Yellowstone River. It transports an average of 2 to 3 million metric tons per year of suspended-sediment at Moorhead, MT (Fig. 1), just north of the Wyoming–Montana state line (Moody and Meade, 1990; Moody et al., 1999).

2.2. Geomorphic features

The study reach on Powder River begins at the stream gage at Moorhead, MT, and ends at the stream gage at Broadus, MT (Fig. 1). The valley length is 58 km and valley slope, S_v , is 0.0017, whereas the river length is 89 km and the channel slope, S_c , is 0.0011. The valley is ~1 km wide at the upstream end and widens to ~2 km at the downstream end.

In our study of terrace aggradation, we focus on a sequence of three terraces along Powder River. These terraces appear to be similar to those first named, described, and mapped half a century ago by Leopold and Miller (1954) largely in the Powder River basin in Wyoming. Leopold and Miller felt that these three surfaces met the primary criteria for a terrace system, i.e., continuity of surfaces and paired surfaces with a tendency for uniform height. They named these terraces (Fig. 2), from oldest (highest) to youngest (lowest): Kaycee, Moorcroft, and Lightning, and we have used the same terminology for the terraces in our study reach. The Kaycee and Lightning were described as fill terraces, and the Moorcroft as a cut terrace on material similar to that underlying the Kaycee terrace. Initial terrace genesis was postulated by Leopold and Miller to be associated with a change in climate during the Holocene from a relatively warm and dry period (7500–4500 years before present, YBP) to a relatively colder and wetter period (4500 YBP–present).

The Kaycee terrace is now 10–20 m above the riverbed (the break in slope between the bank and the bed); the terrace surface has an age ranging from about 4000 to 1600 YBP (Leopold and Miller, 1954; Albanese, 1990), and slopes downward (0.04) from the badland hills toward the river. This terrace has no plains cottonwood trees (*Populus deltoides* Bartr. ex Marsh. ssp. *monilifera* (Ait.) Eckenwalder) but is vegetated with ponderosa pine (*Pinus ponderosa*) and juniper trees (*Juniperus* sp.) along the upper edges where the terrace grades into the badland hills. The remainder of the terrace surface is vegetated with scattered sagebrush (*Artemisia cana*) and grasses and some remnants are used as dry-land wheat fields or winter pasture for cattle. The fill beneath the surface is composed of the Kaycee Formation (with a modern soil on top), which is underlain by some combination of paleosol, Ucross Formation, or Arvada Formation. The paleosol, with a relatively high concentration of calcium carbonate, indicates the position of the water table in late Holocene time (Leopold and Miller, 1954).

Incision into the Kaycee terrace is thought to have begun about 1600–1000 YBP, and may have been the result of the lowering of the regional water table from the level of the

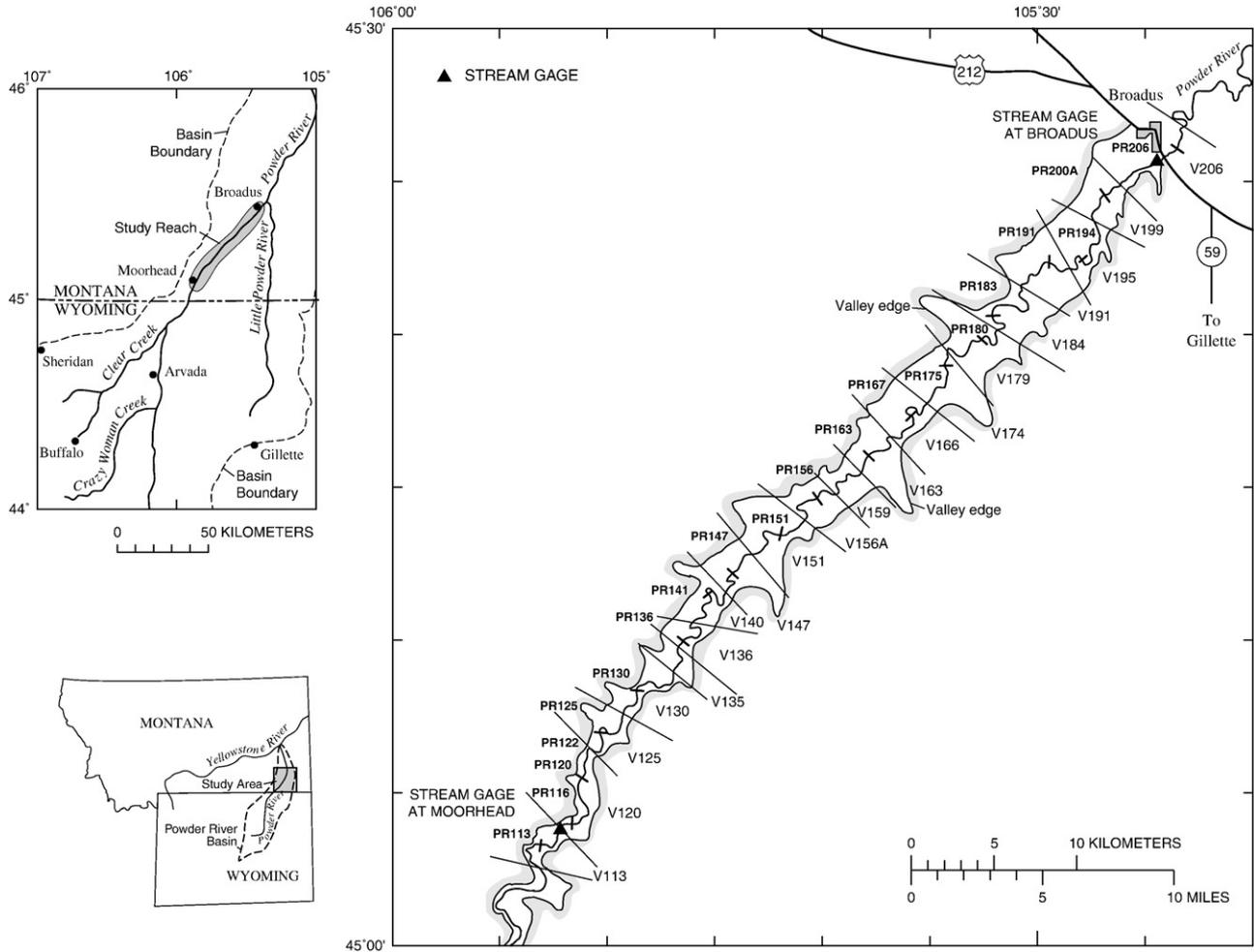


Fig. 1. Location of 19 channel cross sections and the valley reaches within the study reach along Powder River in southeastern Montana, USA. The wide, gray line indicates the valley edge, which is formed by bedrock bluffs. The short, thick lines indicate the channel cross sections and are identified by the prefix “PR”. The longer, thin lines are the boundary between valley reaches, and are identified by the prefix “V”.

paleosol down to the present level (Albanese, 1990). The rate at which the river level dropped relative to the Kaycee terrace was optimal to create the Moorcroft and Lightning terraces. The rate was neither too fast, such that no terraces had time to develop, nor too slow such that lateral erosion had sufficient time to remove all terraces that might have developed (Veldkamp and van Dijke, 2000). The Moorcroft terrace is now 3.1–4.2 m above the riverbed and has mostly widely spaced (100–500 m), large diameter (1–1.5 m) plains cottonwood trees.

Despite their need to have their roots below the top of the water table, plains cottonwood trees can persist, and even be newly established, on higher terrace levels above the floodplain. Extreme floods have been responsible for the establishment of some cottonwood trees on terrace surfaces as high as 1.2 to 3.5 m above the Missouri River. At these heights their survival is more likely than on the floodplain or on point bars subject to annual and ice jam floods (Scott et al., 1997). Seedlings can survive with their root tips as far as 1.0 m above the alluvial water table (Segelquist et al., 1993). Root growth can keep pace with the rate of decline of the alluvial water table, if the rate of decline is slower than $\sim 2.9 \text{ cm d}^{-1}$ (Segelquist et al., 1993), which often happens where ponding has been observed for an

entire growing season after an extreme flood (Jonathan Friedman, personal communication, 2007). Ponding is definitely possible on terraces with high silt content. Silt retards infiltration but also stores water for extended periods. Trees can be established on terraces, and the plains cottonwood trees along the Little Missouri River are 250 to 300 years old (Everitt, 1968; J. Friedman, personal communication, 2007), which may be much younger than the terrace surface on which they now grow. Many similar large trees along Powder River have been thinned on some terrace remnants so that alfalfa and other crops can be grown behind dikes used to contain irrigation water. On undeveloped parts of the terrace, sagebrush is dominant along with various grass species and the cottonwood trees.

