THESIS

WOOD-MEDIATED GEOMORPHIC EFFECTS OF A JÖKULHLAUP IN THE WIND RIVER MOUNTAINS, WYOMING

Submitted by

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ABSTRACT

WOOD-MEDIATED GEOMORPHIC EFFECTS OF A JÖKULHLAUP IN THE WIND RIVER MOUNTAINS, WYOMING

A jökulhlaup burst from the head of Grasshopper Glacier in Wyoming's Wind River Mountains during early September 2003. Five reaches with distinct sedimentation patterns were delineated along the Dinwoody Creek drainage. This thesis focuses on a portion of the jökulhlaup route where erosion of the forested banks created sixteen large logjams spaced at longitudinal intervals of tens to hundreds of meters. Aggradation within the main channel upstream from each logjam created local sediment wedges, and the jams facilitated overbank deposition during the jökulhlaup. Field surveys during 2004 and 2006 documented logjam characteristics and associated erosional and depositional features, as well as initial modification of the logjams and flood deposits within the normal seasonal high-flow channel. Overbank deposits have not been altered by flows occurring since 2003. Field measurements supported three hypotheses:

(i) Logjams present along the forested portions of the jökulhlaup route are larger and more closely spaced than those along adjacent, otherwise comparable stream channels that have not recently experienced a jökulhlaup;

(ii) logjams are not randomly located along the jökulhlaup route, but instead reflect specific conditions of channel and valley geometry and flood hydraulics; and

(iii) the presence of logjams facilitated significant erosional and depositional effects.

This thesis documents a sequence of events in which outburst floodwaters enhance bank erosion and recruitment of wood into the channel and thus the formation of large logjams. These logjams sufficiently deflect flow to create substantial overbank deposition in areas of the valley bottom not commonly accessed by normal snowmelt peak discharges and through this process promote valley-bottom aggradation and sediment storage.

Changes in the occurrence of glacier outburst floods thus have the potential to alter the rate and magnitude of valley-bottom dynamics in these environments, which is particularly relevant given predictions of worldwide global warming and glacial retreat. Processes observed at this field site likely occur in other forested catchments with headwater glaciers.

ACKNOWLEDGEMENTS

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Thank you to Peter Nelson and Sara Rathburn for serving on my committee.

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INTRODUCTION

Jökulhlaups are a type of outburst flood that originate from meltwater ponded by glacial ice (Sturm et al., 1987). Ponding can take the form of supraglacial, subglacial, or ice-marginal lakes. Ice-dammed lakes are present in every glaciated region of the world (Tweed and Russell, 1999) and are likely to have been even more common during the last stages of the Pleistocene (Baker and Bunker, 1985).

Jökulhlaups have a different failure mechanism than moraine-dammed lakes (Cenderelli and Wohl, 2003; Kershaw et al., 2005), but many of the hydrologic and geomorphic effects are similar (Costa and Schuster, 1988). Various types of failure mechanisms for ponded glacial meltwater have produced some of the largest known floods on Earth (Baker and Nummedal, 1978; Teller, 1990; Baker et al., 1993; Kehew, 1993), as well as localized severe flooding that creates persistent erosional and depositional features (Magilligan et al., 2002) and serious hazards for mountainous regions (Vuichard and Zimmermann, 1987; Shroder et al., 1998).

Volcanic and meteorological events can trigger jökulhlaups by causing rapid increases in the volume of water ponded by the ice, or outbursts can result from continually changing meltwater levels within a glacier in the absence of discrete external triggers (Tweed and Russell, 1999; Russell et al., 2000). Evans and Clague (1994) describe *the jökulhlaup cycle* that begins when the damming glacier retreats past a critical threshold such that the ice dam can no longer block the passage of water ponded behind it.

The most common conduit for lake drainage is a subglacial meltwater tunnel. Meltwater drains through these kinds of tunnels as a jökulhlaup until the ice dam can once again block downslope flow of water because the overburden pressure of ice and the rate of glacier movement close the flood conduit (Tweed and Russell, 1999). After the conduit closes, meltwater once more builds up until the threshold for ice-dammed lake drainage is exceeded. Jökulhlaups continue to occur at regular or irregular intervals until the glacier retreats and the formation of a permanent conduit prevents ponding of meltwater or until the glacier advances and creates a water-tight dam (Evans and Clague, 1994).

As glaciers recede, ice dams grow thinner, and progressively less water is required to exceed the threshold for creating jökulhlaups, which thus become more frequent but smaller in magnitude (Tweed and Russell, 1999). The warmer summers and shorter, warmer winters likely to accompany global warming may increase the incidence of jökulhlaups as glacial ice melts and thins more rapidly (Vuichard and Zimmermann, 1987).

When ponded water exceeds the hydrostatic pressure of the restraining ice or moraine dams, the resulting flood is commonly of higher magnitude and shorter duration than normal seasonal high flows produced by snowmelt (Blown and Church, 1985; Costa and Schuster, 1988). The greater energy expended during an outburst flood creates distinctive, persistent erosional and depositional features that have been described by many investigators (Desloges and Church, 1992; Maizels, 1997; Carrivick et al., 2004).

Most of these studies focus on the geomorphic effects of outburst floods on sparsely forested, high-mountain valleys (Carling and Glaister, 1987; Shroder et al., 1998; Cenderelli and Wohl, 2001, 2003; O'Connor et al., 2001) or on Icelandic-type, un-forested glacial outwash plains (Maizels, 1991, 1997; Fay, 2002; Magilligan et al., 2002). The most common pattern described for the geomorphic effects of outburst floods in mountainous environments is that of erosion in narrower, steeper valley segments; deposition in wider, lower gradient sections of the valley; and downstream decreases in the magnitude and extent of geomorphic effects as the outburst flood attenuates and approaches the magnitude of more commonly occurring peak flows (Cenderelli and Wohl, 2003).

Jökulhlaups may also occur in forested mountainous environments, where wood entrained by the flood may potentially mediate erosional and depositional effects. Descriptions of how wood alters the geomorphic effects of jökulhlaups or moraine-dam outbursts in mountain environments are quite rare, however. Blown and Church (1985) briefly mention that overbank flows created large accumulations of wood along the floodplain of the Nostetuko River in British Columbia during a 1983 jökulhlaup.

The primary objective of this thesis is to examine wood-mediated geomorphic effects along the path of the 2003 Grasshopper Glacier jökulhlaup in the Wind River Mountains of Wyoming.

Existing work documents a wide variety of geomorphic effects of wood in stream channels and across floodplains (Abbe and Montgomery, 2003; Wohl, 2014). Wood within the channel increases flow resistance and hydraulic roughness (Curran and Wohl, 2003; Wilcox et al., 2011) and promotes flow separation and localized scour of the bed and banks (Daniels and Rhoads, 2003). Wood alters bed-surface grain size and sediment storage (Faustini and Jones, 2003; Ryan et al., 2014) to the point of creating forced-alluvial reaches of lower gradient than might otherwise be present (Montgomery et al., 1995, 2003) and alters bedform configuration (Baillie and Davies, 2002; Gomi et al., 2003). Wood increases resistance of the channel boundary to erosion (Brooks et al., 2003) and creates jams that promote channel avulsion, anastomosing channel planform, and overbank flow (Collins and Montgomery, 2001; Jeffries et al., 2003; O'Connor et al., 2003; Montgomery and Abbe, 2006; Collins et al., 2012), as well as enhanced floodplain deposition (Sear et al., 2010).

Although most investigators have not focused on the geomorphic effects of wood during large floods, the mobility of at least some of the wood already present in a channel is likely to increase during a large flood such as a jökulhlaup. Erosion of forested stream banks during the flood probably also increases the volume of wood present in the channel. The greater abundance of wood in transport could create large jams where an obstacle is present or during the waning stages of the flood, and these jams could enhance localized erosional and depositional features by promoting flow separation, increasing hydraulic roughness, and decreasing conveyance of the channel.

Although a rainfall- or snowmelt-generated flood could also erode the stream banks and form logjams, the higher magnitude and shorter duration of a jökulhlaup relative to other types of floods could greatly enhance bank erosion and wood recruitment during the flood, creating larger and more closely spaced logjams than those resulting from other types of peak flows.

These possibilities are examined along forested portions of the 2003 Grasshopper Glacier jökulhlaup route. Specifically, the following hypotheses are tested:

(i) Logjams present along the forested portions of the 2003 jökulhlaup route are larger and more closely spaced than those along adjacent, otherwise comparable stream channels that have not recently experienced a jökulhlaup. This hypothesis reflects the possibility that increased transport and recruitment of wood during the jökulhlaup enhanced the size and frequency of logjams along the channel. Because the geomorphic effects of individual wood pieces are enhanced where logjams form, the existence of larger and more closely spaced logjams as a result of a jökulhlaup substantially increases the persistent geomorphic effects of the jökulhlaup.

(ii) Logjams are not randomly located along the jökulhlaup route, but instead reflect specific conditions of channel and valley geometry and flood hydraulics. The specific conditions referred to indicate that the logjams are more likely to occur within a limited range of conditions, rather than being randomly distributed across all the channel and valley types present along the jökulhlaup route.

(iii) The presence of logjams facilitated significant erosional and depositional effects. Significant here implies greater magnitude and persistence than erosional and depositional effects associated with annual peak snowmelt flows. This hypothesis, if corroborated, suggests that jökulhlaups can exert an important influence on the geomorphic evolution of valley bottoms in forested mountain catchments with headwater glaciers.

FIELD AREA

The study site is located in Wyoming's Wind River Range, which contains 63 glaciers, including seven of the ten largest glaciers in the U.S. Rocky Mountains (Bonney, 1987; Pochop et al., 1990). Grasshopper Glacier lies at the Continental Divide between high-level erosion surfaces and the eastward-facing cirque walls of the mountain front (**Figure 1**).

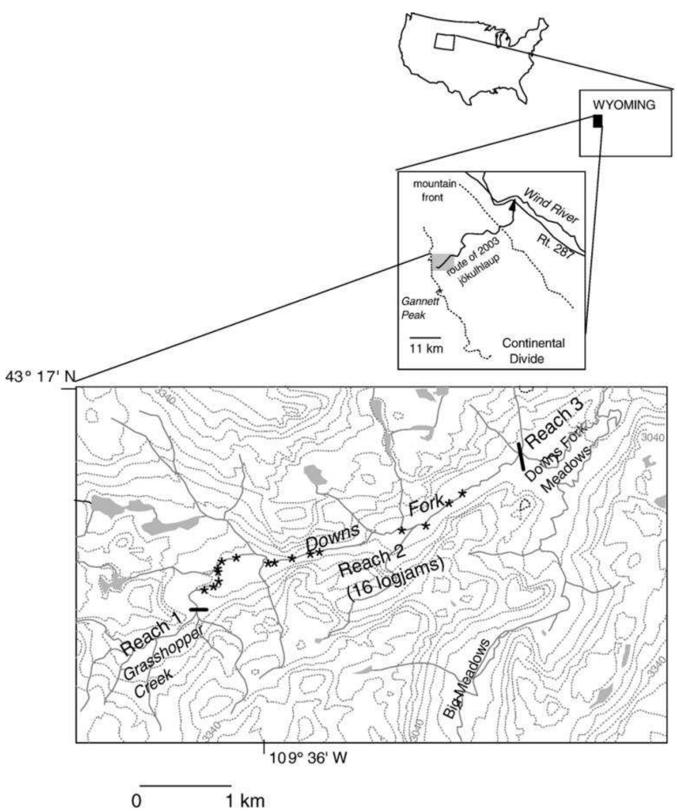

Figure 1.(A) Location map, showing route of the 2003 jökulhlaup and reaches 1 through 3. Smaller inset maps show the location of the State of Wyoming within the continental United States and the location of the detailed map within Wyoming. Within the detailed map, dotted gray lines indicate topographic contours at 60-m intervals, gray polygons indicate lakes, and solid gray lines are streams. Asterisks indicate location of logjams in study reach 2. Streamflow is from lower left toward upper right.

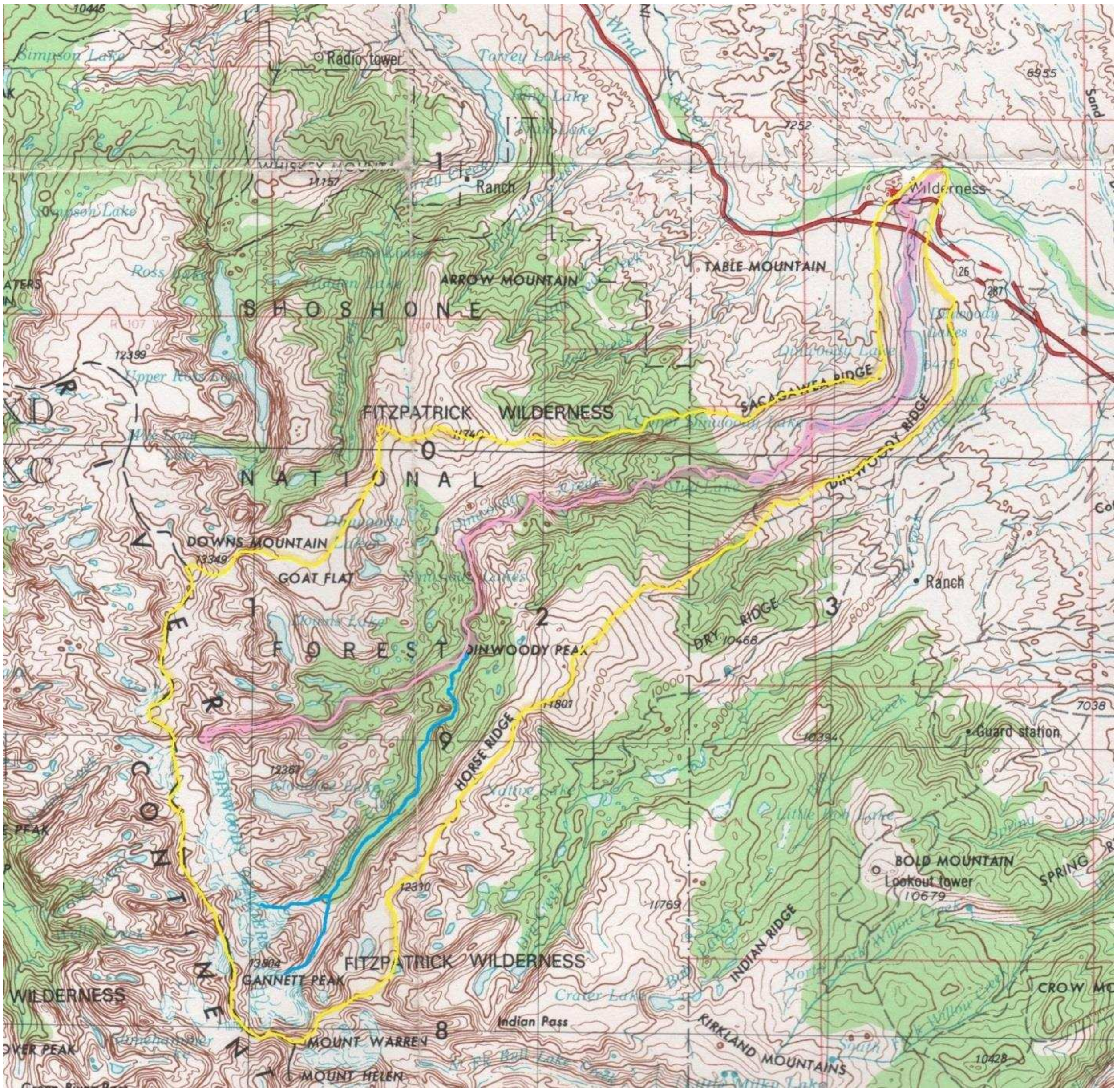

Figure 1.(B) Dinwoody watershed area (outlined in yellow)showing route of jökulhlaup (in pink) along Downs Fork which is tributary to Dinwoody Creek (in blue). Below the confluence, the jökulhlaup flows down the mainstem Dinwoody Creek. Topographic contours at 60 m intervals.