The Lightning terrace is now 2.1–3.4 m above the riverbed and is rarely developed for agriculture because of its proximity to the river. It forms the banks in more locations than the Moorcroft terrace, and often has cottonwood trees (0.1–1.0 m diameter) in narrow clusters along old meander scars. The terrace is underlain by the Lightning Formation (fine or medium sand devoid of bedding with occasional lenses of fine gravel or sand; Leopold and Miller (1954) with no evidence of a soil profile. Leopold and Miller (1954) noted that in some reaches

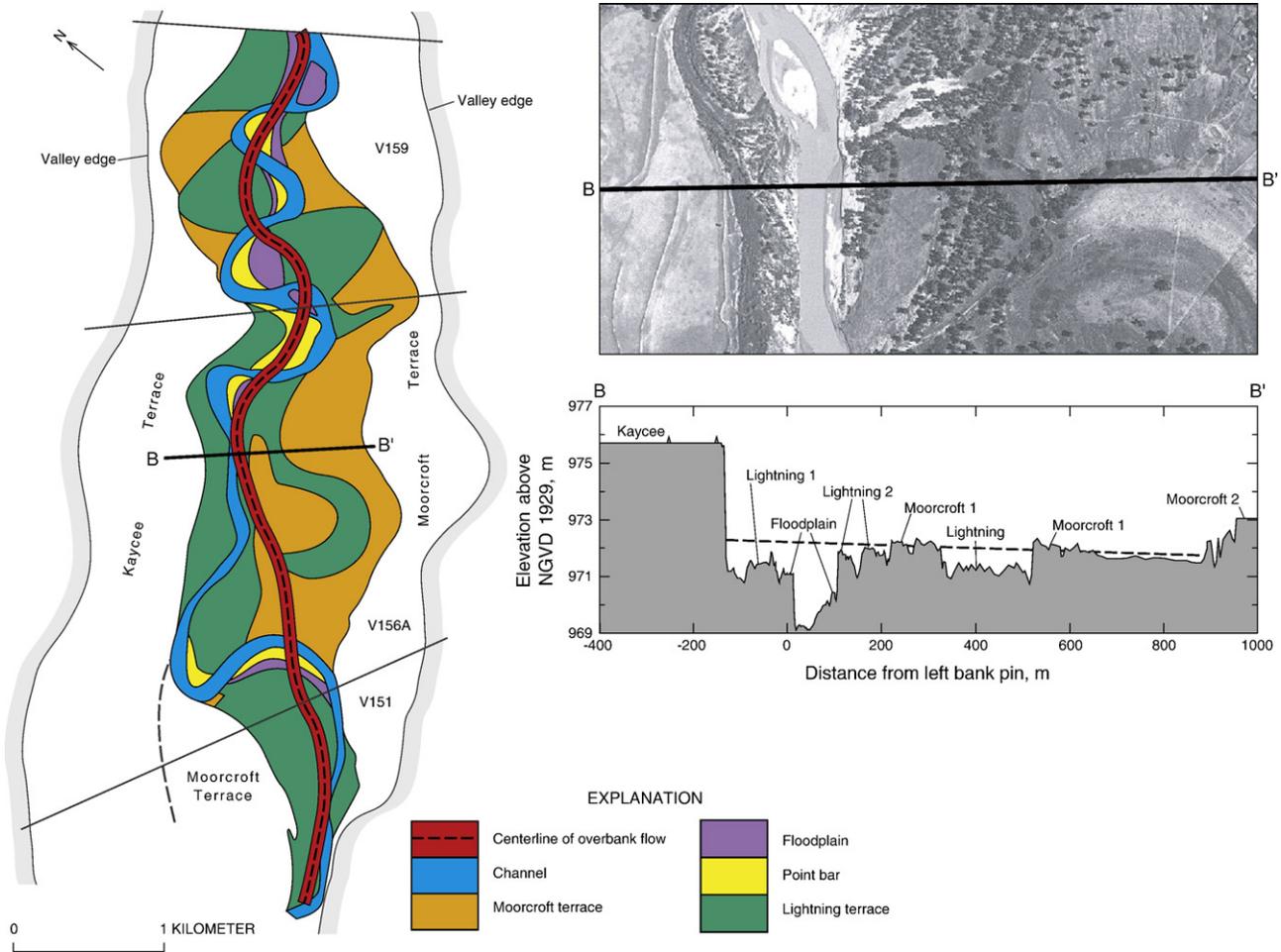


Fig. 2. Left: Geomorphic features in the valley of Powder River near PR156A. Flow is from the bottom to the top of the figure. Only those parts of each feature that were inundated by the 1978 flood are in color. Parts of the Moorcroft terrace and the all remnants of the Kaycee terrace that are shown as white were not inundated. The edge of the badlands bordering the valley is indicated by a wide, gray line. Solid black lines running across the inundated areas are the boundaries between valley reaches V151, V156A, and V159. Middle right: A swath of a 1978 post-flood aerial photo is centered over the line B–B' and shows the different vegetation communities. Bare sand areas are white, uniform dark gray areas are generally grass on floodplains, Lightning terrace, and on some Moorcroft terraces, and uniform light gray areas on the left bank with dikes are alfalfa fields on the Kaycee terrace. The cottonwood trees show as distinct circular areas of very dark gray. The diameter of the circular area gives some indication of the diameter of the tree. Lower right: A vertical profile along B to B' across the valley including the inundated terraces. The profile coincides with the channel cross section at PR156A, which was established after the 1978 flood and is about 30 m upstream from PR156. NGVD 1929 refers to the National Geodetic Vertical Datum of 1929. Ground surface elevations were determined by spirit-level survey conducted in 1996. The dashed line shows the inferred level of high-water during the 1978 flood. Terraces names were given by Leopold and Miller (1954).

“it probably is the modern floodplain, but generally there is present a slightly lower, narrow, and relatively inconspicuous flat bordering the stream, which is actually the present floodplain”.

The remaining channel banks are formed by floodplains inundated several times a decade or by point bars flooded each year. Floodplain segments are 1.0–1.7 m above the riverbed, vegetated with cottonwoods (0.01–0.1 m diameter) and with willows (*Salix exigua*, 0.01 m in diameter) interspersed with native and non-native grasses. Sedge grass (*Carex* sp.) forms a distinct vegetation band along the river edge near the break in slope between the bed and the bank and gradually disappears as the floodplain merges into a point bar. Upper parts of point bars have scattered young cottonwoods (0.01–0.1 m diameter), but willows (0.01 m in diameter) dominate the surface in curving rows associated with scroll bars. The lower parts of point bars are mostly unvegetated or sparsely vegetated sand and gravel.

2.3. Flood of May 1978

The flood of May 1978 happened at the beginning of our study, one year after 19 channel cross sections on Powder River were established, monumented, and measured. (Because we initially had insufficient data to establish the significance of this flood in the context of alluvial deposition, we continued to collect data annually for several decades). This extreme flood was caused by an anomalous circulation pattern (Parrett et al., 1984; Hirschboeck, 1988) rather than by annual spring snowmelt. A stationary front was fed by southerly, low-level flow that brought moisture from the Gulf of Mexico and resulted in an abnormal period of extended precipitation and runoff that included rain-on-snow at low elevations and snowmelt from higher elevations (Parrett et al., 1984). During 16–19 May 1978, intense rain (maximum 128 mm) fell on ground already saturated by previous rain and snow (3–8 May

1978, maximum 100 mm) over northeastern Wyoming and southeastern Montana (Parrett et al., 1984). The average precipitation at Broadus, MT during the month of May is 57 mm, but in 1978 it was 180 mm. At Moorhead, the daily discharge peaked at $779 \text{ m}^3 \text{ s}^{-1}$ on 20 May 1978 (Fig. 3A); and downstream at Broadus, this decreased to $711 \text{ m}^3 \text{ s}^{-1}$ on 21 May 1978 (U.S. Geological Survey, 1979). Daily peak discharge did not increase between Moorhead and Broadus because tributaries are few, small in size, and many are blocked by small reservoirs (Hubert Gay, personal communication, 1979; Riggs, 1985). The downriver decrease of $68 \text{ m}^3 \text{ s}^{-1}$ was attributed by Parrett et al. (1984) to channel and valley storage between Moorhead and Broadus. Record mean daily suspended-sediment concentrations were measured on 22 May 1978 at both Moorhead ($41,000 \text{ mg L}^{-1}$) and at Broadus ($22,600 \text{ mg L}^{-1}$) (Parrett et al., 1984).

3. Methods

The volumes of sediment eroded from and deposited on each geomorphic feature (Moorcroft terrace and Lightning terrace, floodplain, point bars, and channel) were determined within 19 separate valley reaches that included a channel cross section and

a valley transect. Valley transects started either at the left or right bank of the river, were approximately orthogonal to the downvalley direction, and extended across the inundated surface. The boundary between valley reaches was one-half the distance between adjacent valley transects (Fig. 1).

3.1. Erosion and deposition of sediment in the channel

Changes in sediment volume were monitored at 19 channel cross sections established in 1975 and 1977 and spaced about 5 km apart. Powder River cross sections are identified by the prefix “PR” followed by the number of river kilometers downstream from the mouth of Crazy Woman Creek, a tributary in Wyoming (Fig. 1). Elevations along a sections were measured before and after the 1978 flood (Moody and Meade, 1990; Moody et al., 2002). Areas of erosion and areas of deposition were normalized by the width of the channel to give an average thickness of erosion or deposition across the channel. Surface area of the channel was determined from a map of the channel in 1978 (Martinson and Meade, 1983). Volume of sediment eroded and deposited in the study reach was determined by multiplying the total channel surface area by the average change in thickness for all 19 cross sections. The error in the erosion and deposition volume was equal to the typical survey error (0.01 m) reported by Moody and Meade (1990) divided by the average erosion (0.18 m) and deposition (0.23) thickness in the channel (Table 1).

3.2. Erosion of geomorphic features

Volume of sediment eroded from the banks during the 1978 flood was computed as the height of the geomorphic feature times the surface area eroded. The height of each feature above the riverbed was equal to the average of all heights measured on both sides of the river at the 19 cross sections (Table 1). The surface area eroded was measured by using aerial photographs (1:10,000 scale) taken in 1976 before the flood and photographs taken in 1978 at a similar low river stage after the flood. The bank line after the 1978 flood was drawn on the 1976 photographs using primarily cottonwood trees, other large vegetation, and other distinct surficial landmarks. The line had a lateral uncertainty of 2–4 m on the terrace and floodplain surfaces. This line was more difficult to draw on the point bars where vegetation was small, the surface was less distinct, and much of the surface was undifferentiated sand and gravel. The length of the bank line and the surface area eroded between the bank lines in 1976 and 1978 was measured for each geomorphic feature in each valley reach. Computed volumes of sediment were summed to give a total eroded volume for each geomorphic feature. The measurement error within each valley reach was equal to the length of the 1976 bank line for each feature times the average estimated uncertainty of the bank line (3 m) times the height of the feature (Table 1). The total error (expressed as a percent) was the sum of the measurement errors for each valley reach divided by the total eroded volume (additional discussion of map errors is given by Martinson (1984)).

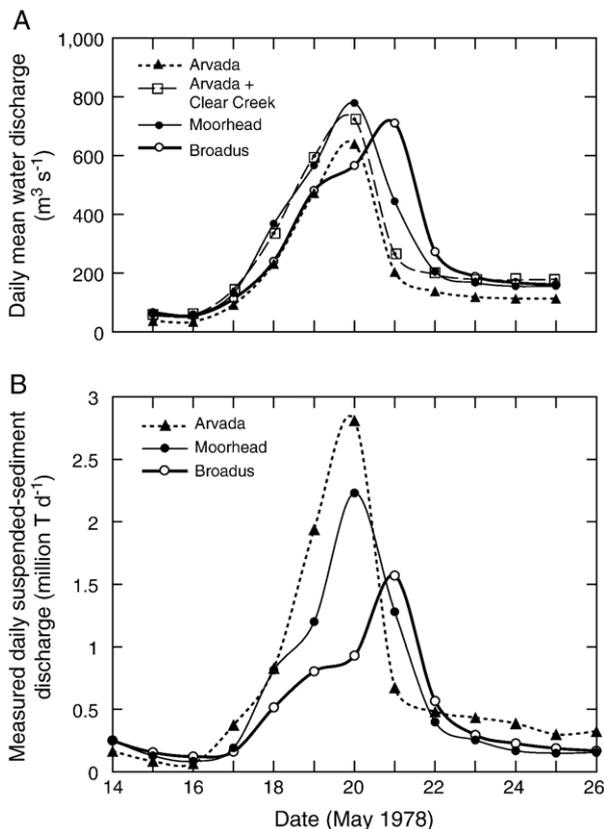


Fig. 3. (A) Daily mean water discharge. (B) Sediment discharges for the Powder River at Arvada, WY, Moorhead, MT, and Broadus, MT, during the flood of May 1978. The combined water discharge of Powder River at Arvada, WY (dotted line) plus Clear Creek near its confluence with Powder River between Arvada and Moorhead is shown as a dashed line in graph A. Concerning the measured sediment discharge at Moorhead, see the footnote to Table 1.

Table 1
Sediment budget for 1978 flood on Powder River^a

Geomorphic feature	Percent of flooded area	Average height above riverbed (m)	Erosion				Deposition					
			Planform area (km ²)	Sediment volume (m ³)	Sediment mass (million tons)	Estimated error (%)	Surface area (km ²)	Reach average thickness (m)	Sediment volume (m ³)	Sediment mass (million tons)	Estimated error (%)	
			Kaycee Terrace	0	15	0.01	216,000	0.32	16	0	0	0
Moorcroft Terrace	20.1	3.5	0.40	1,486,000	2.23	21	9.73	0.063	613,000	0.92	34	
Lightning Terrace	47.3	2.7	0.73	1,884,000	2.83	26	22.88	0.16	3,660,000	5.49	11	
Floodplain	9.6	1.4	0.18	226,000	0.34	35	4.65	0.17	790,000	1.18	24	
Pointbar 1 (5 valley transects)	5.5	0.5	0.04	21,500	0.03	115	2.67	0.29	774,000	1.16	22	
Pointbar 2 (8 cross sections)				635,000	0.95	13			2,023,000	3.03	18	
Channel	17.5	0.18	–	1,526,000	2.29	6	8.48	0.23	1,950,000	2.92	4	
Total erosion sources using Pointbar 1						8.0			Total depositional sinks		11.7	
Total erosion sources using Pointbar 2						9.0			Total depositional sinks		13.5	
<i>Suspended-sediment Load, million T</i>												
Input at Moorhead Gage (16–23 May 1978)						5.9 ^b	16		Output at Broadus Gage		4.5	16
Grand Total Sources using Pointbar 1						13.9			Grand Total Sinks		16.2	
Grand Total Sources using Pointbar 2						14.9			Grand Total Sinks		18.0	

Channel average erosion and deposition thickness was determined from 19 channel cross sections; bulk density was assumed to be 1500 kg m⁻³; Pointbar 1 and Pointbar 2 are two alternative methods of calculating erosion and deposition.

^a Total flood area in the 89 km long and the study reach was 48.4 km².

^b One of the conventions of the kind of portrayal shown in Fig. 4 is that the budget be forced to balance. To balance the excess of deposition over erosion as listed in the bottom lines of this table, an extra 2.3 million tons were added in the budget diagram (Fig. 4) to the measured input (5.9 million tons) at Moorhead. To justify this addition, we report here that five of the daily samples collected during the peak of the flood at Moorhead had to be collected unconventionally from the river surface at the edge of flow for safety reasons rather than from the centroid of the flow using a bridge-mounted depth-integrated sampler. Moorhead bridge was inaccessible during the peak days of the flood. As surface-water samples dipped from the edge of the flow are likely to contain substantially smaller concentrations of suspended-sediment than a fully depth-integrated sample from the centroid of flow, this is among the more likely sources of much of the disparity between the measured totals for erosion and deposition listed in the table.

3.3. Deposition on geomorphic features

Volume of sediment deposited on each geomorphic feature was equal to the surface area of the feature times the reach average thickness measured along a valley transect. Thicknesses of each new sediment layer were measured in small trenches dug, mostly during October 1978, at intervals of 5–20 m along each valley transect. Average depositional thickness, for each feature along a transect, was assumed to be representative of the total area occupied by the feature within each valley reach. Flood boundaries were determined at the ends of the valley transects, from interviews with ranchers along the river, and were plotted on USGS 1:24,000 topographic quadrangle maps. Volume of deposited sediment in each valley reach was summed to give a total depositional volume for each geomorphic feature. The error (expressed as a percent) in the deposition volume was equal to the standard error (%) of all the values of the reach average thickness for each geomorphic feature (Table 1).

A second method was used to calculate erosion and deposition on point bars. The first method (pointbar 1 in Table 1) used aerial photographs to estimate erosion; however, it

was difficult to determine the two bank lines around point bars and only deposition could be measured along the valley transects (only 5 of the 19 valley transects included point bars). The second method (Pointbar 2 in Table 1) used 8 of the 19 channel cross sections that included point bars. Change in cross-sectional area was calculated by differencing the elevations (interpolated to every 0.1 m) between topographic surveys made before and after the flood and accumulating the areas of erosion and deposition separately. Results from all eight cross sections were averaged and assumed to be representative of a 1-m-wide swath across any location on a point bar along Powder River. The measured water-line distance around each point bar in 1976 was multiplied by the average change in area of erosion and area of deposition to obtain a second estimate.

3.4. Particle-size distribution

Particle sizes of each newly deposited sediment layer found in the small trenches along the valley transects were determined in the field by comparison with a sediment reference card. Size classes were gravel (>2 mm), coarse sand (0.5–2 mm), medium sand (0.250–0.500 mm), fine sand (0.125–0.250 mm), very fine

sand (0.063–0.125 mm), silt (0.004–0.063 mm), and mud [mixture of silt and clay-sized (<0.004 mm) particles]. In addition, samples of selected size classes were collected in the field from different layers along each valley transect. Particle-size distributions for these samples and sand samples from other depositional features were measured in the laboratory using standard sieve analysis for sands and pipette-settling analysis for silt and clays (Guy, 1969).