The glacier is presently approximately 2.5 km wide and covers 3.6 km^2. The glacier flows 2.8 km, dropping from 3650 m at its headwall to its terminus at 3435 m. A jökulhlaup burst from an ice-dammed lake at the head of the glacier during early September 2003, draining an estimated volume of 3.2 million m^3 of water from the 0.12-km^2 lake. Meltwater drained beneath 18 to 21 m of glacial ice in the vicinity of the tunnel inlet and entrained subglacial sediment as it flowed for 2.8 km beneath the glacier. The sediment-laden flood then followed the steep, narrow valley of Grasshopper Creek to Downs Fork and along Downs Fork to the confluence with Dinwoody Creek, before continuing down Dinwoody to the Wind River valley at the base of the mountains.

The Wind River Mountains are composed predominantly of high-grade Archean gneiss and granites that were thrust westward over Phanerozoic sediments during the Laramide orogeny (Frost et al., 1986). Granites dominate the southern half of the range, whereas older gneisses form most of the northern half, in which the study area lies.

Mean annual precipitation is 100 to 130 cm over most of the mountains, locally rising to more than 150 cm. The western side of the range receives more precipitation than the eastern side because moisture moves predominantly eastward from the Pacific Ocean. Vegetation is strongly zoned by elevation, from alpine tundra above 3000 m elevation, to spruce-fir (*Picea engelmanni, Abies lasiocarpa*) forest at 3000 to 2500 m, and lodgepole pine (*Pinus contorta*) forest at lower elevations within the mountains (Reed, 1976). Streamflow in the study area is dominated by the seasonal snowmelt and glacier-melt peak, which rises in mid-June and gradually falls through August.

The 2003 jökulhlaup was recorded at a U.S. Geological Survey stream gage on Dinwoody Creek approximately 33 km downstream from the ice-dammed lake (gage 06221400, drainage area 245 km^2). The gage recorded a peak of 36.5 m^3/s on 9 September 2003, which rose from a base flow of 7 m^3/s on 6 September. Peak annual flow has been recorded at this site since 1956, although several years of records are missing. Annual peak snowmelt flow at the gage averaged 23 m^3/s (range 14 to 42 m^3/s) during the period 1956 to 2005. Peak flow most commonly occurs during June or July, but a peak flow on 11 September 1973 may also have been a jökulhlaup based on its late occurrence. The 2003 peak flow has a recurrence interval of five years at the gage.

The route followed by the 2003 jökulhlaup has pronounced downstream variability in valley and channel morphology. During a field reconnaissance in summer 2004, the flood route was divided into five reaches based on depositional patterns associated with these downstream variations (**Table 1**).

Table 1. Characteristics of depositional reaches along the route of the 2003 Grasshopper Glacier jökulhlaup. Intervening distances (e.g., 6.1 to 8.2 km) represent reaches that remained stable or experienced erosion.

Reach	Distance from outburst lake (km)	Average valley gradient (m) Average valley-bottom width (m)	Description
1	4.2–6.1	0.045 200	Bar deposition across entire valley bottom; individual bars up to 0.2 km wide and 1.4 m high
2	8.2–13.2	0.025 80–400	16 log jams recorded in valley in 2004; largest jam up to 12 m long (parallel to channel), 35 m wide, 3.5 m high
3	13.2–15.3	0.013 300–600	Aggradation and log jams promoted channel avulsion; deposition of sediment in an ephemeral lake
4	16.7–21.2	0.025 500–1000	In-channel filling of riffle-pool sequences downstream from Dinwoody Falls
5	32–38.6	<0.01 700-800	Silt and clay deposition in Mud Lake, the Dinwoody Lakes, and the lower Dinwoody River valley to the Wind River floodplain

The first, upstream-most reach is relatively steep and narrow. Jökulhlaup deposition primarily took the form of expansion and longitudinal bars (Baker, 1978; O'Connor, 1993; Cenderelli and Cluer, 1998; Cenderelli and Wohl, 2003) composed of sediment ranging from boulders ≤1 m in diameter down to sand.

The second reach, of moderate gradient and valley-bottom width, is heavily forested. This reach, which is the focus of this thesis, was characterized by localized in-channel and overbank deposition that was strongly influenced by the formation of channel-spanning logjams composed of trees introduced to the channel by streambank erosion.

Downstream from reach 2, the valley widens into a broad glacial trough. Here the floodwaters temporarily ponded in an ephemeral lake. When the lake level subsided, 0.6 to 1.5 m of granitic silty sand was deposited over an area of approximately 28 ha. One-meter-high standing waves were observed along this portion of Dinwoody Creek during the jökulhlaup (S. Kearney, backcountry horsepacker, personal communication, November 2003). The creek avulsed from its pre-flood meandering course through this reach, creating numerous overflow channels and filling the pre-flood channel with sediment (**Figure 2**).

The valley grows locally narrower and steeper downstream in reach 4, where deposition occurred mainly within the active channel and consisted primarily of sand-sized and finer sediment. Reach 5 was characterized by substantial sediment deposition in Mud and Dinwoody Lakes, more than 30 km downstream from the jökulhlaup source. The existence of the wider, lower gradient valley segments that are designated reaches 3 and 5 suggests that substantial attenuation of the jökulhlaup peak discharge likely occurred before the peak reached the stream gage and that the recurrence interval of the 2003 peak was likely much longer than five years at upstream sites.

Figure 2. Aerial view of reach 3 in late September 2003 showing the extent of deposition during jökulhlaup (pale gray color) in floodplain wetlands, channel avulsion, and overbank flows. Flow is from right to left. Approximate maximum width of deposition is 400 m.

METHODS

Several sources of data and analysis were employed to evaluate the geomorphic role of logjams within reach 2 of the 2003 jökulhlaup route. Sixteen logjams were mapped with GPS units during an initial field season in autumn 2004. Cross-sectional profiles and longitudinal profiles of the channel and water surface were surveyed at these logjams using a transit level. Surveys included high-water marks from the jökulhlaup and normal peak flow level where this could be estimated in the field using changes in bank morphology and stratigraphy, riparian vegetation, and high-water marks such as debris.

A total of 37 cross sections and ten water-surface longitudinal profiles were surveyed in 2004. The majority of these surveys were associated with ten logjams spread along the length of reach 2.

Random-walk clast counts of bed and bar sediments were conducted at three sites. Volume of sediment deposited upstream from ten logjams was measured by surveying the cross sectional area of channel with and without in-channel sediment deposition and the longitudinal distance between the two, and assuming that the sediment-filled wedge immediately upstream from each jam tapered gradually with increasing distance upstream. The field site was re-visited during summer 2006 to re-survey cross-sectional and longitudinal profiles and to re-measure grain-size distributions at sites of previous clast counts. Twenty-four of the cross sections and all of the longitudinal profiles surveyed in 2004 were re-surveyed.

Older jökulhlaup deposits along the valley margins of the flood route were qualitatively evaluated. The location and size of logjams and associated upstream sediment storage were also surveyed in the adjacent Dinwoody Creek drainage upstream of the confluence with Downs Fork (Figure 1), which was beyond the 2003 jökulhlaup route. These field data were used to:

(i) Compare the magnitude of seasonal peak discharges and the September 2003 jökulhlaup;

(ii) examine wood-mediated geomorphic response to the jökulhlaup, as reflected in correlations among channel and valley morphology, size and spacing of logjams, sediment storage upstream from logjams, and downstream trends in logjam or sediment characteristics; and

(iii) evaluate the persistence of modifications to channel morphology and sediment movement associated with logjams by quantifying changes between 2004 and 2006 and by relating depositional features from the 2003 jökulhlaup to older deposits present along the valley margins.

Seasonal peak discharge was calculated using surveyed cross-sectional geometry from 2006, high-flow indicators, and the Manning equation. The Manning equation is: $Q = (1/n)\ A\ R^{2/3}\ S^{1/2}$ where *Q* is discharge (m^3/s), *n* is a coefficient for hydraulic resistance, *A* is cross-sectional area (m^2), *R* is hydraulic radius (m), and *S* is water-surface slope (m/m). Visually estimated *n* values based on similar channel geometry, gradient, and grain size relative to examples in Barnes (1967) were used and were varied by ±15% as part of a sensitivity analysis. The Manning equation calculates peak discharge at a single cross section and is the simplest approach to estimating instantaneous peak discharge in the field. Given the uncertainty in estimating pre-flood channel geometry at surveyed sites, the Manning equation provides a useful, first-order approximation of flood hydraulics.

RESULTS

Peak Discharge Estimation

Four of the 24 cross sections surveyed during 2004 and 2006 were chosen to calculate instantaneous peak discharge for normal seasonal peak flows and for the 2003 jökulhlaup. The cross sections chosen span the length of reach 2 and have well-defined banks and high-flow indicators on each side of the channel. Use of the Manning's equation to calculate instantaneous peak discharge of normal seasonal high flows (here, the annual snowmelt peak flow) resulted in values that ranged from 6.7 m^3/s at the upstream end of reach 2 to 7.6 m^3/s at the downstream end.

The same procedure, and the same cross sections, were used to estimate peak flow of the 2003 jökulhlaup. Because the necessary data to model flow along the entire length of the jökulhlaup or to evaluate parameters such as flow duration or temporal rate of attenuation were not measured, only maximum peak discharge was calculated at each of the four surveyed modeling reaches and then used to examine downstream attenuation in the instantaneous peak.

These calculations were conducted using first the 2004 survey data and then the 2006 data. The 2004 data likely reflect in-channel deposition during the waning stages of the jökulhlaup, as well as cumulative bank erosion during the jökulhlaup. The 2006 data are more likely to represent pre-jökulhlaup channel-bed elevations in that much of the jökulhlaup deposition within the main channel has been transported downstream, but the 2006 data also include continuing bank erosion. Direct comparison of re-surveyed cross sections indicates minimal post-jökulhlaup bank erosion, however, as discussed below. Consequently, I felt that the 2006 surveys provided the most appropriate channel geometry to use in calculating the peak jökulhlaup discharge.

The magnitude of peak discharge during the jökulhlaup appears to have undergone substantial attenuation within reach 2, dropping from an estimated value of 70 m^3/s at the upstream-most cross section to 48 m^3/s 50 m downstream, 36 m^3/s 1200 m further downstream, and 19 m^3/s at the downstream-most cross section.

Because of uncertainties in roughness values, channel cross-sectional geometry, and water-surface slopes at the time of the jökulhlaup peak discharge, each of these estimates has an associated uncertainty of ±25%.

The 2003 peak discharge of 36.5 m^3/s recorded downstream at the stream gage presumably reflects the much greater drainage area (drainage area of 245 km^2 at the gage). Based on the calculations for sites along reach 2, the jökulhlaup exceeded normal seasonal peak discharge by a ratio that varied from approximately 10 at the upstream site in reach 2 to >2 at the downstream site in this reach.

Wood-Mediated Channel Response To The Jökulhlaup

As is the case along the entire route of the 2003 jökulhlaup, reach 2 has longitudinal variability of valley-bottom width and gradient, as well as longitudinal variability in bedrock confinement of the active channel. Segment-scale (tens to hundreds of meters) channel gradient varies from 0.05 to 0.007 m/m along the 5 km of reach 2. The steeper channel segments occur along narrower sections of valley, and in some places the channel is incised into bedrock along both banks.

A low (1 to 3 m above the active channel), longitudinally discontinuous terrace is present along one or both channel banks in the wider sections of the valley. This terrace extends back to the valley walls or to a high glacial terrace and/or moraine 4 to 5 m above the low terrace. The distance from the active channel to the valley walls or higher surface varies from 50 to 500 m. Logjams numbered 2 through 16 occurred in valley bottom segments averaging 100 to 150 m in width.

The 2003 jökulhlaup left extensive deposits of cobble- to sand-sized sediment on the terrace. These deposits gradually decreased in volume and became finer-grained downstream. Upstream overbank deposits included substantial gravel and cobbles, whereas overbank deposits farther downstream were primarily fine to coarse sand. Overbank deposits locally reached >1 m in thickness (**Figure 3**).

Deposition tended to be most extensive where the terrace was lower in elevation and where logjams exacerbated overbank flooding during the jökulhlaup. In the middle and lower portions of reach 2, overbank deposition during the jökulhlaup occurred only at sites where logjams formed.

Figure 3. Photograph taken in 2006 of overbank deposits from the 2003 jökulhlaup. The pre-jökulhlaup top-of-bank shows as a dark line <1 m above the water surface. The younger conifers growing along the channel are rooted in this surface. The jökulhlaup deposits are the lighter colored sediments, a meter thick, that are burying the bases of the older conifers growing farther back from the channel. This site is near the downstream-most logjam in reach 2. Flow is from right to left.

Overbank deposition was more continuous in the upper portions of reach 2, which are immediately below the first substantial decrease in gradient and increase in valley width that occur along the 2003 jökulhlaup route. Indirect discharge estimation indicates that flow must be 1.5 to 3.7 times greater than average annual peak, depending on the specific site, to overtop the channel banks. Because peak discharge of the jökulhlaup decreased to just over twice the average annual peak flow at the downstream end of the study reach, logjams became progressively more important in forcing overbank flow in the lower part of this reach.

Extensive cutbanks observed in 2004 suggest that the jökulhlaup caused discontinuous bank erosion, which would have increased recruitment of wood into the channel from the riparian conifer forests that grow to the margins of the active channel (**Figure 4**). Formation of the logjams during the 2003 jökulhlaup was also indicated by the condition of much of the wood in the jams. Trees commonly still had abundant bark and numerous small, delicate branches and roots.