3.5. Sediment budget

A sediment budget is based on the conservation of mass within a control volume. Boundaries of the control volume were defined by the stream gage and sediment-monitoring station at the upstream (Moorhead) and at the downstream (Broadus) ends of the study reach and by the area within the valley inundated by the 1978 flood. The difference between the input of sediment, I (metric tons, T), at the upstream end and the output of sediment, O (metric tons), at the downstream end must equal the change in sediment mass, ΔS (metric tons) within the study reach. This can be written as

$$I - O = \Delta S + e = (D - E) + e \quad (1)$$

where D (metric tons) represents deposition, E (metric tons) represents erosion, and e (metric tons) represent the error in the field measurements. Suspended-sediment input was measured daily by using a depth-integrated sample taken at a single vertical on the bridge at Moorhead, or from a surrogate sample collected as near as possible to the Moorhead bridge (see footnote to Table 1). Similarly, the output was measured daily (depth-integrated sample) at a single vertical on the bridge at Broadus. Data from these single verticals were calibrated with periodic measurement of suspended-sediment at a cross section with multiple verticals. The Moorhead station (U.S. Geological Survey number 06324500) has been operated continuously since it was established in 1929, and the Broadus station (U.S. Geological Survey number 06324710) was established in 1975 and operated until 1995. The combined empirical relation between suspended-sediment concentration, C (mg L^{-1}), and daily mean discharge, Q ($\text{m}^3 \text{s}^{-1}$), for Moorhead and Broadus is

$$C = 156Q \quad \text{for } Q < 273 \text{ m}^3 \text{ s}^{-1} \quad (2)$$

with $R^2=0.65$ and with a measurement variability at $100 \pm 20 \text{ m}^3 \text{ s}^{-1}$ of about $\pm 16\%$.

4. Results

4.1. 1978 flood

Most of the area, flooded by the 1978 flood, was Lightning and Moorcroft terraces (32.6 km^2 or 67.4% of the total flooded area, Table 1). The flooded planform area of the Lightning terrace (47.3%) was larger than the adjacent floodplain area (9.6%), which agrees with the description given above by Leopold and Miller (1954) that the floodplain is often “relatively inconspicuous”. Bankfull discharge, Q_{bank} ($\text{m}^3 \text{ s}^{-1}$), at Moorhead is about

$170 \text{ m}^3 \text{ s}^{-1}$; and hourly stage data showed that the water rose above bankfull on 18 May 1978 at about 1400 h and fell below bankfull on 23 May 1978 at about 0700 h. Peak discharge at Moorhead was $929 \text{ m}^3 \text{ s}^{-1}$ (hourly data) at 2100 h on 20 May 1978; and at Broadus, the peak discharge was $775 \text{ m}^3 \text{ s}^{-1}$ (hourly data) at 1430 h on 21 May 1978. The phase speed of the peak discharge was 1.4 m s^{-1} using the distance along the channel (89 km , blue area in Fig. 2), and 0.9 m s^{-1} using the distance along the valley (58 km , wide red line in Fig. 2).

The shape of the flood wave at Moorhead differed substantially from that measured at Broadus. Both hydrographs indicated a constant discharge ($\sim 57 \text{ m}^3 \text{ s}^{-1}$) on 16 May, but the rising discharge increased more rapidly at Moorhead than at Broadus. The discharge hydrograph at Broadus had a definite “shoulder” where the rate of increase in discharge slowed between 19 and 20 May (Fig. 3). The rate of decrease in the falling discharge was less at Moorhead than at Broadus, but by 23 May 1978 both hydrographs indicated nearly constant discharge ($\sim 160 \text{ m}^3 \text{ s}^{-1}$).

4.2. Sediment budget

The sediment budget was organized into sediment sources and sediment sinks. One source is the upstream input of suspended-sediment at Moorhead, and one sink is the downstream output at Broadus (Table 1; Fig. 4). Within the study reach, sediment erosion from the geomorphic features is considered as sources and sediment deposition on these features as sinks. Thus, the conservation of mass Eq. (1) can be rewritten with the sources on the left-hand side and the sinks on the right-hand side:

$$I + E = O + D + e \quad (3)$$

The erosion (E) and deposition (D) can be further separated into individual geomorphic features [Kaycee (K), Moorcroft (M), and Lightning (L) terraces, floodplains (F), point bars (P), and channel (C)] so that

$$E = E_K + E_M + E_L + E_F + E_P + E_C \quad (4a)$$

$$D = D_K + D_M + D_L + D_F + D_P + D_C \quad (4b)$$

Suspended-sediment transported into and out of the study reach was computed from the daily values of sediment discharge (tons d^{-1} or T d^{-1}) published annually by the U.S. Geological Survey (U.S. Geological Survey, 1979). The sediment discharge peaked on 20 May 1978 at Arvada and Moorhead, but a day later at Broadus (Fig. 3B). Plots of the sediment concentration as a function of water discharge showed a counterclockwise hysteresis at Moorhead and Broadus, but the magnitude of the hysteresis decreased for a given discharge. For example, at $500 \text{ m}^3 \text{ s}^{-1}$ the sediment concentration at Moorhead was 9000 mg L^{-1} but at Broadus it was 5000 mg L^{-1} .

Sediment load per day was summed over the 8-d duration of the flood when the flow was relatively unsteady from 16 May through 23 May 1978. Suspended-sediment transported into the study reach at Moorhead was measured at 8.2 million T (see

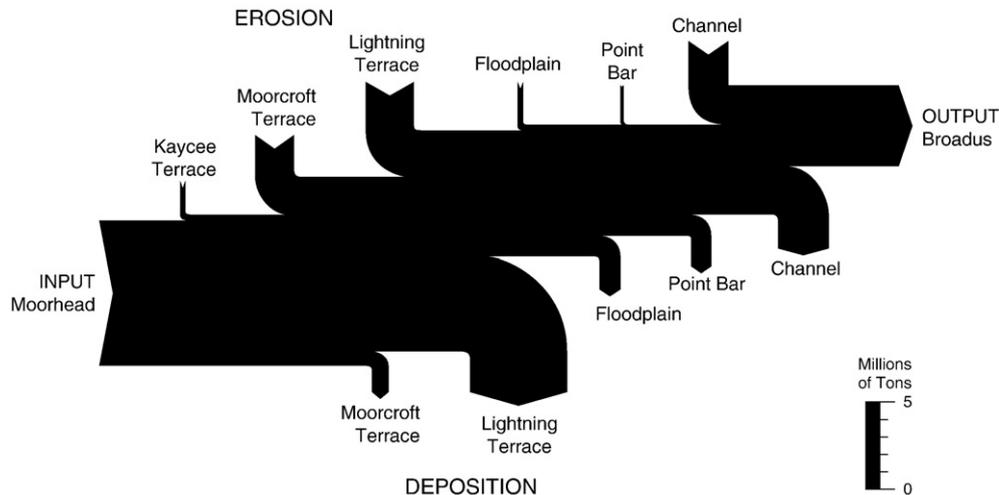


Fig. 4. Sediment budget (in metric tons) for the 58-km reach of Powder River valley between Moorhead and Broadus, MT, for the 8-d duration of the flood of May 1978. Data as listed in Table 1 (pointbar 1 option) and explained in footnotes.

footnote to Table 1) and 4.5 million T was transported out of the study reach at Broadus. Sediment was eroded from all geomorphic features within the study reach: the Lightning terrace (2.72 million T), Moorcroft terrace (2.16 million T), Kaycee terrace (0.32 million T), and channel (2.29 million T) (Table 1; Fig. 4). Sediment deposition within the study reach was primarily on the Lightning terrace (5.49 million T) (Fig. 5A), with lesser amounts on pointbars (1.16–3.03 million T) and on the bed of the channel (2.92 million T). The sources and sinks are shown in Fig. 4, and the principal results gained from an inspection of Fig. 4 are:

- (i) Erosion of all six features of the fluvial landscape (channel, point bars, floodplain, and three terraces) during the 1978 flood contributed substantially to the sediment budget of the 58-km valley reach. Aggradation was measured on all features except the Kaycee terrace.
- (ii) Sediment transferred between features or eroded from one feature and redeposited on the same feature exceeded the mass transported out of the study reach during the flood.
- (iii) The Lightning terrace gained twice as much new sediment through aggradation as it lost by bank erosion. In fact, the mass of newly deposited sediment atop the Lightning terrace, taken by itself, was slightly greater than the mass transported out of the study reach during the flood.
- (iv) Any individual sediment particle that was transported into the study reach during the flood is much more likely to have been deposited on one of the flooded surfaces in the study reach than to have been transported out of the study reach.

4.3. Overbank deposition

Particle sizes in the overbank deposition ranged from clay to gravel. Newly deposited gravel and coarse sand were found on the floodplains, primarily in old meander channels. Most of the newly deposited sediment on point bars adjacent to the channel

was very fine to coarse sand (61%) (Table 2). As the elevation of the geomorphic features increased, the percentage of sand decreased from 48% on the Lightning terrace (Fig. 5A) to 17%

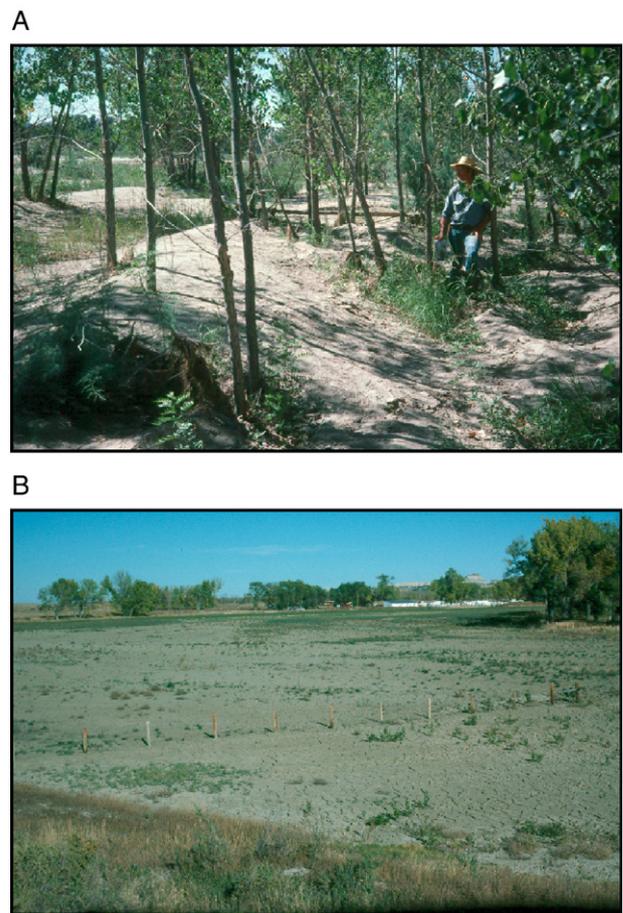


Fig. 5. (A) Photograph of the characteristic sand deposits on Lightning terrace, which were lee dunes downstream from vegetation stems, September 1978. View is upvalley. (B) Photograph of the characteristic mud deposits on the Moorcroft terrace, which coarsen upward, October 1978. Drying and cracking of the bottom mud layer can be seen at the surface.

Table 2
Particle-size distribution of sediment deposited on geomorphic features

Geomorphic feature	N	Percentage in each size class						
		Gravel	Coarse sand	Medium sand	Fine sand	Very fine sand	Silt	Clay
Size range in millimeters		>2.00	2.00–0.50	0.25–0.50	0.125–0.250	0.063–0.125	0.063–0.004	<0.004
Moorcroft Terrace	10	0.0	0.0	0.6	2.8	13	62	22
Lightning Terrace	20	0.0	0.0	1.3	11	36	40	12
Floodplain	7	8.8	1.7	3.3	25	29	22	10
Pointbar	4	0.0	1.0	4.1	20	36	36	2.9

Particle sizes were determined in the field using a sediment-size reference card; N is number of samples in the average.

on the Moorcroft terrace. More silt and mud were deposited on these two terraces than on the floodplains. Most of the new sediment deposited on the Moorcroft terrace was mud (Fig. 5B). Two depositional features dominated the aggradation of the terraces. First was the coarsening-upward sequences on both terraces, which was relatively uniform in spatial distribution. Second were lee dunes deposited around and downstream from trees and groups of trees and shrubs. Lee dunes depended on the distribution of the trees and the proximity to the channel, and were not uniform in spatial distribution.

Particle size coarsened upward in most of the newly deposited layers sampled along the valley transects. Average total thickness of these layers was 0.16 m on the Lightning terrace and 0.063 m on the Moorcroft terrace (Table 1) and consisted of 3–4 layers (mud, silt, very fine sand, and fine to medium sand). The bottom layer was usually mud with a relatively uniform thickness (0.01–0.04 m) regardless of the water depth (Figs. 6 and 7). The next higher layer was silt, but the thickness showed no discernible relation with water depth.

The third layer was very fine sand, which was more variable in thickness than the silt layer. Thickness of very fine sand seemed to have an inverse relation with water depth at some valley transects but this relation was not consistent along the entire study reach. Very fine sand, however, often had a maximum thickness landward of but not at, the channel bank where one would expect the greatest thickness. During other overbank floods on Powder River, we have observed that similar sand deposits were eroded by undercutting and sloughing at the channel bank as the flood stage dropped leaving the maximum thickness landward of the channel bank. The large variations in thickness of very fine sand along the V163 transect (Fig. 6) reflects the presence of scroll bars and vegetation. Fine and medium sand were generally thickest closer to the channel, but at a few scattered locations the coarsening-upward sediment deposits were capped by a layer of fine sand.

Lee dunes were composed of fine and very fine sand. Their characteristic shapes were typically an approximate ellipse with a perfoliate appearance relative to the tree in the planform

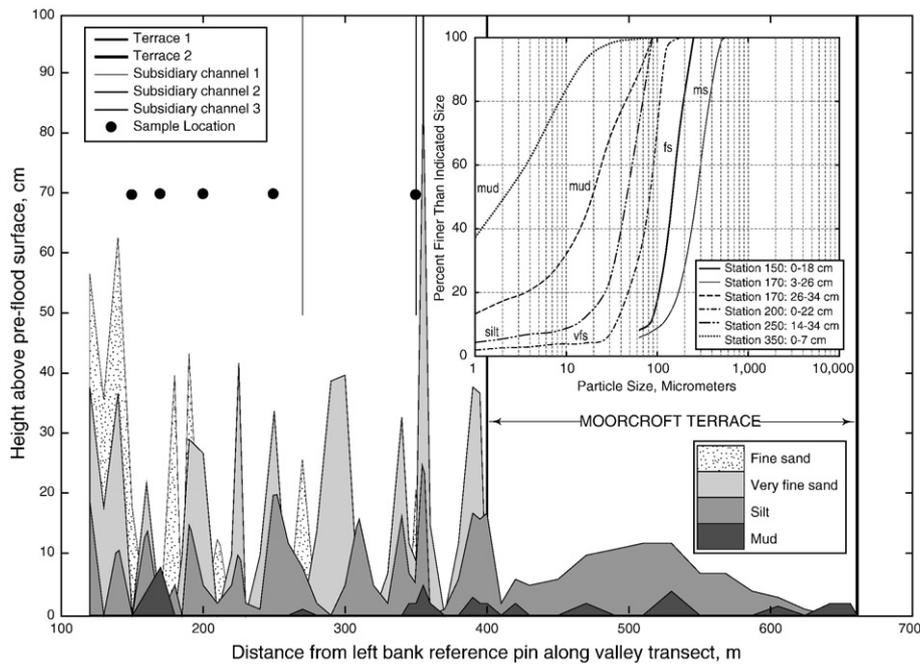


Fig. 6. Thickness and particle sizes of the sediment deposits on the terraces along valley transect V163. The solid black circles mark the horizontal locations where sediment samples were collected for particle-size analysis. The cumulative particle-size distribution is shown and the vertical location of these samples from within the total overbank deposit is listed on the insert graph. The pre-flood surface was not flat. The thin vertical lines at 270, 350, and 355 m indicate subsidiary channels. Solid vertical lines show the edges of the Moorcroft terrace. Abbreviations for particles sizes are very fine sand, vfs; fine sand, fs; medium sand, ms; and coarse sand, cs.

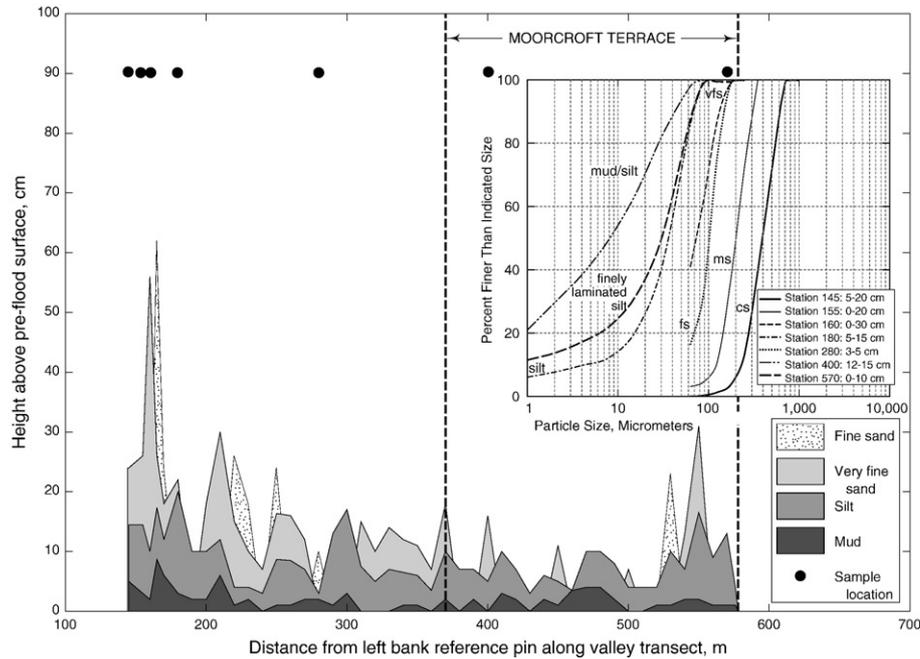


Fig. 7. Thickness and particle sizes of the sediment deposits on the terraces along valley transect V191. The solid black ovals mark the horizontal locations where sediment samples were collected for particle-size analysis. The cumulative particle-size distribution is shown, and the vertical location of these samples from within the total overbank deposit is listed on the insert graph. The pre-flood surface was not flat. Dashed vertical lines show the edges of the Moorcroft terrace. Abbreviations for particles sizes are very fine sand, vfs; fine sand, fs; and medium sand, ms.