Logjams did not span the channel in the Downs Fork drainage prior to the 2003 glacial outburst flood (C. Voss, backcountry outfitter, personal communication, September 2004). Comparison of the longitudinal spacing, size, and sediment storage of logjams along the jökulhlaup route on Downs Fork and the adjacent Dinwoody Creek indicates that logjams are much more widely spaced along Dinwoody (**Figure 5**).

Figure 4. Photographs across the channel taken in 2006 of the extensive logjam at the downstream end of reach 2. This jam, which is 50 m long across the channel and 5 m wide, forced flow toward the right bank, exacerbating bank erosion during the jökulhlaup. Flow is from right to left. Upper view is toward the right bank, lower view is toward the left bank.

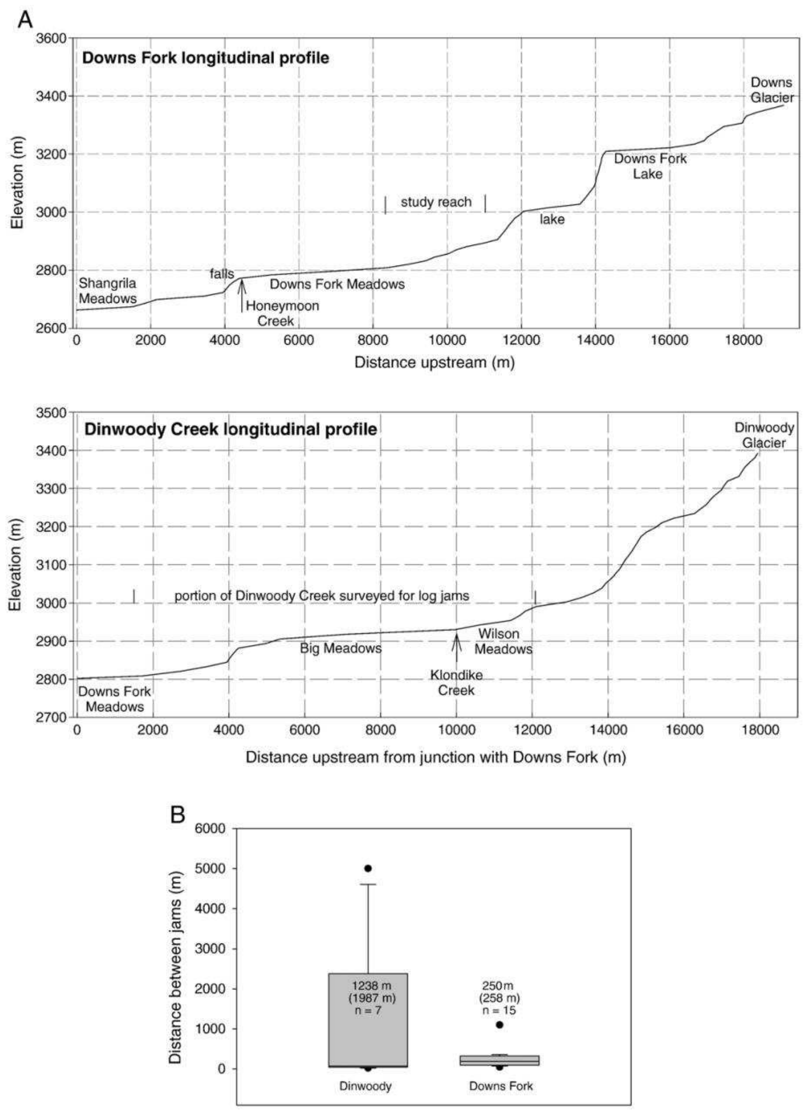

Figure 5. (A) Longitudinal profiles of Downs Fork and Dinwoody Creek reaches surveyed for logjams, listing features shown in **Figure 1. Figure 5 (B)** Plot comparing downstream spacing of logjams along Downs Fork and Dinwoody Creek reaches. Mean, standard deviation in parentheses, and sample size for each box plot are indicated.

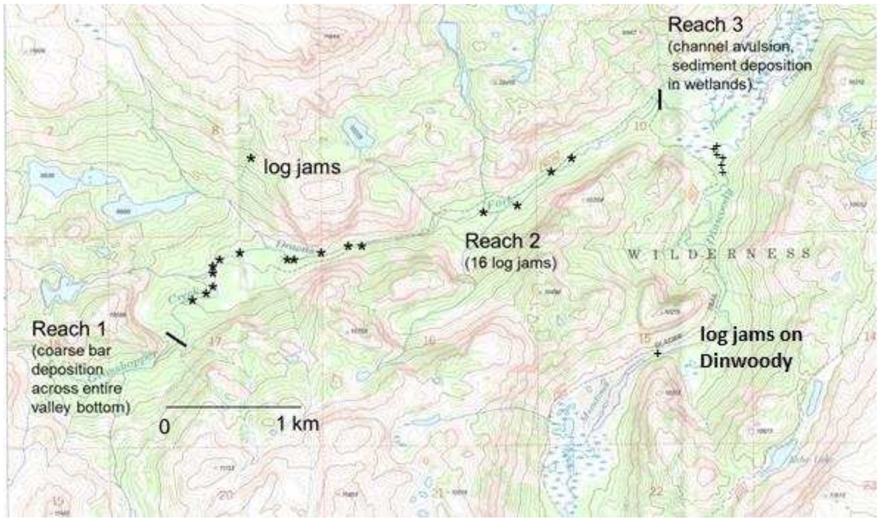

Figure 5.(C) Map showing location of logjams on Dinwoody above the confluence with Downs Fork.

With the exception of one logjam associated with an avalanche from the valley walls, the logjams along Dinwoody Creek create negligible upstream sediment storage, in contrast to the sediment aggradation consistently observed at logjams along Downs Fork. The differences in logjam characteristics and associated sediment storage between the two reaches, along with the differences in logjam characteristics on Downs Fork prior to and following the 2003 jökulhlaup, strongly suggest that bank erosion during the jökulhlaup initiated wood recruitment into the channel, facilitating the formation of logjams.

The measured characteristics of logjams in Downs Fork and Dinwoody Creek support the first hypothesis: that logjams present along the forested portions of the 2003 jökulhlaup route are larger and more closely spaced than those along adjacent, otherwise comparable stream channels that have not recently experienced a jökulhlaup.

Individual logjams surveyed in 2004 along the jökulhlaup route varied from 14 to 22 m in length across the channel and from 2 to 6 m in width (**Table 2**). Hydraulic jumps 0.75 to 1.5 m deep occur within the channel at the base of each logjam. Erosional and depositional features associated with the logjams indicated that many of the jams created backwater effects that increased overbank flow and deposition, as well as deflecting flow and enhancing local bank erosion.

Erosional features mainly took the form of a greater length and occurrence of vertical and undercut banks in the vicinity of logjams. Localized erosion also occurred on the low terrace in the form of scour around obstructions such as large boulders, tree trunks, and roots. At one site where a logjam formed just below a tributary junction, aggradation within the channel forced the jökulhlaup flow over the banks and created a small headcut (<1 m tall) on the low terrace.

Table 2. Characteristics of individual logjams within reach 2.

Logjam number	Length (m)	Width (m)	Length, diameter of key pieces (cm)	Estimated volume of sediment stored upstream (m^3)
1	14	3	150, 20; 150, 25	118
2	16	2	170, 20	671
3	15	3	160, 20	2303
4	22	2	120, 15; 160, 20	5076
5	20	2	150, 18; 150, 20	2775
6	15	3	200, 25; 180, 20	2827
7	14	5	150, 15; 160, 20	435
8	10	2	200, 18; 180, 20	577
9	11	6	150, 25; 150, 20	788
10	16	4	200, 15; 160, 18	2065
Average	15	3	163, 19.7	1764

Depositional features included streambed aggradation upstream of jams and greater extent and thickness of overbank deposition in the vicinity of logjams. Sediment deposition upstream from individual logjams varied from approximately 120 to 5075 m^3, with no downstream trend (**Figure 6**).

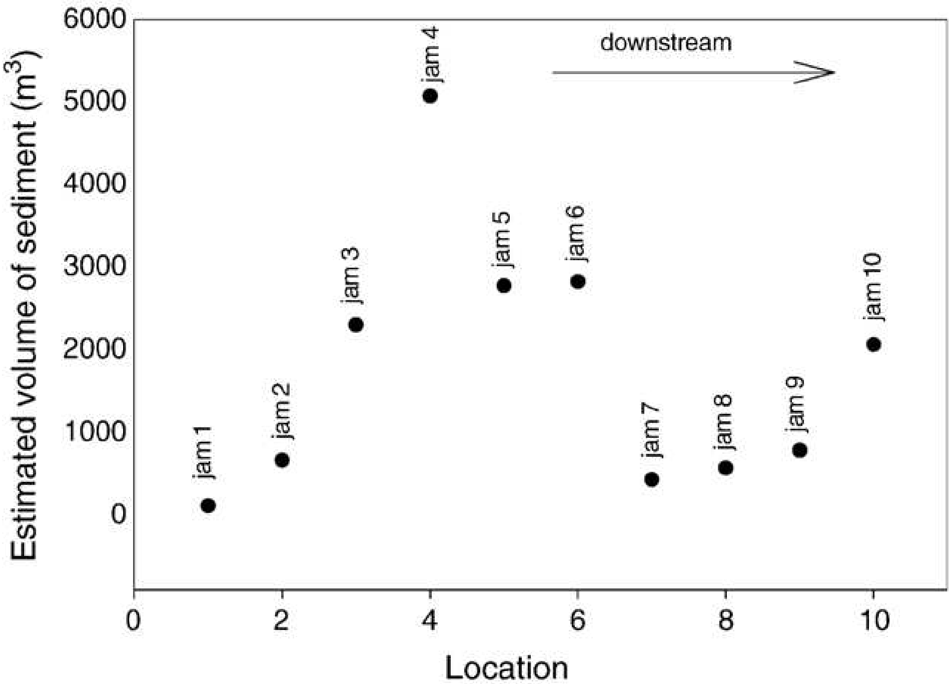

Figure 6. Volume of sediment deposited within channel upstream from each of the 10 logjams on which field surveys focused in 2004. Distance along x-axis does not reflect relative downstream distance of logjams.

The volume of each sediment wedge does not correlate with logjam dimensions. Rather, channel geometry appears to control the accumulation of sediment upstream from the logjams. Sites with taller banks that limited overbank flow and associated deposition, wider sections of channel, and lower gradient reaches all had sediment wedges with greater volume. Clast counts at sediment deposits upstream from logjams showed no downstream fining in 2004 and negligible change in grain size from 2004 to 2006. Clast size fractions measured in 2004 (D_{16} 8 to 10 mm; D_{50} 45 to 63 mm; D_{84} 66 to 126 mm) changed by 10 mm or less in 2006.

Logjams tended to be largest and most closely spaced where reach-scale gradient is 0.01 to 0.03, banks are alluvial rather than bedrock, and valley bottoms are 100 to 150 m wide. Reaches of steeper gradient, bedrock banks, and greater confinement did not have logjams (**Figure 7**). This supports the second hypothesis in suggesting that logjams are not randomly located along the jökulhlaup route, but instead reflect specific conditions of channel and valley geometry and flood hydraulics.

Persistence Of Wood-Mediated Channel Response

During summer 2006, 24 of the cross sections surveyed in autumn 2004 were re-surveyed as a means of quantifying short-term changes in channel geometry in the vicinity of logjams. Ten of these cross sections were upstream from logjams, seven were downstream, five were at logjams, and the remaining two cross sections were not in the vicinity of logjams but were at the downstream end of reach 2 where in-channel deposition of fine sediment occurred at the transition to the broad, low-gradient valley morphology of reach 3 (**Table 3**). In some cases, locating cross sections at exactly the same endpoints in 2004 and 2006 was difficult. Table 3 summarizes results for cross sections for which I was confident that the endpoints were identical.

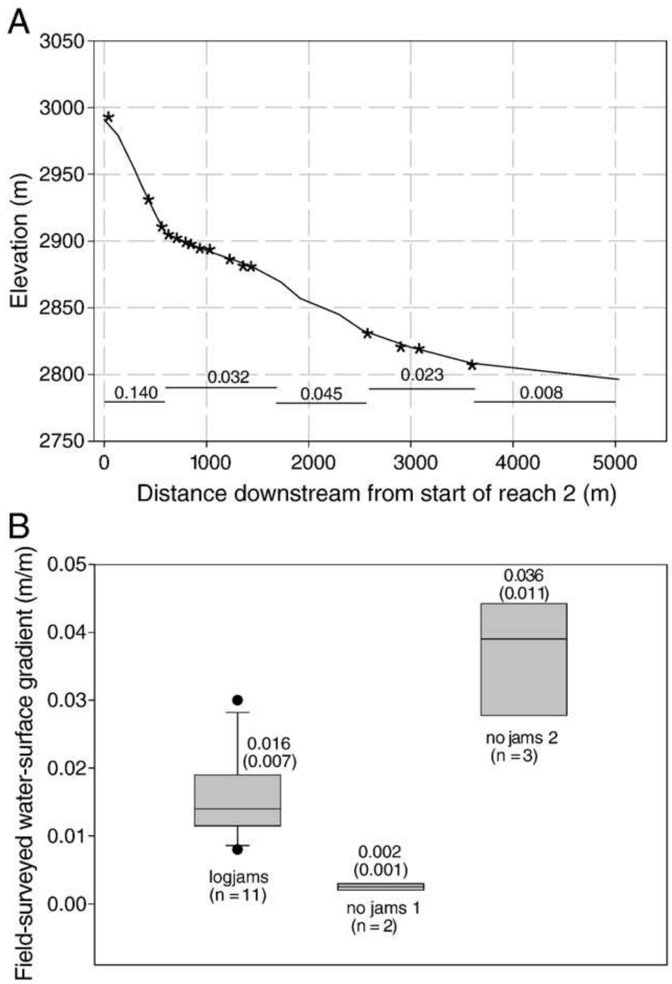

Figure 7. (A) Location of logjams (*****) in reach 2 plotted on longitudinal profile. Average gradient for each segment of the profile, as calculated from 1:24,000-scale topographic maps, is indicated below the profile. **(B)** Field-surveyed, water-surface gradient for sites with and without logjams. Number of surveyed reaches is listed below each box. Logjams are channel segments within reach 2 that contained jams; no jams 1 are channel segments at the transition between reaches 2 and 3, with no logjams; no jams 2 are channel segments within reach 2 with no logjams. Mean and standard deviation (in parentheses) are listed above each box. The line within each box indicates the median value, box ends are the upper and lower quartile, whiskers are the 10th and 90th percentiles, and dots are outliers.

The following trends were observed between the 2004 and 2006 surveys: of the ten cross sections upstream from logjams, six degraded, and four had negligible change; of the seven cross

Table 3. Changes at selected surveyed cross sections between 2004 and 2006. Cross section numbering starts at 1 for the upstream-most cross section surveyed and proceeds downstream for successively higher numbers. Net change describes vertical change in thalweg elevation (m) and cross-sectional area (m 2). Position of cross section is immediately upstream from logjam (above), at the logjam, or immediately downstream (below) from the logjam.

Cross section Number	Net change, 2004–2006	Position with respect to logjam
1	Degradation (−0.2 m vertical, −0.6 m^2)	Above
2	Aggradation (+0.6 m vertical,+12.4 m^2)	Below
4	Aggradation (+0.3 m vertical, +2.8 m^2)	At
5	Degradation (−0.2 m vertical, −0.8 m^2)	At
6	Aggradation (+0.4 m vertical, +2.4 m^2)	At
16	Degradation (−0.6 m vertical, −4.6 m^2)	At
20	Degradation (−0.2 m vertical, −2 m^2)	Above
21	Degradation (−0.6 m vertical, −5 m^2)	Above
23	Degradation (−0.4 m vertical, −5 m^2)	Below
24	No net change	Not near logjam
25	Degradation (−0.2 m vertical, −2 m^2)	Not near logjam

sections downstream from logjams, three degraded, two were unchanged, and the two upstream-most cross sections aggraded; of the five cross sections at logjams, two aggraded and three degraded; and of the two cross sections at the downstream end of the reach, one degraded, and one had no net change.

Net vertical change in thalweg elevation at each cross section varied between 0.2 to 0.6 m, with no trend in relation to downstream distance within the reach. Net change in cross-sectional area varied between 0.6 to 12.4 m^2, again with no downstream trend.

In addition, the magnitude of cross-sectional change does not vary with the downstream distance along the reach or the position with respect to the logjam, although the direction of change (aggradation versus degradation) does vary slightly with cross-sectional position relative to the logjam. Cross sections upstream from logjams are more likely to have degraded between 2004 and 2006.

Some previous studies have demonstrated a consistent downstream progression in sediment mobilization and channel cross-sectional change following introduction of a large volume of sediment (e.g., Madej and Ozaki, 1996; Wohl and Cenderelli, 2000). The lack of strong or consistent correlations between downstream distance and parameters such as net vertical change in thalweg elevation or net change in channel cross-sectional area are interpreted to indicate that localized controls on transport capacity associated with channel and logjam geometry exert a stronger influence on re-distribution of sediment deposited by the jökulhlaup than does distance downstream.

Comparison of eight water-surface longitudinal profiles in the vicinity of logjams that were surveyed in 2004 and again in 2006 indicates that the water-surface gradient increased with time at the two upstream-most jams but decreased at all other sites, including the channel segment at the downstream end of reach 2 (**Figure 8**).

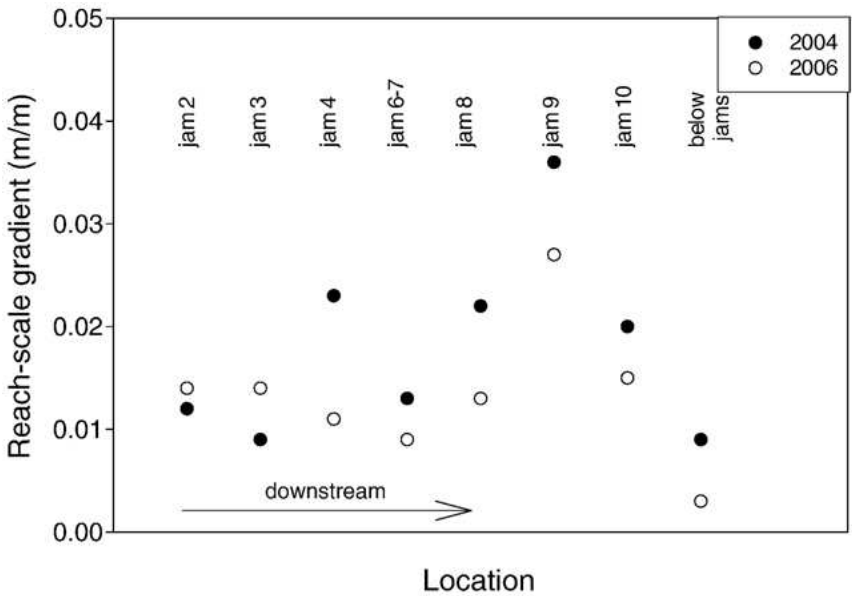

Figure 8. Changes in reach-scale water-surface gradient in the vicinity of logjams between 2004 and 2006 surveys. Distance along x-axis does not reflect relative downstream distance of logjams.

The magnitude of the change at each site varied, and this likely reflects local erosional and depositional processes as well as slight variations in discharge. Although the 2004 and 2006 surveys were done at about the same time of year, each field season extended over several days during which flows fluctuated in response to daily changes in air temperature and associated glacier melt and in response to short rain storms. The difference in direction of gradient change between the two upstream sites and the other sites may indicate a gradual downstream movement of sediment deposited within the channel during the jökulhlaup, but daily fluctuations in discharge between individual measurements prevent the thorough evaluation of this possibility.

DISCUSSION

Peak 2003 jökulhlaup discharge in reach 2 of the study area varied from being approximately ten times the size of normal seasonal peak flows at the upstream end of the reach to just over twice as large approximately 5 km downstream. This pattern of rapid downstream attenuation in magnitude of peak discharge is similar to those documented for other glacial outburst floods (e.g., Cenderelli and Wohl, 2001; Kershaw et al., 2005) and was associated with reach-scale changes in the spatial extent and grain-size distribution of overbank deposits.

The primary effect of the 2003 jökulhlaup on the valley bottom in reach 2 appears to have been overbank deposition. The flood does not seem to have had sufficient power to create observable or persistent overbank erosion except for localized scour features. Overbank and in-channel sediment deposition were influenced by logjams that appear to have formed as a result of bank erosion and increased recruitment of wood to the active channel during the jökulhlaup.

Volumes of in-channel deposition during the jökulhlaup and subsequent re-working during 2004 to 2006, however, do not show consistent longitudinal patterns analogous to those found in previous studies of channel response to rapidly introduced large volumes of sediment (Lisle and Hilton, 1992; Madej and Ozaki, 1996; Wohl and Cenderelli, 2000). This suggests that local channel and logjam configuration, and associated hydraulics, exerted a stronger influence than system-wide trends in sediment dynamics.

Much of the local influence on sediment deposition and subsequent re-working comes from the presence of logjams. The geomorphic influences of wood on hydraulics and sediment storage have been extensively documented by numerous investigators (Keller and Swanson, 1979; Abbe and Montgomery, 2003; Jeffries et al., 2003; MacFarlane and Wohl, 2003; O'Connor et al., 2003), including the effects of wood on channel processes during large floods caused by intense precipitation in the Appalachian Mountains (Williams and Guy, 1973; Kite and Linton, 1993; Hicks et al., 2005).

The latter studies document debris jams created by mass movement (Williams and Guy, 1973) and fluvial erosion (Hicks et al., 2005) of streambanks and floodplains and by associated entrainment of trees. Jams within the stream channel caused upstream aggradation and overbank flooding (Hicks et al., 2005) and enhanced bank erosion (Williams and Guy, 1973). Jams on the floodplain created localized upstream scour and downstream deposition (Kite and Linton, 1993).

These effects are similar to those that were observed along the route of the 2003 jökulhlaup. The only mention that could be found in the literature, however, of in-channel effects of wood during a jökulhlaup comes from Blown and Church (1985), who briefly described high-velocity overbank flows that uprooted valley-bottom trees and created large overbank accumulations of wood debris along the Nostetuko River in British Columbia, Canada.

The jokulhlaup increased the mobility of floatable wood in the channel. Floatable here refers to the scaling relation between length and diameter of wood and channel width and depth: pieces that are shorter than average channel width and of smaller diameter than average flow depth are more readily mobilized than longer or larger diameter wood pieces. With a large amount of floatable wood moving simultaneously in the channel, channel blockages can form simply by the presence of "congested flow" (Braudrick et al. 1997) where the interactions of large quantities of wood in transport cause the wood to accumulate and become jammed. Where the jökulhlaup increased bank erosion and recruitment of whole trees into the channel, ramps and bridges formed with one or both ends of the tree on the stream margins. This channel-spanning wood could accumulate smaller floatable wood *in situ* or a short distance from treefall where the log became hung up on an obstacle in the channel. With the increased large wood recruitment and smaller wood mobility due to the jökulhlaup, large and more closely spaced logjams were formed.

The formation of in-channel logjams and associated in-channel and overbank deposition that were described along reach 2 of Downs Fork undoubtedly occur along other forested mountain valleys that are subject to jökulhlaups. The fact that much of the overbank and in-channel deposition within reach 2 occurred only in association with logjams strongly suggests that the formation of logjams exerts a critical influence on sediment dynamics along this portion of the jökulhlaup route.

Logjams might be expected to persist for decades in the relatively dry climate of the study area. Conversations with elderly residents of nearby communities indicate that channel-spanning logjams were present along Downs Fork half a century ago but that backcountry outfitters removed these jams to make it easier to cross the channel on horseback. These earlier logjams might have been formed during previous jökulhlaups from Grasshopper Glacier because jökulhlaups may occur repeatedly from a single glacier, and glaciers throughout the Wind River Range have been retreating for decades (Naftz et al., 2002), thus creating the meltwater necessary to supply repeated jökulhlaups.

Coarse sediments that likely resulted from overbank deposition during pre-2003 jökulhlaups suggest that meltwater floods may create persistent influences on valley-bottom geometry along Downs Fork. Extensive longitudinal and expansion bars of cobble- to boulder-sized sediment (approximately 0.3 to 1 m diameter) are present in the wider sections of valley at the transition between study reaches 1 and 2 (**Figure 9**).

Although neither the age of these boulder deposits nor the number of depositional events that occurred are certain, individual boulders are well-covered with lichens and partially buried in soil, and conifers at least a century in age (with diameters up to 1.2 m at breast height) grow on top of portions of the boulder bars. The older coarse-grained jökulhlaup deposits have a characteristic undulating bar and swale topography that serves to distinguish the deposits from the glacial terraces at the upstream valley margins of reach 2.

Terraces have a generally planar surface with an abrupt bench toward the present stream channel. Terraces also tend to be higher on the cutbank side of the meandering stream, whereas depressions in the terrace's point bar deposits form a swale closer to the valley margin. Longitudinal differences in elevation of the valley bottom associated with terrace

Figure 9. (Upper) Looking upstream along right valley margin at head of reach 2 to partially buried longitudinal and expansion bars created by pre-2003 jökulhlaups. Individual bars are 5 to 10 m across. **(Lower)** View downstream along the crest of a boulder bar at the head of reach 2 created by pre-2003 jökulhlaup. Large clasts in foreground have intermediate diameters of 0.5 to 0.8 m.

morphology and older jökulhlaup sediments determined where the banks were overtopped along the 2003 jökulhlaup route by the backwater formed upstream from the logjams.

Older, fine-grained jökulhlaup deposits are likely present as well but are difficult to distinguish from contemporary overbank deposition in the terraces of the downstream areas of study reach 2.

The presence of older overbank deposits similar to those created during the 2003 jökulhlaup indicates the influence that these high-magnitude floods exert on the valley bottom of Downs Fork over longer time spans. Reach 2 is the first wide, lower gradient segment of the valley downstream from the glacier. It thus seems likely that any large volumes of sediment originating from the glaciated headwaters would be at least temporarily deposited in this zone of lower transport capacity.

Deposition may be enhanced by the fact that narrow, steep portions of the valley downstream create hydraulic constrictions that facilitate backwater effects and sediment deposition during very large floods capable of filling the narrow portions from valley-wall to valley-wall. No evidence in terms of either post-2003 erosion or deposition, or indirect calculations of peak discharge magnitudes, were found that normal snow- and glacial-melt peak flows go overbank along Downs Fork. Consequently, the only means by which normal peak flows may rework valley-bottom deposits is through lateral channel migration or channel avulsion.

Although aerial photographs suggest that channel avulsion occurred in reach 3 even before the 2003 jökulhlaup, no indication that avulsion was common along other portions of Downs Fork was found. Lateral channel migration is limited by the extensive forest cover.

The presence of conifers at least a century in age growing in close proximity to the active channel suggests that lateral channel migration is of limited occurrence, and longitudinal variations in valley width and gradient associated with lateral bedrock confinement of the active channel indicate that any lateral channel migration would also be of limited extent longitudinally.

Little evidence of debris flows or landslides along Downs Fork was found. The resistant crystalline rocks outcropping along the valley in the study area produce rockfalls and talus slopes along the valley margins but appear to contribute relatively little sediment of gravel size or finer. The talus slopes also commonly cover <20% of the total valley-bottom width. Finer sediment present along the valley bottom in the study area thus appears to result primarily from fluvial rather than hillslope processes.

In addition to sediment delivery to streams, debris flows, landslides, and fire history also affect wood recruitment to streams. The high-elevation forests of the Dinwoody drainage most likely have a mean fire recurrence interval of 150 to 700 years for stand-replacing fires, and in the historic interval, a mean fire recurrence interval for small fire events of 10 to 300 years (Rice et al., 2012). Small fires, both lightning- and human-caused, are evident in the Dinwoody drainage and have contributed to localized wood input to streams. Although there are insufficient data to quantify the Downs Fork fire history, the larger stand-replacing fires could contribute up to 15% of the wood budget and associated large wood recruitment to streams (Benda and Sias, 2003).

The combined evidence of limited overbank deposition during normal seasonal peak flows, limited lateral re-working of overbank deposition by lateral channel migration during normal seasonal peak flows, and limited contribution of fine sediment from hillslope processes support the third hypothesis: that the presence of logjams facilitated significant erosional and depositional effects of greater magnitude and persistence than the effects created by normal snowmelt peak discharges.

The results suggest that relatively infrequent jökulhlaups exert a substantial influence on valley-bottom geometry through substantial overbank deposition along segments of slightly wider and gentler gradient valley bottom. A conceptual model was developed to illustrate the jökulhlaup effects on channel and valley processes (**Figure 10**). In this model, a jökulhlaup is geomorphically effective when its peak discharge exceeds the conveyance of the channel, which reflects flow history and boundary resistance.

The extent to which conveyance is exceeded, and overbank flow occurs, reflects distance downstream, channel and valley geometry, and wood dynamics (recruitment into the channel and transport or storage within the channel). Wood dynamics become increasingly important farther downstream because, as the ratio of jökulhlaup peak discharge to average annual peak discharge decreases, logjams are relatively more important in forcing overbank flow and enhancing local erosional and depositional effects during the jökulhlaup.