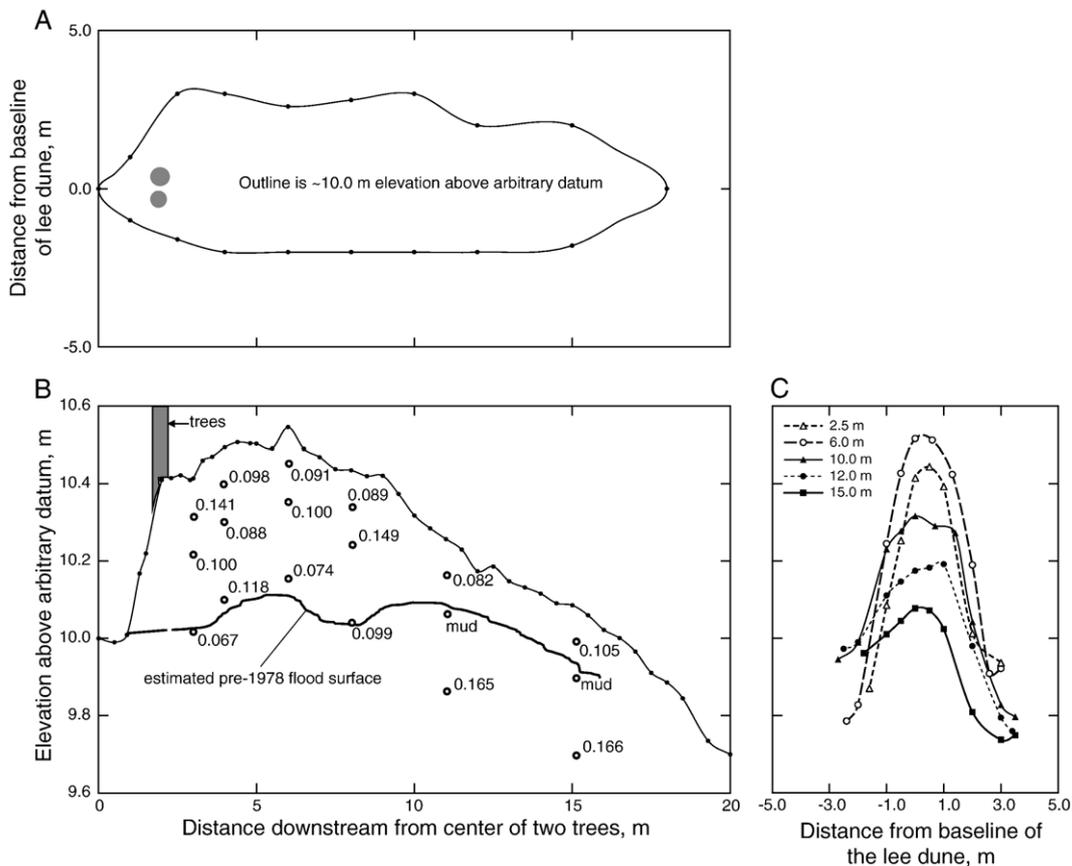


Fig. 8. Views a typical lee dune on the Lightning terrace (surveyed in October 2007). (A) Planform, (B) Downstream vertical profile with median particle size given in millimeters next to the sample location shown as an oval, and (C) Cross section viewed looking upstream.

outline, an airfoil in the downstream vertical profile, and parabola in cross section. For example, flow around two cottonwood stems (effective diameter of 1.2 m) eroded a depression at ~2 m upstream from the trees and deposited a sand within an elliptical area ~5-m-wide and ~18-m long (Fig. 8). Lee dunes closer to the channel were generally larger than those farther away from the channel.

5. Discussion

5.1. 1978 flood

The 1978 flood was an extreme flood and deviated from the frequency curve more than the next seven highest floods. While the recurrence interval was estimated to be ~50 years, the limited length of the discharge series makes extrapolation of recurrence intervals beyond 10–20 years quite questionable (Klemeš, 1989). It was not as high (by at least 2 m) as the earlier 1923 flood (~2800 m³ s⁻¹), the only other extreme flood recorded since Powder River valley was first settled by homesteaders in the late 1800s (Follansbee and Hodges, 1925). The 1923 flood plots as an outlier on the frequency curve for annual snowmelt floods (Charles Parrett, personal communication, 2007) and was caused by an unusual period of prolonged rain (160 mm over four days; Follansbee and Hodges, (1925)) at the end of September suggesting another anomalous circulation pattern (Hirschboeck, 1988). In general, climate variability creates flood variability (Redmond, 2002) and this may have caused past and recent extreme floods on Powder River as it did on the Upper Mississippi River during the late Holocene (Knox, 1987). The relative magnitudes of the 1978 flood (5 times $Q_{\text{bank}} \sim 170 \text{ m}^3 \text{ s}^{-1}$) and certainly the 1923 flood (16 times Q_{bank}) place both floods in the large flood population (>2–3.5 times Q_{bank}) defined by Knox (1993). Extreme floods tend to be associated with warmer periods having rapid snowmelt (Meyer, 2001) and greater potential for coincident and more intense rainfall (Karl and Knight 1998; Schneider 2004).

5.2. Terrace evolution

Extreme floods like the 1978 flood alter terraces in two important ways. First, extreme flood usually erode some remnants of all the terraces and this decreases the planform area. Second, extreme floods continue to aggrade the terrace surfaces with deposits of sediment whose thickness and particle size decrease as the terrace height increases. As terrace surfaces grow higher they are inundated less frequently because the overbank critical discharge becomes larger and the recurrence interval is longer (Wolman and Leopold, 1957; Moody et al., 1999; Moody and Troutman, 2000). Thus, terraces aggrade at a much slower rate than floodplains.

5.2.1. Planform erosion

Lateral erosion during the 1978 flood removed 0.73 km² from the planform area of the Lightning terrace or 3.2% of the flooded area, 0.40 km² from the Moorcroft terrace or

4.1% of the flooded area, and 0.01 km² from the Kaycee terrace, which was not flooded. More area was eroded from the Lightning terrace than from the Moorcroft or Kaycee terraces because the Lightning terrace forms more of the active channel banks than the other two terraces, or it is closer to the active channel and only protected in some reaches by narrow floodplains or point bars. Whereas the Moorcroft is set back from the channel in many reaches and protected from erosion by the Lightning terrace, floodplain, and point bar. Loss of planform area was most evident at the upstream side of a meander bend where the flood left the channel. It eroded the terrace creating a convoluted edge with scour holes, small eroded channels, and isolated remnants of the terrace surface. This loss of planform area by lateral erosion is irreversible. With enough time this process will eventually remove all remnants of the terraces and they will become extinct.

5.2.2. Terrace aggradation

The mass or volume of the sediment composing the terraces was increased by terrace aggradation but decreased by lateral erosion. This was especially true for the Moorcroft terrace. It gained 0.92 million T by aggradation representing a reach average increase in elevation of 0.063 m (Table 1), but lost more by lateral erosion (2.16 million T). On the other hand, the Lightning terrace gained 5.49 million T by aggradation representing a reach average increase in elevation of 0.16 m, but lost less by lateral erosion (2.72 million T). Thus, the net volume of the Moorcroft terrace within the study reach decreased, but the net volume of the Lightning terrace increased. On average, the elevation of the terrace surfaces are higher than before the 1978 flood.

Elevations of the present terrace surfaces reflect the cumulative aggradation on the original surface by numerous extreme floods, like the 1923 and 1978 floods. If one assumes that the original terrace surfaces had uniform height above the riverbed (this assumption is not essential but is easier to visualize), then deposition associated with these extreme floods would produce non-uniform aggradation for three reasons: (i) the different heights of the terraces, (ii) the different magnitude of each flood, and (iii) the different locations of remnants of the terraces relative to the active river channel.

For a simple river system with no meanders, or at a specific cross section, the thickness of aggradation on a terrace depends primarily on the particle settling velocity, the relative terrace height above the riverbed, and the magnitude of the flood (Moody and Troutman, 2000). The terrace height determines a minimum discharge required for inundation, and this minimum discharge continues to increase as aggradation by each successive flood increases the height of the terrace—just as in the case of a floodplain (Wolman and Leopold, 1957; Moody et al., 1999). For regular annual floods on Powder River, water discharge controls the suspended-sediment concentration and thus the net depositional thickness, in approximately linear fashion (Moody and Troutman, 2000). During extreme floods we expect the depositional thickness to continue to increase,

but not necessarily in a linear fashion. Thus, we can readily assume that extreme floods in the past have so differed in their discharges that aggradation on terrace surfaces has been non-uniform. Moreover, the magnitude of extreme floods has been shown to increase in response to modest changes in the climate during the late Holocene (Knox, 1993,) thus, the thickness of aggradation on terraces may increase, in the future, if current changes in climate are similar. However, the simple relation between thickness of aggradation and flood magnitude is altered when the local processes created by a meandering channel are introduced.