The sequence of erosional and depositional events illustrated in **Figure 10** could certainly occur during a rainfall- or snowmelt-generated flood along a forested mountain valley, but the greater magnitude of the jökulhlaup likely enhances localized bank erosion and the formation of relatively large, closely-spaced logjams.

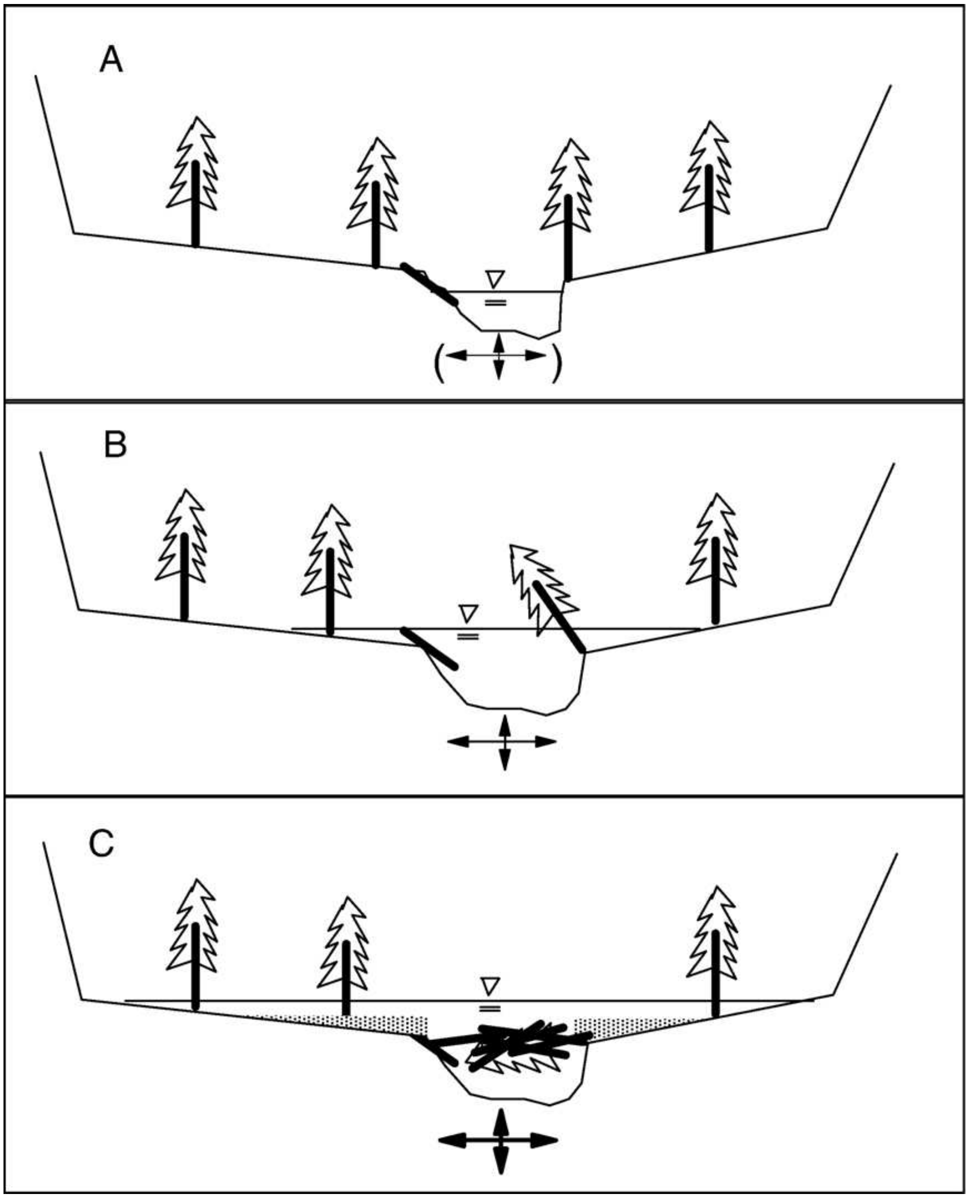

Figure 10. (A) Annual snowmelt peaks are contained within the bankfull channel. Individual annual peaks create relatively little erosion and deposition along channel banks and bed (arrows below channel), but repeated occurrence year after year can create channel change. Relatively little wood is recruited into the channel and logjams do not completely span the channel. **(B)** During a jökulhlaup, overbank flow occurs locally. Shorter duration but higher magnitude flow creates greater magnitude of change along the channel boundaries, as well as affecting overbank areas. **(C)** More wood is recruited into the channel during a jökulhlaup. Wood accumulating in logjams can enhance local erosional and depositional change in the channel (larger arrows below channel). Logjams can also cause channel aggradation and backwater, locally forcing flow across the bankfull threshold and onto the floodplain, where overbank deposition (gray shading) results. The effect of logjams is most pronounced where the jökulhlaup might not create overbank flow in the absence of logjams.

The results from this study have particularly important implications for valley evolution over hundreds to thousands of years because of the possibility for variations in the magnitude and frequency of jökulhlaups. Jökulhlaups may occur when a glacier is advancing, retreating, or is stable (Tweed and Russell, 1999).

The greater volumes of meltwater generated during periods of glacial retreat, however, certainly enhance the potential for frequent jökulhlaups. If jökulhlaups increase in frequency during periods of glacial retreat, such as that following the Neoglacial or the Little Ice Age, then these time intervals could represent periods of enhanced valley-bottom deposition and sediment storage, followed by either stability or gradual downstream transport of stored sediment during intervening periods with fewer outburst floods.

Current predictions of the complete melting of alpine glaciers within the continental United States and in many other regions of the world during the next few decades suggest that we are entering a period likely to be characterized by greater jökulhlaup activity and associated sediment movement but that valley bottoms may then become more stable or degrade once the glaciers have completely melted.

The geomorphic effects of the jökulhlaup, and specifically of logjams during the jökulhlaup, are similar to those described for high-magnitude rainfall floods in other forested mountainous regions. The possibility of repeated occurrence of jökulhlaups as mountain glaciers retreat is thus analogous to a scenario in which changes in storm frequency or location produce unusually frequent high-magnitude floods, except that the jökulhlaup scenario requires only continued warming of air temperatures.

CONCLUSIONS

The basin-scale geomorphic effects of the 2003 jökulhlaup reflected distance from the outburst source and valley geometry. Deposition grew progressively finer downstream, and confined, steeper valley reaches had less deposition, whereas the lowest gradient, least confined reaches had extensive overbank deposition and formation of multiple channels. Within reach 2, the combination of moderate valley confinement and gradient, readily-eroded banks, and extensive forest cover in the valley bottom created a situation in which the higher discharge of the jökulhlaup caused overbank flow and deposition and enhanced bank erosion and recruitment of wood into the channel.

Limited mobility of wood introduced during the flood initiated numerous logjams along the 5 km of reach 2. The logjams created backwater effects and enhanced overbank flow and bank erosion, as well as inducing local upstream sediment deposition.

In the three years after the jökulhlaup, subsequent flows did not modify the overbank deposits and caused only minor changes in logjam configuration and sediment storage within the channel. These observations, along with the presence of much older jökulhlaup deposits on terraces along the valley margins, suggest that periodic jökulhlaups play an important role in persistent valley-bottom aggradation along Downs Fork.

Downs Fork is unlikely to be unique with respect to other forested catchments with headwater glaciers in mountains throughout the western United States and other regions of the world. This thesis documents a sequence of events in which outburst floodwaters enhance bank erosion and recruitment of wood into the channel and thus the formation of large logjams. These logjams sufficiently deflect flow to create substantial overbank deposition in areas of the valley bottom not commonly accessed by normal snowmelt peak discharges and through this process promote valley-bottom aggradation and sediment storage.

Changes in the occurrence of glacier outburst floods caused by climatic warming or other processes thus have the potential to significantly influence the rate and magnitude of valley-bottom dynamics in these environments. This is particularly relevant given current predictions of global warming and the retreat of headwater glaciers around the world.

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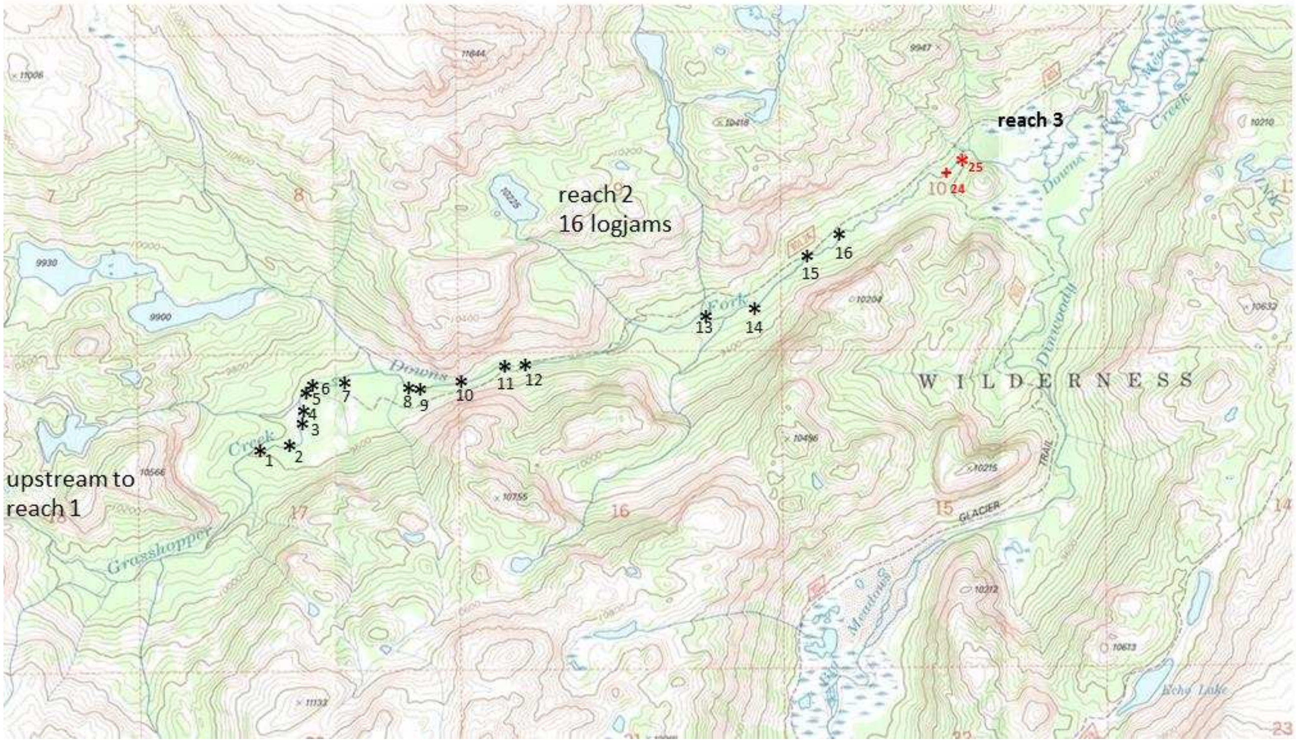
APPENDIX I

COMPARISON OF CROSS-SECTIONAL PROFILES SURVEYED IN 2004

AND RE-SURVEYED IN 2006



Map showing location of logjams in reach 2. Map base is Ink Wells and Fremont Peak North USGS 1:24,000 topographic quadrangle maps, contour interval 40 feet. Section lines in red show 1 mile scale (1 mile = 1.6 km).

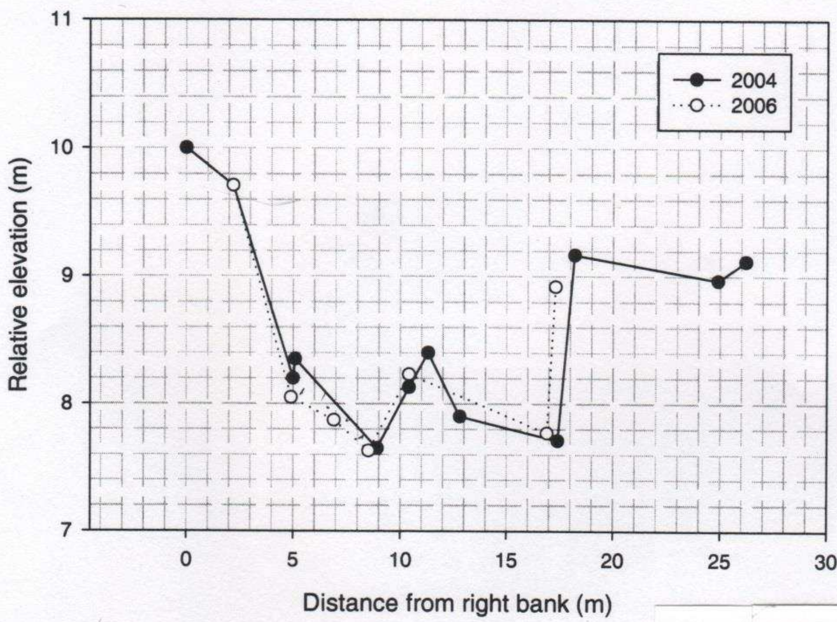
Key to location of Appendix cross sections.

Xs 1 above logjam 3 Xs 2 below logjam 3	Xs 7 above logjam 6 Xs 8 below logjam 6	Xs 13 above logjam 9 Xs 14 below logjam 9	Xs 20 above logjam 16 Xs 21 above logjam 16
Xs 3 above logjam 4 Xs 4 at logjam 4	Xs 9 above logjam 7 Xs 10 at logjam 7	Xs 15 above logjam 10 Xs 16 above logjam 10	Xs 22 below logjam 16 Xs 23 below logjam 16
Xs 5 at logjam 5 Xs 6 at logjam 5	Xs 11 above logjam 8 Xs 12 below logjam 8	Xs 18 above logjam 15 Xs 19 below logjam 15	Xs 24 above bridge (+24) Xs 25 above bridge (*25)

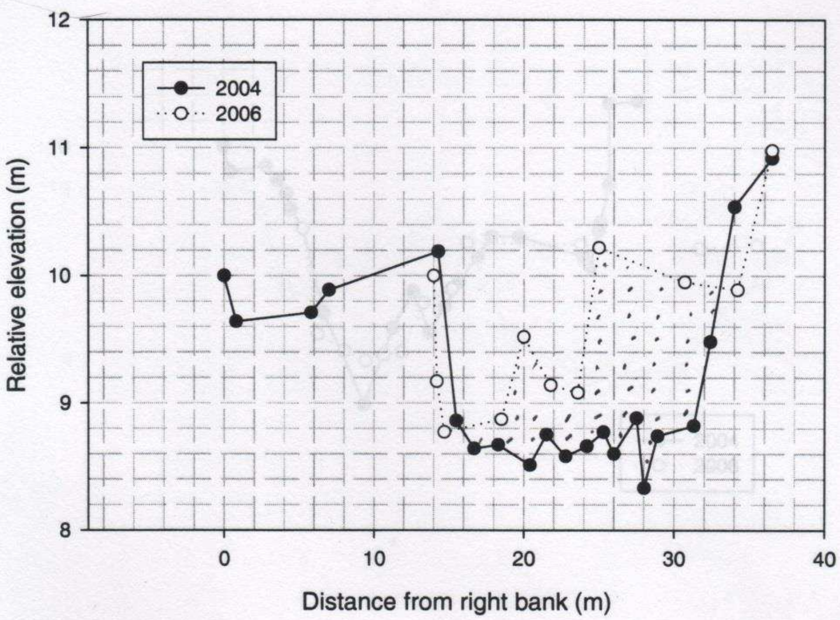
The appendix contains 24 cross sections of reach 2 surveyed in 2004 and 2006. Cross sections are referenced in relation to one of 16 logjams. Both logjams and cross sections have their own numbering sequences, which increase in a downstream direction through the length of reach 2. The map designates logjams with an * and a number. Cross sections are numbered Xs 1 through 25, and their locations are shown in the table above in relation to logjam number. The location of two cross sections not associated with logjams are denoted +24 and *25 at the downstream end of reach 2.

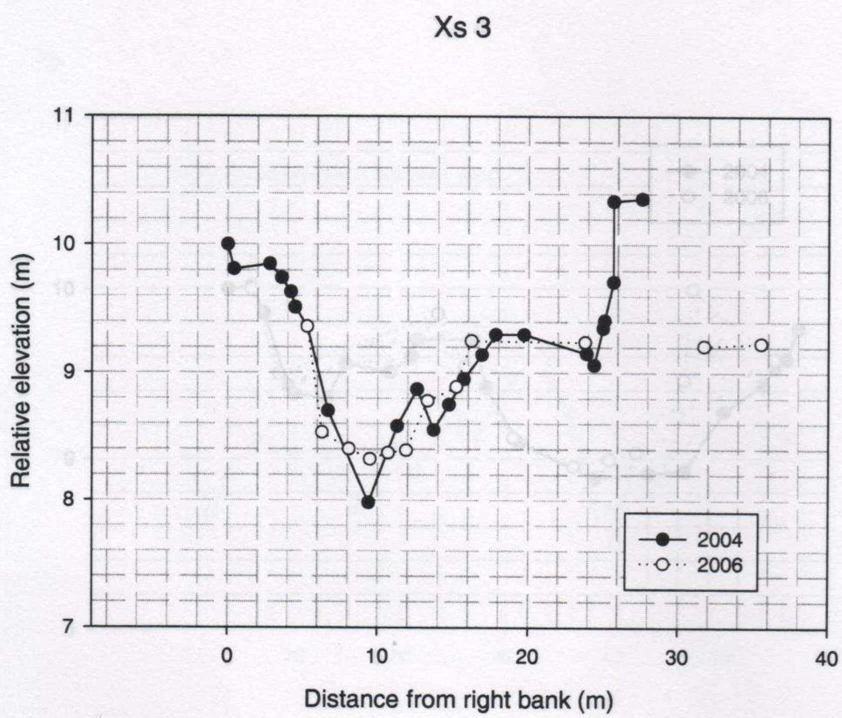
Cross sections were not monumented in the field due to wilderness regulations. This made re-location of end points and orientation of cross sections imprecise using GPS locations, tape and compass, photographs, and longitudinal profile data from the 2004 field season. Some of the comparison graphs in the Appendix have the notation *match uncertain between years* when the location for the 2006 re-survey was less certain. Cross section 17 was not re-surveyed in 2006.

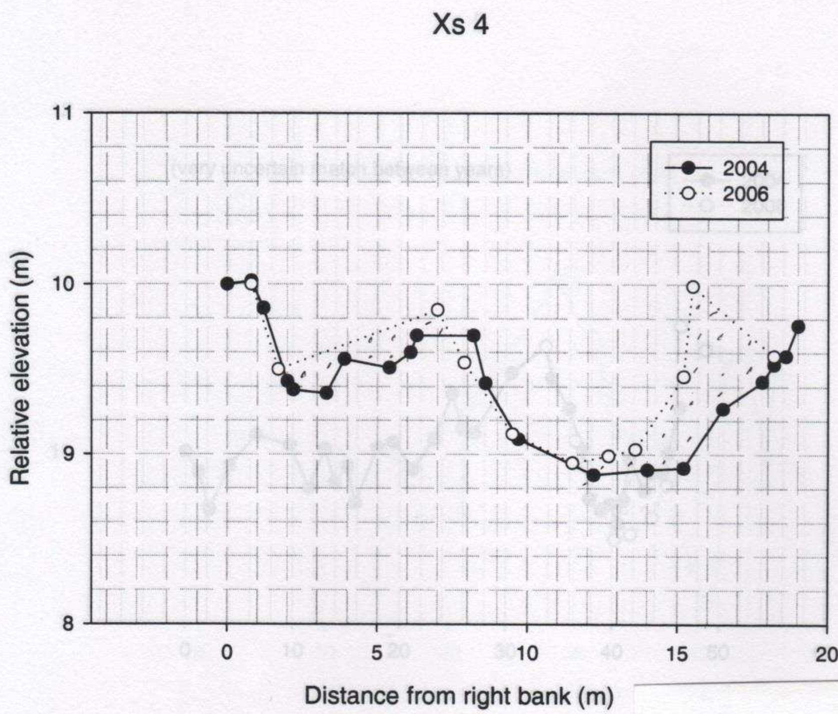
Xs 1


2004
2006
11
10
9
8
7
Relative elevation (m)
0
5
10
15
20
25
30
Distance from right bank (m)

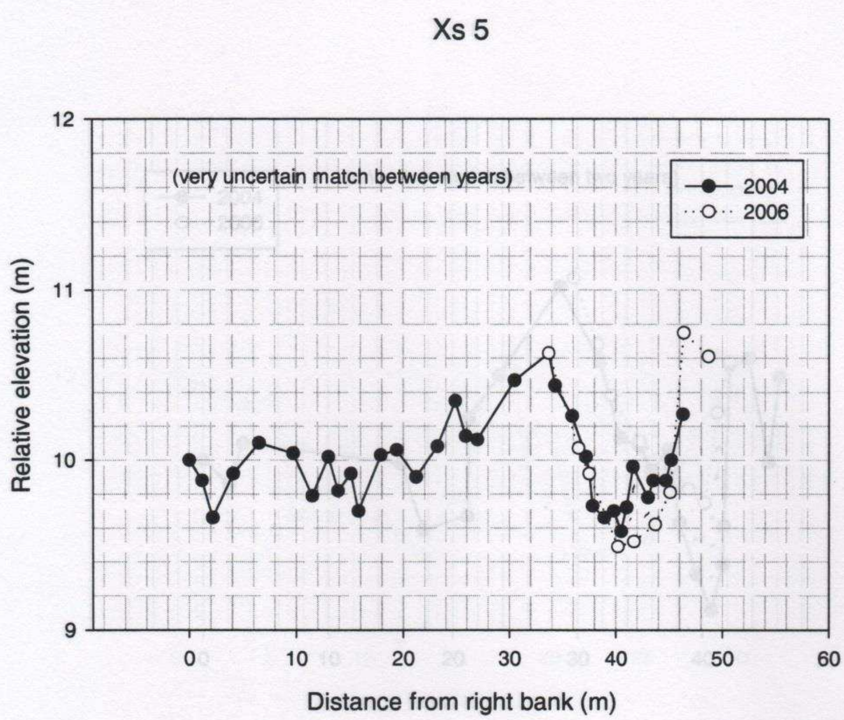
Xs 2


2004
2006
Relative elevation (m)
12
11
10
9
8
0
10
20
30
40
Distance from right bank (m)

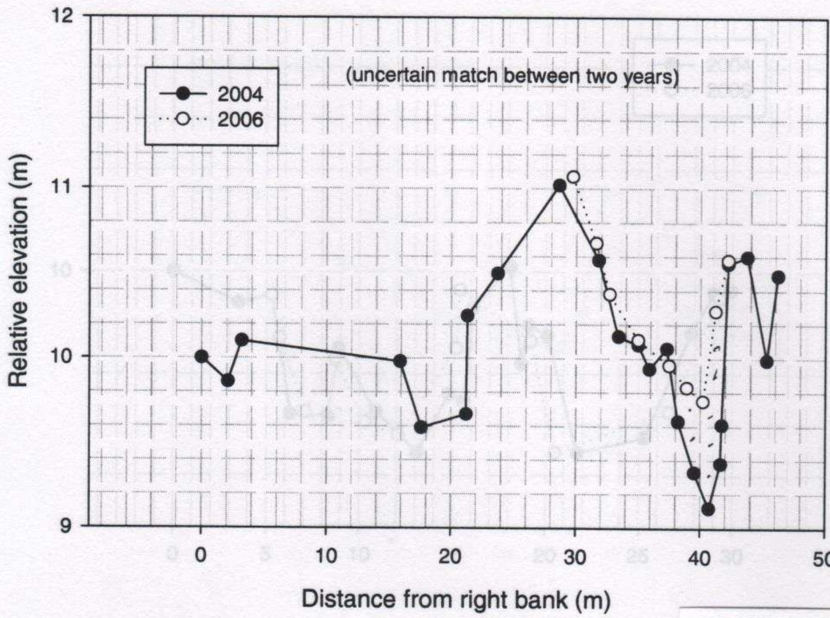

Xs 3
11
10
9
8
7
0
10
20
30
40
Relative elevation (m)
Distance from right bank (m)
2004
2006



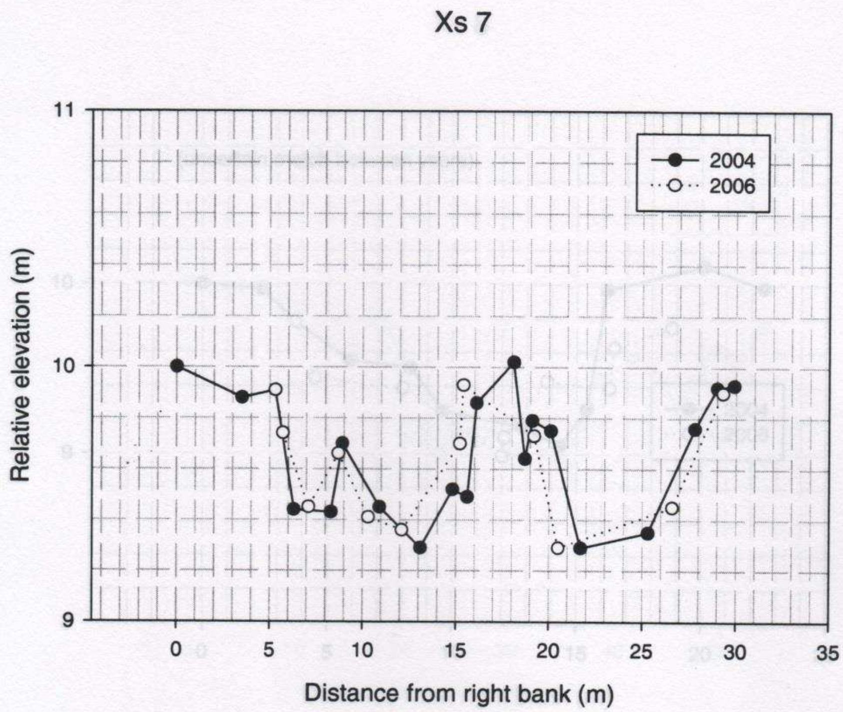
Xs 4
11
10
9
8
Relative elevation (m)
0
5
10
15
20
Distance from right bank (m)
2004
2006


Xs 5
(very uncertain match between years)
2004
2006
Relative elevation (m)
12
11
10
9
0
10
20
30
40
50
60
Distance from right bank (m)

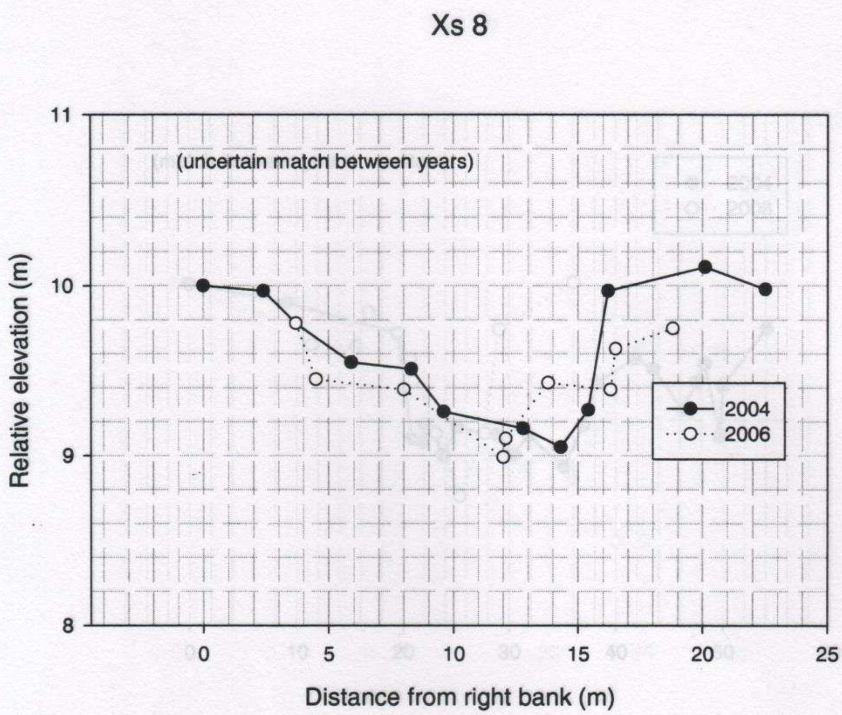
Xs 6

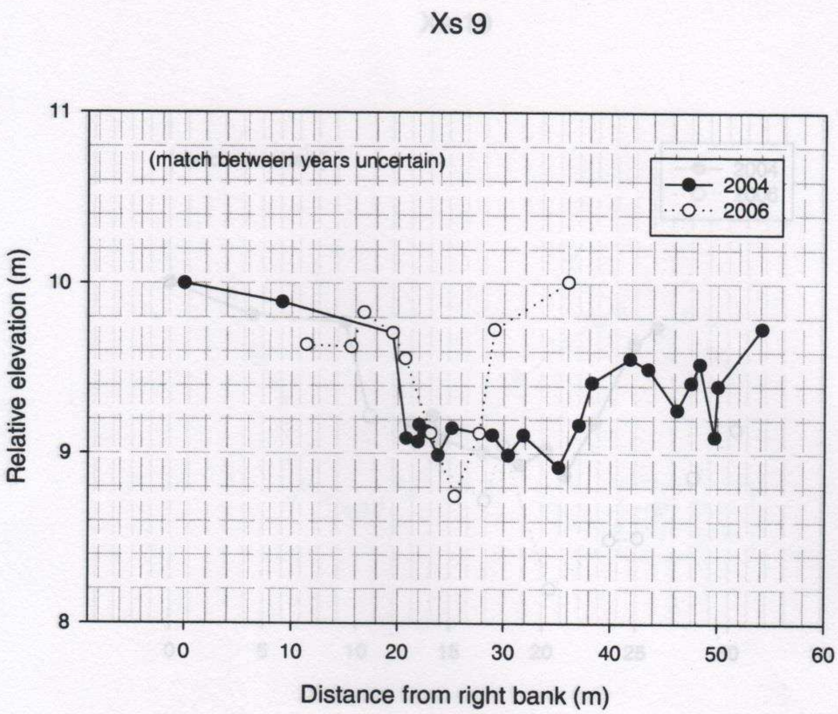


(uncertain match between two years)
2004
2006
Relative elevation (m)
12
11
10
9
0
10
20
30
40
50
Distance from right bank (m)