The local processes that cause aggradation on selected terrace remnants depend upon the spatial configuration of the terraces relative to the river channel. A general example is shown in Fig. 2 where some remnants of the Moorcroft terrace were inundated (orange color) and the elevation was increased, but other remnants were not inundated (white areas) and the elevation remained unchanged. Overbank sedimentation also differs depending upon the location inside versus outside of a bend (Howard, 1992). A specific example was a remnant of the Moorcroft terrace located on the right bank at a sharp bend in the river within the valley reach V135 (Fig. 1). Flood water left the channel at this bend and deposited an average thickness of 0.24 m of sediment on the right bank, which was four times the average for the entire study reach. The paired Moorcroft terrace on the left bank of the sharp bend was set back about 400 m and was protected from inundation by a point bar, floodplain, and Lightning terrace between the channel and the foot of the terrace. Localized river processes also can be aided or impeded by the ranchers who build dikes on the Moorcroft terrace to keep irrigation water in, but also kept flood water and sediment out.

Episodic aggradation during a single flood can raise the elevation of one surface to the elevation of another surface. One example of episodic aggradation was within the valley reach V174 where a piece of floodplain was elevated by a much-greater-than-average thickness of new sediment (average = 0.44 m, maximum = 0.98 m), in a few days, to the equivalent height of the Lightning terrace. Another example of episodic aggradation was the deposition of ~0.20 m of new sand (175 m upriver from PR 163), which elevated, in a few days, an upper Lightning terrace to approximately the level of an adjacent lower Moorcroft terrace. Based on historical accounts and interviews with local ranchers, the larger 1923 flood on Powder River elevated a portion of the Lightning terrace near PR120 (see Fig. 1, site of a historic battlefield in 1876; Vaughn, 1961, pp. 66, 89) to the height of the Moorcroft terrace. This process does not increase the planform area of the Moorcroft terrace because the original surface was the Lightning terrace and there will be a discontinuity of lithological characteristics at the contact between the two terraces. Thus, local aggradation by extreme floods is an evolutionary process that creates terrace with multiple elevations and explains some of the overlap in heights of the floodplain and terrace surfaces (Fig. 9). Along a reach of river, terraces will have multiple elevations, and thus, approximate a continuum of elevations. However, at any

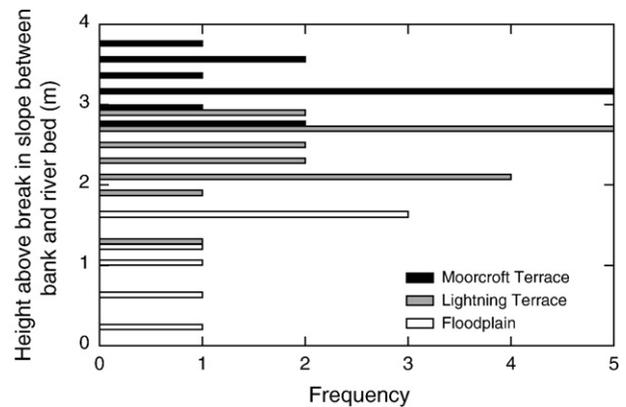


Fig. 9. Histogram of the heights of the floodplain and terraces at 19 cross sections across Powder River. Some cross sections had terraces on both sides, and other cross sections did not.

specific cross section of the river, terraces will still lie at discrete heights above the riverbed.

5.3. Coarsening-upward sequence

Coarsening-upward sequences are caused by a combination of temporal changes in the vertical and longitudinal relative concentrations of sediment particle sizes and the effects of vegetation on sediment transport.

5.3.1. Time-varying changes of sediment concentration

As the water rises, the source of sediment for the overbank flow originates from farther below the water surface. The vertical distribution of sediment concentration within the water column (Rouse, 1950; Middleton and Southard, 1984) depends on the Rouse number, $p = w / \kappa u_*$, where κ is von Karman's constant = 0.408, w (m s^{-1}) is the particle settling velocity, and u_* (m s^{-1}) is the shear velocity. The shear velocity is given by $u_* = \sqrt{ghS}$, where g is the acceleration of gravity (9.8 m s^{-2}), h (m) is the water depth, and S is the bed slope. No direct measurements of the vertical distribution of sediment concentration were made during the flood, so we calculated theoretical distribution using equations given by Middleton and Southard (1984) to model the changes in the relative concentrations of sediment. Water depths in the channel at the beginning of the flood were typically between 1 and 2 m, and the respective Rouse numbers for silt and clay would have been small enough (Table 3) such that the concentrations were vertically uniform throughout the water column. However, for sand, a concentration gradient probably extended throughout the water column, and as the water deepened with time over the inundated surface, the relative proportions of sand- to clay-size particles increased with depth below the water surface. For example, the estimated relative proportion of very fine sand to clay at 0.01 m below the surface was 0.44. As the water level rises this proportion would increase to 0.63 at 1.0 m below the surface (~50% increase). Likewise for medium sand, the relative proportion would increase with time from 0.01 at 0.01 m to 0.18 at 1.0 m (>1000%). Thus, as the water depth increased with time, the relative proportion of the sand increased.

Table 3
Suspended-sediment properties during 1978 flood at Broadus

Size	Median settling velocity (mm)	Critical shear velocity (m s ⁻¹)	Rouse number ^a <i>p</i>	Arvada, Wyoming			Broadus, Montana		
				Concentration (mg L ⁻¹)			Concentration (mg L ⁻¹)		
				19 May	20 May	21 May	18 May	22 May	
Clay	<0.004	<0.0000007	~0.003	0.000015	16,185	15,867	12,160	5211	10,170
Very fine–Fine silt	0.004–0.016	0.000018	0.004–0.007	0.00036	7470	7740	6400	3088	5424
Medium–Coarse silt	0.016–0.062	0.00034	0.007–0.014	0.0068	9960	8514	8640	7527	4746
Very fine sand	0.062–0.125	0.0031	0.014–0.017	0.061	5810	4257	3520	2702	1582
Fine sand	0.125–0.250	0.0095	0.017–0.019	0.19	2075	1935	960	579	678
Medium sand	0.250–0.500	0.026	0.019–0.022	0.52	0	387	320	193	0
Total					41,500	38,700	32,000	19,300	22,600

^a Rouse number, *p*, was computed for bankfull depth of ~1.4 m or shear velocity = 12 cm s⁻¹.

Relative suspended-sediment concentration can change because the longitudinal advective transport from upstream sources also varies with time. This advection process (proposed by Iseya, 1989) applies where the fine fraction or “wash load” is supply limited, and the concentration of fines decreases with time to produce a relative increase in the concentration of coarse particles.

5.3.2. *Effect of vegetation*

Stems of vegetation along the banks and across the normally subaerial surfaces of the bends reduce the water velocity sufficiently to deposit sediment (Dietz, 1952), and, under some conditions, to deposit mud. The criterion for deposition is that the boundary shear velocity is less than the critical shear velocity for maintaining particles in suspension. The effective

boundary shear velocity, u_*^v , in a flow with vegetation is given by Smith (2004, 2006) as

$$u_*^v = \sqrt{\frac{u_*^2}{(1 + \sigma_D)}} = \sqrt{\frac{(ghS)}{1 + \sigma_D}} \tag{5}$$

The drag function, σ_D , is

$$\sigma_D = \alpha \left[\frac{C_D}{2} \right] \left[\frac{u_{\text{mean}}}{u_*} \right]^2 \tag{6}$$

where α is the non-dimensional vegetation density parameter, and C_D is the drag coefficient for a single, cylindrical stem. Drag on a vertical cylinder in open-channel flow can be affected by open-channel turbulence and surface waves (Shen,

Table 4
Effect of vegetation on shear velocity during 1978 flood^a

Vegetation	Water depth (m)	Shear velocity on bare surface u_* (m s ⁻¹)	Stem diameter D_s (m)	Spacing l (m)	Vegetation density parameter α	Shear velocity in vegetation u_*^v (m s ⁻¹)
Grass						
Wide spacing	0.010	0.013	0.005	0.05	0.02	0.0060
Close spacing	0.010	0.013	0.001	0.01	0.10	0.0030
Wide spacing	0.10	0.041	0.005	0.05	0.2	0.0051
Close spacing	0.10	0.041	0.001	0.01	1.0	0.0023
Wide spacing	1.00	0.13	0.005	0.05	2	0.0041
Close spacing	1.00	0.13	0.001	0.01	10	0.0018
Willows						
Wide spacing	0.10	0.041	0.01	1	0.001	0.036
Close spacing	0.10	0.041	0.01	0.2	0.025	0.014
Wide spacing	1.0	0.13	0.01	1	0.01	0.053
Close spacing	1.0	0.13	0.01	0.2	0.25	0.012
Wide spacing	2.0	0.18	0.01	1	0.02	0.053
Close spacing	2.0	0.18	0.01	0.2	0.50	0.011
Cottonwood trees						
Wide spacing	1.0	0.13	0.2	2	0.05	0.026
Close spacing	1.0	0.13	0.2	0.4	1.25	0.0052
Wide spacing	2.0	0.18	0.2	2	0.10	0.024
Close spacing	2.0	0.18	0.2	0.4	2.5	0.0049

^a Shear velocity on bare surface is equal to the square root of the product of the depth and slope = 0.0017; the drag coefficient, $C_D = 1.2$, for a cylinder was used for the stems; bed roughness parameter z_0 was $k_s/30$, where $k_s = 0.125$ -mm sand.