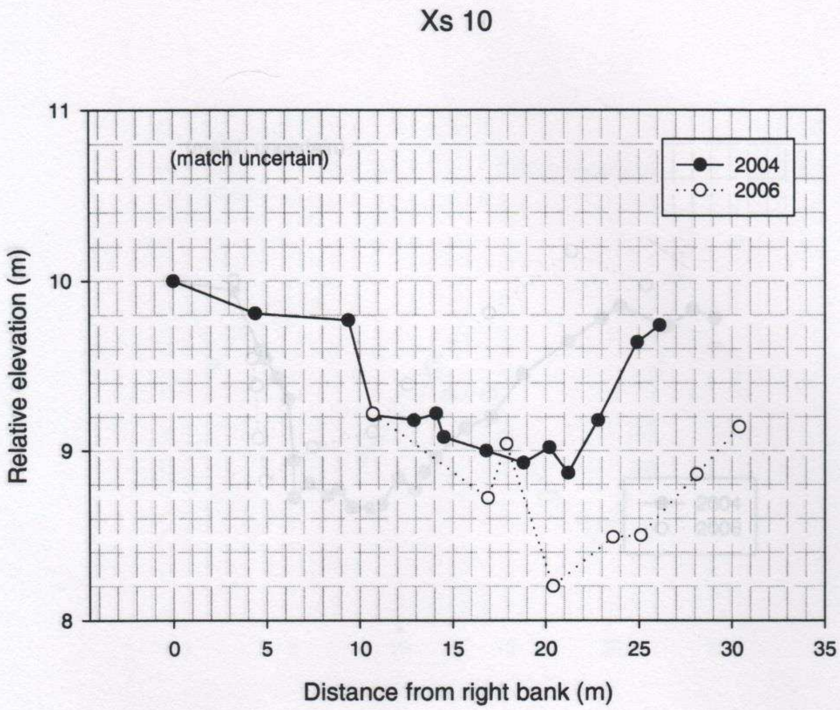


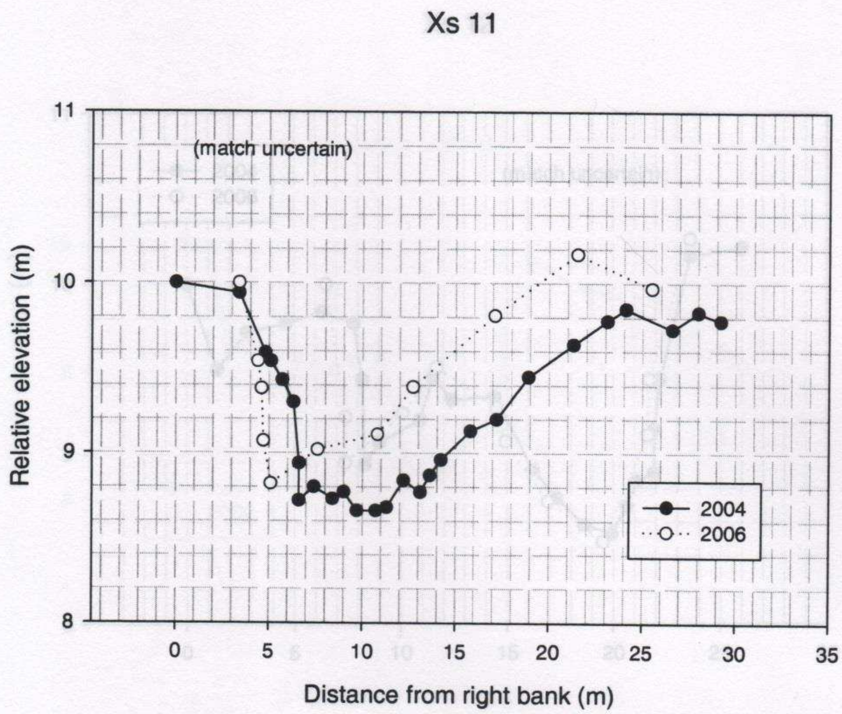
Xs 7
11
10
9
Relative elevation (m)
0
5
10
15
20
25
30
35
Distance from right bank (m)
2004
2006


Xs 8
(uncertain match between years)
Relative elevation (m)
Distance from right bank (m)
2004
2006
11
10
9
8
0
5
10
15
20
25



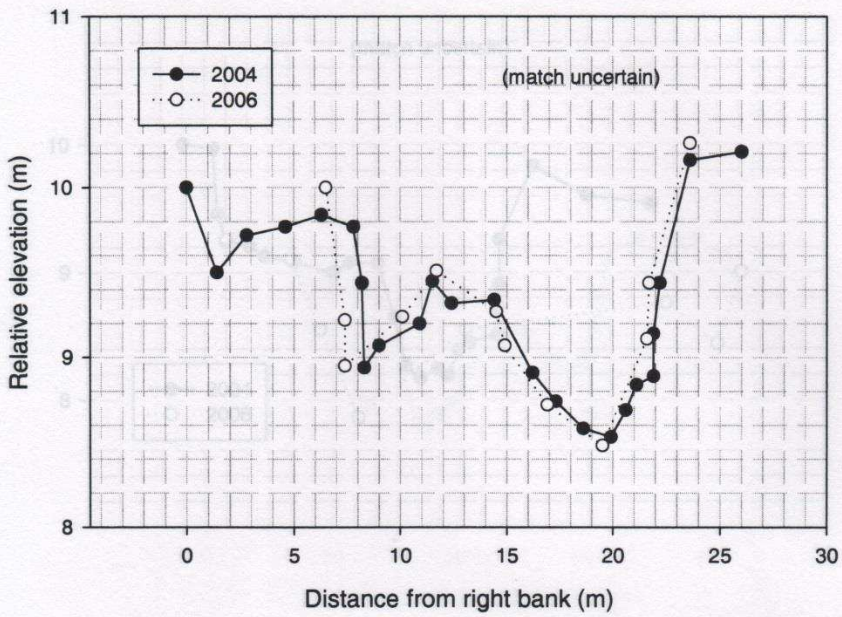
Xs 9
(match between years uncertain)
2004
2006
Relative elevation (m)
11
10
9
8
0
10
20
30
40
50
60
Distance from right bank (m)


Xs 10
(match uncertain)
2004
2006
Relative elevation (m)
Distance from right bank (m)
11
10
9
8
0
5
10
15
20
25
30
35

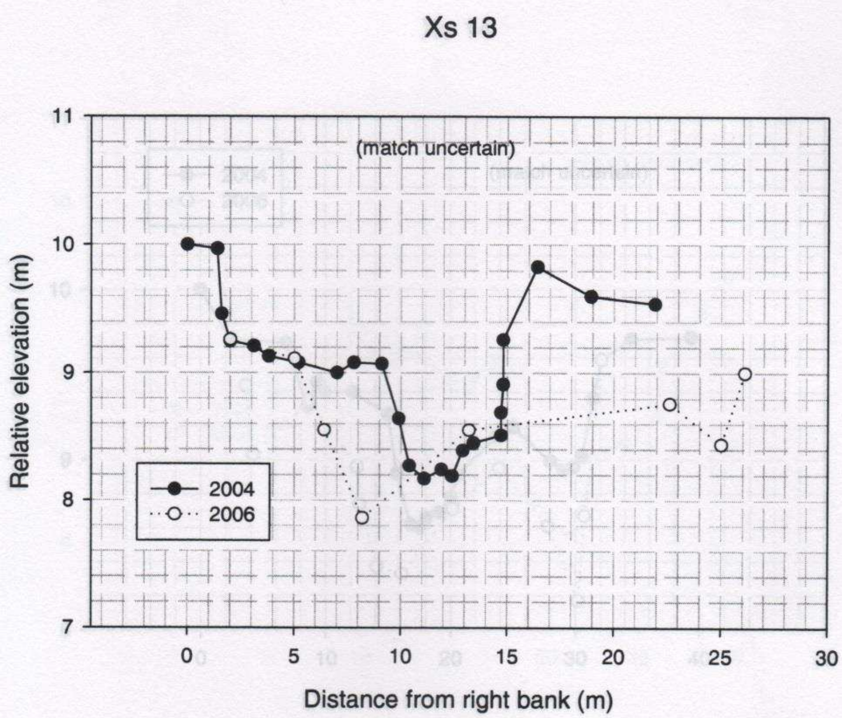


Xs 11
(match uncertain)
Relative elevation (m)
11
10
9
8
0
5
10
15
20
25
30
35
Distance from right bank (m)
2004
2006

Xs 12

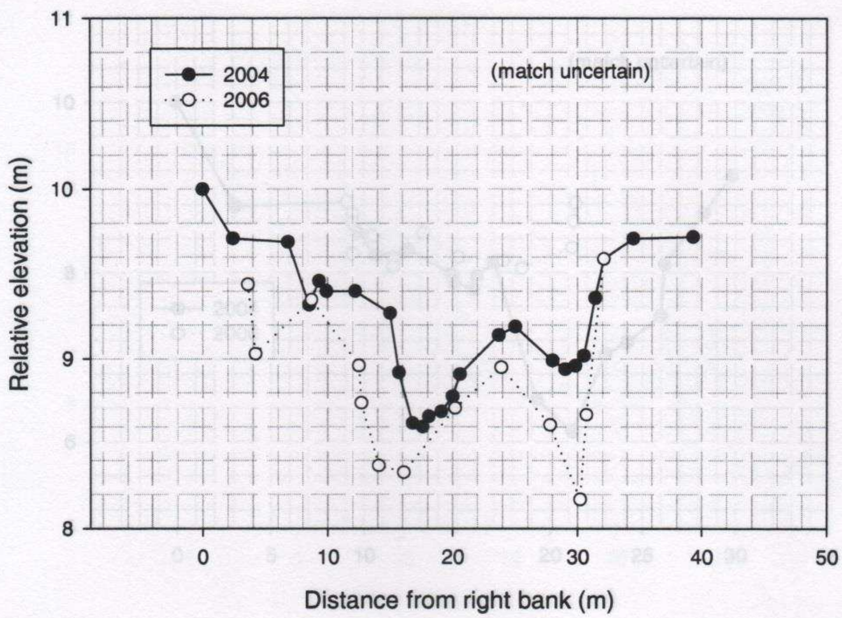


2004
2006
(match uncertain)
Relative elevation (m)
11
10
9
8
0
5
10
15
20
25
30
Distance from right bank (m)

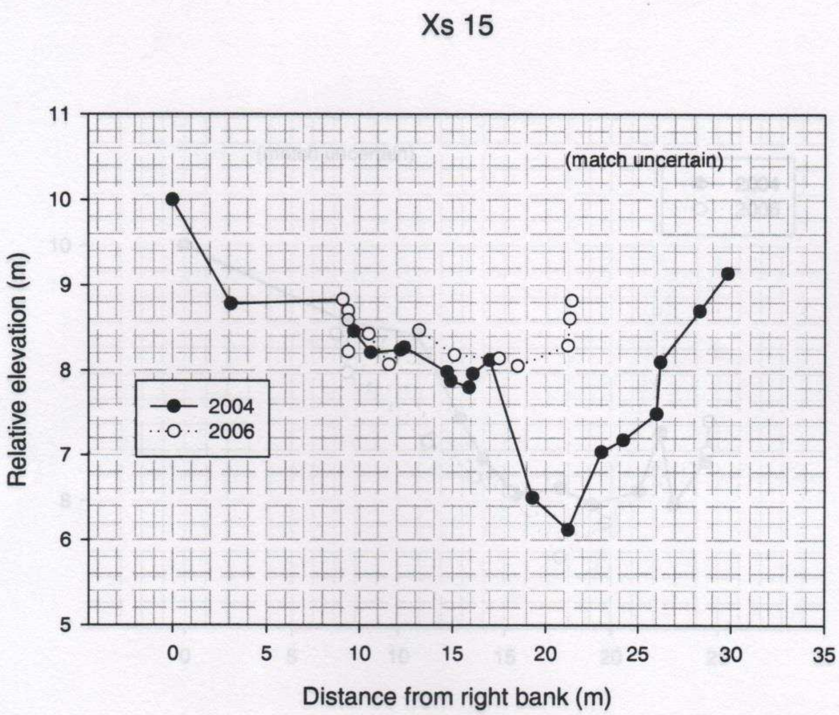


Xs 13
(match uncertain)
2004
2006
Relative elevation (m)
Distance from right bank (m)
11
10
9
8
7
0
5
10
15
20
25
30

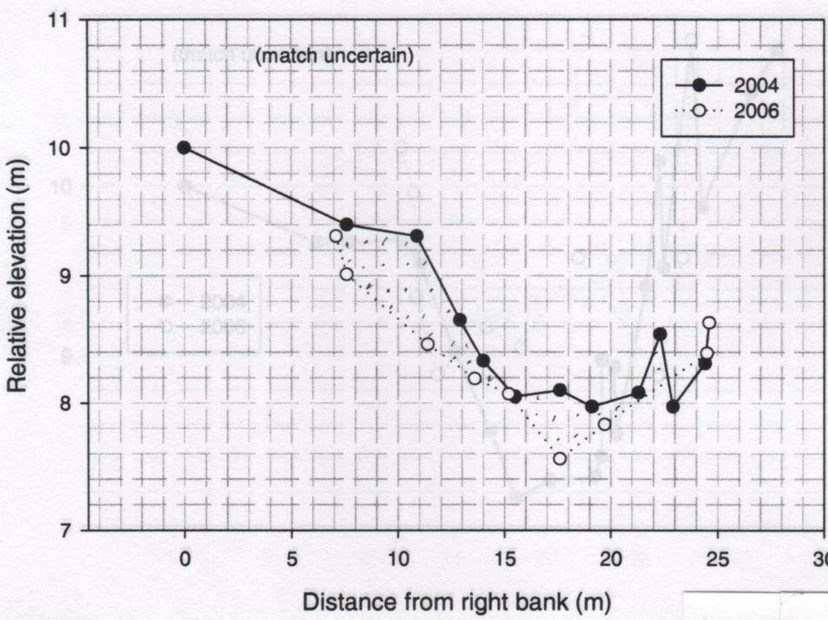
Xs 14



2004
2006
(match uncertain)
Relative elevation (m)
11
10
9
8
0
10
20
30
40
50
Distance from right bank (m)