1973), but these are not important effects for shallow flow over floodplains and terraces so we have used a typical value of $C_D=1.2$ from laboratory experiments (Schlichting, 1987). The effective boundary shear velocity, u_*^v , without vegetation ($C_D=0$) reduces to u_* . The ratio of the depth-average mean velocity to the shear velocity, $\frac{u_{\text{mean}}}{u_*}$, is

$$\frac{u_{\text{mean}}}{u_*} = \frac{1}{\kappa} \left[\ln \left(\frac{h}{z_0} \right) - 0.74 \right], \quad (7)$$

where the bed roughness parameter, z_0 , is often taken to be the grain diameter/30 or equal to the height of ripples or bed features if they are present (Smith, 2004, 2006).

5.3.2.1. Coarsening-upward sequences. We modeled the effect of different vegetation on the shear velocity and overbank sediment deposition using Eqs. (5)–(7). A typical grain diameter of 0.125 mm was used to calculate the ratio in Eq. (7). A few different combinations of stem diameters and spacing, and different water depths were used to compute values of u_*^v that would bracket typical conditions on the floodplain and terraces (Table 4). Grass reduced the shear velocity on a bare surface below the critical velocity for most silt-size particles (>0.004 mm). For example, a stem diameter of 0.001 m and spacing of 0.01 m reduced the critical velocity below that for clay-size particles (Table 4). The effective shear velocity decreases as the depth increases for the grass characteristics listed in Table 4 so that clay- and silt-size particles would be deposited for any water depths >0.01 m. The first layer of the coarsening-upward sequence was 0.02- to 0.07-m-thick mud and silt, which always had imbedded pieces of partially decomposed grass. Grass particles were much less common in the overlying very fine sand layers.

Sand is deposited from suspension close to the bank. Sand would not have been deposited across the entire grassy surface because settling velocities for sand are greater than those for mud and silt-size particles. For example, in water ~0.02 m deep with mean velocity of ~0.03 m s⁻¹, calculations indicate that medium sand would have been deposited within 0.03–0.07 m of the bank, fine sand within 0.07–0.22 m, and very fine sand within 0.22–0.75 m; but the silt and clay would have been deposited over a distance of 0.75–290 m and 290–1430 m, respectively. As the grass fills up with mud and fine silt, the trapping efficiency of the sand-size particles decreases and eventually not even mud or fine silt-size particles are trapped.

Shear velocity increases as the water depth increases above the height of the grass and also in response to the smoother boundary provided by the newly deposited mud. The relative concentration of sand to clay will increase (as discussed above) and the sand will be transported farther from the bank before being deposited (for example: in 1 m of water, very fine sand will be transported 7–22 m). Additionally, sand can be transported as bedload from the initial thick deposit near the edge of the terrace or bank of the floodplain (Pizzuto et al., in press). This tends to produce more uniform thick layers over larger areas of the floodplain or terraces than occupied by the initial sand deposit. Coarse sand layers will cover, protect, and

preserve each underlying finer layer in the coarsening-upward sequence from erosion.

We estimate that most of the newly deposited sediment on the Lightning and Moorcroft terraces was deposited on the rising limb of the water discharge. No reversal occurred of the coarsening-upward sequence on the falling limb (such as those reported by Knox, (1987), for large floods on tributaries to the upper Mississippi River). Knox attributed the coarsening-upward sequence and its reversal to a substantial source of silt in the form of slumps at the base of the banks, which was added to the source of silts in the channel bed. Four possible reasons combine to explain the lack of late-stage fining-upward deposits on the falling limb of the discharge: (i) water fell more rapidly than it rose during the flood of 1978 (Fig. 3), (ii) water receded so quickly from the terraces that there was little time to deposit more sediment; (iii) the waters that receded back into the channel were relatively clear having already deposited most of their suspended-sediment; and (4) the short, early-season grass became so clogged with rising-stage deposits and so covered with newly deposited sand that it could no longer trap fine sediment during the falling river stages.

5.3.2.2. Lee dunes. Lee dunes have been called shadow dunes (Hesp, 1981), wake deposits (Miller and Parkinson, 1993), and anchor dunes (Cooke et al., 1993) in the eolian and fluvial literature. In both fields, lee dunes provide information about the depositional environment. In the present study, the edge of the planform outline of a lee dune indicates the critical shear velocity for fine and very fine sand (~0.017 m s⁻¹). Any definite stratigraphy within the deposit was not resolved by the initial sampling pattern for particle size, but visual inspection during the sampling indicated that more detailed sampling might reveal stratigraphy even within the rather narrow range of particle sizes found in this study. Stratigraphy could provide information on the transport of particle size with time as the lee dunes grew during the flood. The rate of growth of the deposit may indicate the suspended-sediment concentrations, but controlled laboratory experiments would be necessary to determine this relation. The height of eolian lee dunes depends on the width of the vegetation (Hesp, 1981). In the case of fluvial lee dunes, the height also must be limited by the water depth, and thus, provides a direct minimum estimate of the water depth on the floodplain or terrace. The length of eolian lee dune has been shown to be a function of the width of the vegetation and the wind velocity (Hesp, 1981). At the present time, we are not aware of any quantitative relations between fluvial lee dune length and flow velocity across floodplain or terrace surface. However, this information would be useful in establishing the depositional environment associated with the formation of lee dunes, which might be found in the geological record.

The morphology of the lee dunes on the Lightning terrace, which were revisited and sampled for particle size in 2007, had changed little in the course of 29 years. The dune ridges, which may have been triangular in cross section immediately after the flood, have probably rounded off with time so that the cross-sectional profiles now have a parabolic shape. In some cases the

trees around which the dune formed have died, fallen over, and begun to decompose. Eventually these dune deposits will appear on the present terraces without their formative agents—the trees. These depositional features have probably been incorporated into the stratigraphic record of floodplains and terraces without any indication of the presence of a tree, thus the interpretation must rely on the morphology and additional stratigraphy that can be found in lee dunes in the future.

6. Summary and conclusions

Fluvial terraces are geomorphic features that have been abandoned by the stream in the sense that processes of accretion are no longer actively increasing the planform area of the terrace. However, terraces continue to evolve by episodic aggradation during extreme floods, such as the 1978 flood on Powder River. This aggradation is a race against terrace degradation by lateral erosion, which will eventually erode all terrace remnants and the terraces will become extinct. The flooded planform area of the Lightning and Moorcroft terraces decreased by 3.2% and 4.1% during the 1978 flood, but the same terraces, on average, aggraded vertically by 0.16 m and 0.063 m, respectively. The thickness of the episodic aggradation depends on the heights of the terraces, the magnitude of the extreme flood, and the location of the terraces relative to the active channel. Thus, the discrete heights of the original terraces have evolved with time such that along a reach of a river, a terrace can have multiple elevations that approximate a continuum, but at a specific cross section of the river, one can still see terraces with discrete heights relative to each other. If the magnitude of extreme floods increases in response to climate change, then the thickness of the episodic aggradation on terraces, often thought to be above the reach of floods, will probably increase. Local episodic aggradation can sometimes elevate a floodplain to the height of a terrace or a terrace to the equivalent height of a higher terrace.

Two characteristic types of deposition were observed after the 1978 flood. Coarsening-upward sequences were common on the floodplain, Lightning terrace, and Moorcroft terrace; and lee dunes were composed of fine and very fine sand and were common on the floodplain and Lightning terrace. The percentage of sand in samples of the newly deposited overbank material decreased from 48% on the Lightning terrace to 16% on the higher Moorcroft terrace, and consequently, the percentage of silt and clay increased from 52% on the Lightning terrace to 84% on the Moorcroft terrace.

Vegetation controlled the type, thickness, and stratigraphy of the aggradation on terrace surfaces. Grass reduced the shear velocities and caused the deposition of coarsening-upward sequences. Overbank flow was initially shallow (~ 0.02 m deep) and relatively slow (~ 0.03 m s⁻¹) and a mud (silt and clay) layer was deposited among the grass stems. As the water depth increased, the relative proportion of coarser sediment increased, and progressively coarser sand layers were calculated to have been transported farther from the channel and deposited on top of the fine sediment. These successive coarser layers require greater critical shear velocities to erode, and thus, protected

and preserved coarsening-upward sequences on the floodplain and terrace surfaces.

Lee dunes were composed of fine and very fine sand and were deposited around and downstream from individual trees and groups of trees or shrubs on the floodplain and Lightning terrace. The planform shape, dune height, and dune length provide some information about the flow variables such as water depth, flow velocity, and suspended-sediment particle size. The results and additional laboratory investigations should be able to further quantify the relation between lee dune morphology and flow variables, which can be used in interpreting depositional stratigraphy at other locations.

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