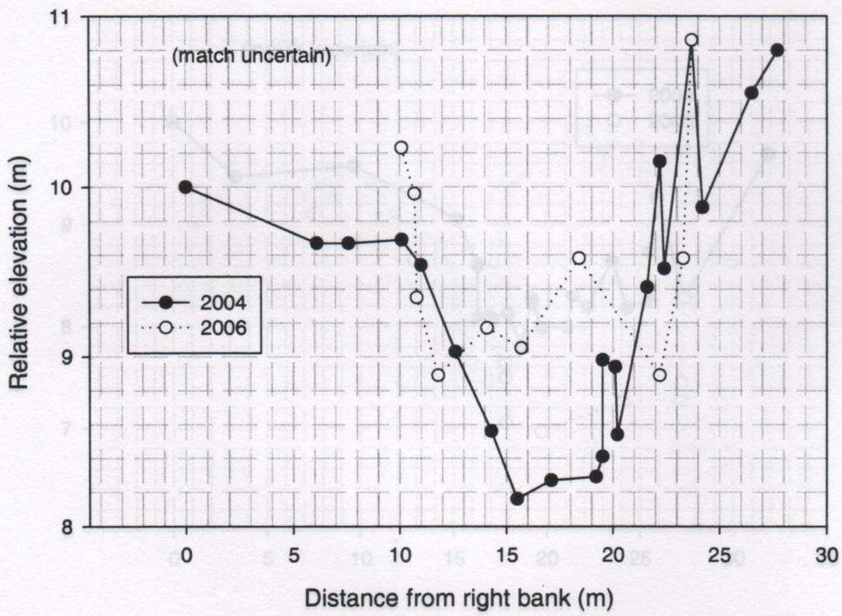

Xs 15
(match uncertain)
Relative elevation (m)
Distance from right bank (m)
2004
2006
11
10
9
8
7
6
5
0
5
10
15
20
25
30
35

Xs 16



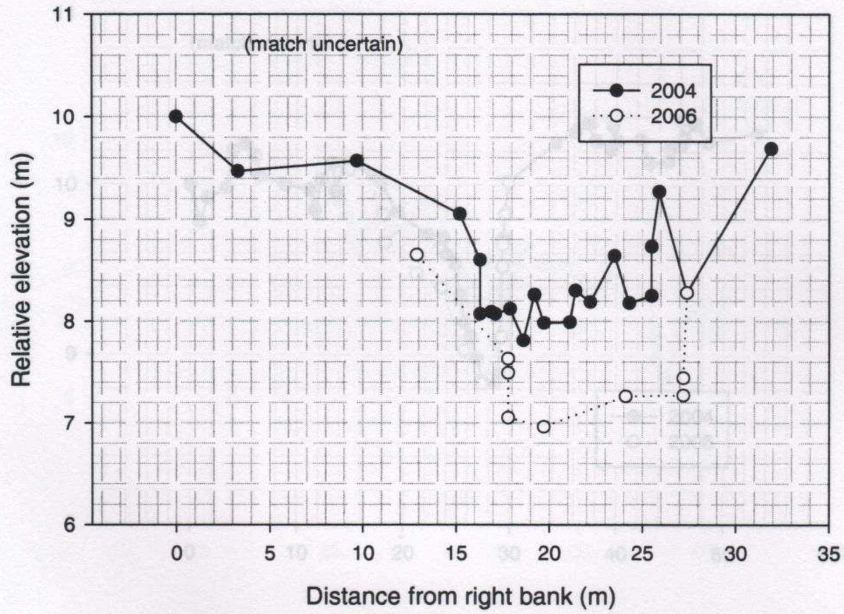
(match uncertain)
2004
2006
Relative elevation (m)
11
10
9
8
7
0
5
10
15
20
25
30
Distance from right bank (m)

Xs 18

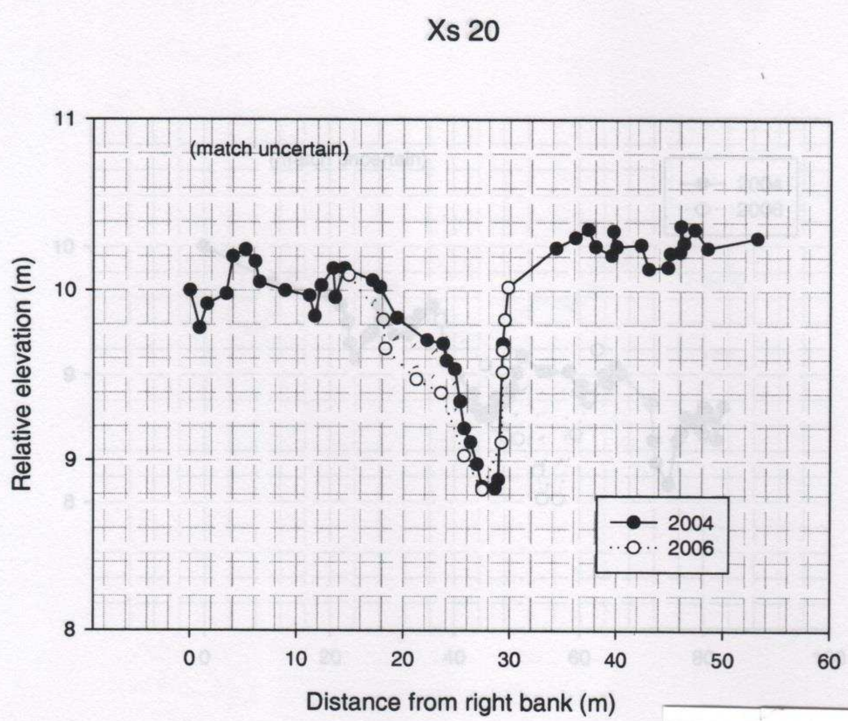


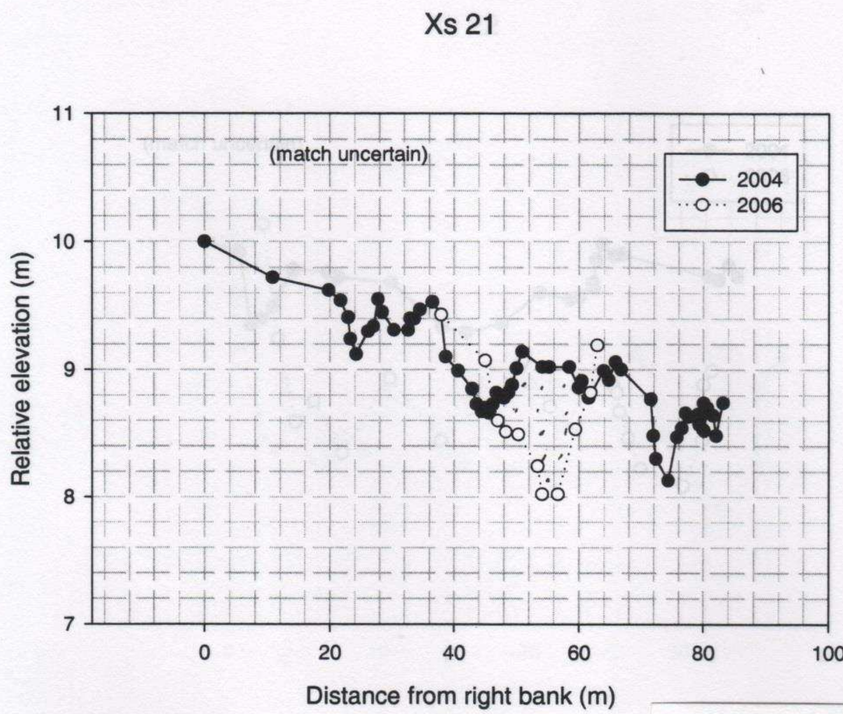
(match uncertain)
2004
2006
Relative elevation (m)
Distance from right bank (m)
11
10
9
8
0
5
10
15
20
25
30

Xs 19

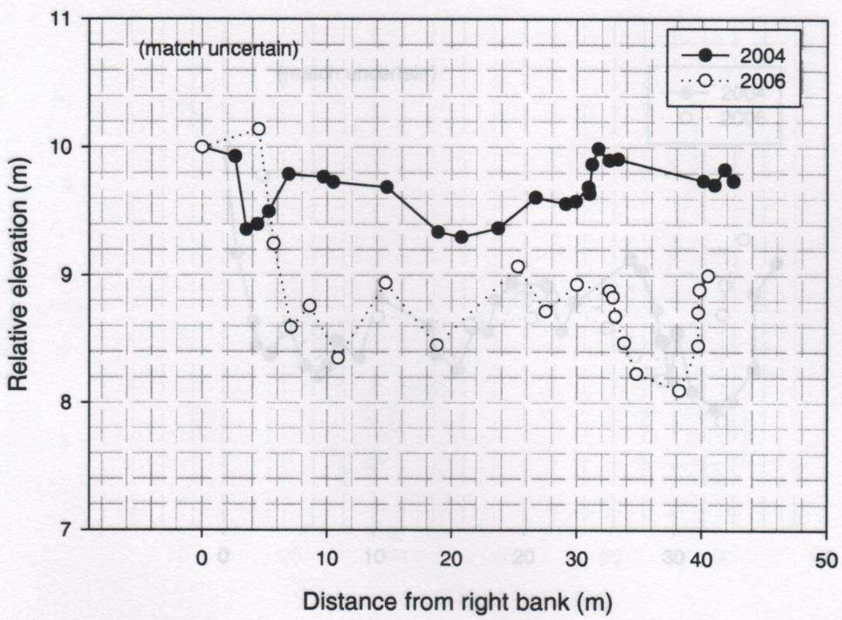


(match uncertain)
2004
2006
Relative elevation (m)
11
10
9
8
7
6
0
5
10
15
20
25
30
35
Distance from right bank (m)


Xs 20
(match uncertain)
Relative elevation (m)
11
10
9
8
0
10
20
30
40
50
60
Distance from right bank (m)
2004
2006

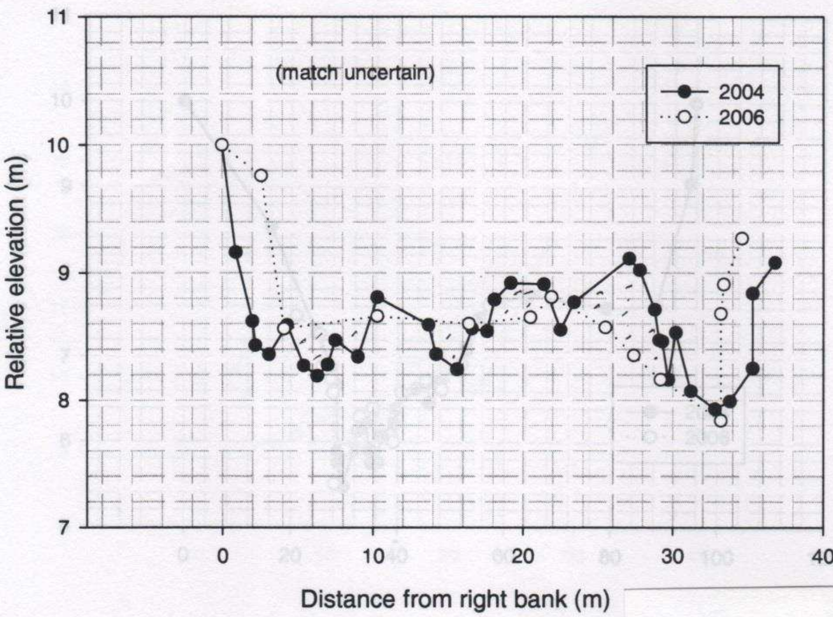

Xs 21
(match uncertain)
2004
2006
Relative elevation (m)
11
10
9
8
7
0
20
40
60
80
100
Distance from right bank (m)

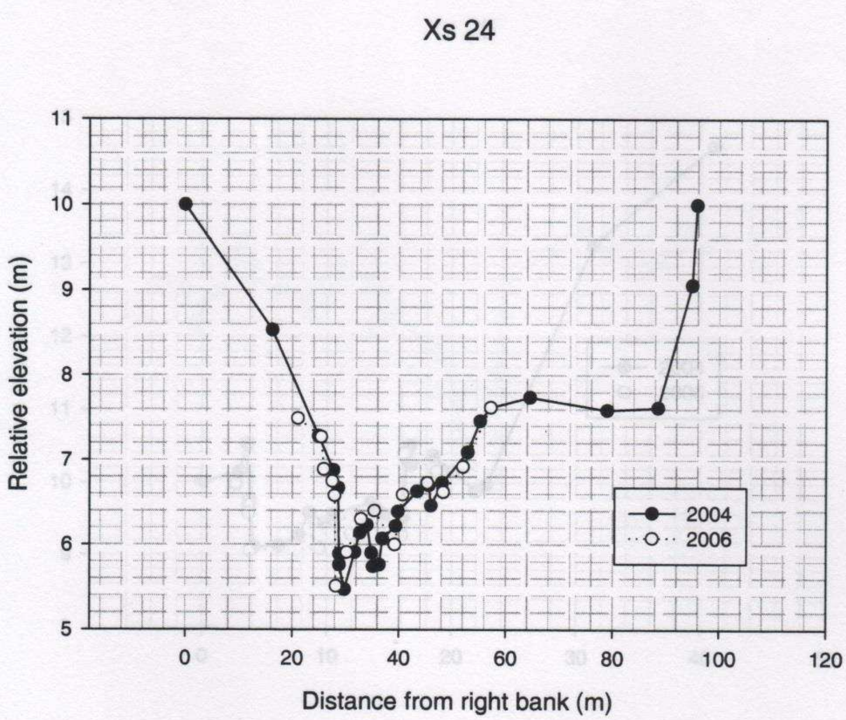
Xs 22



(match uncertain)
2004
2006
Relative elevation (m)
11
10
9
8
7
0
10
20
30
40
50
Distance from right bank (m)

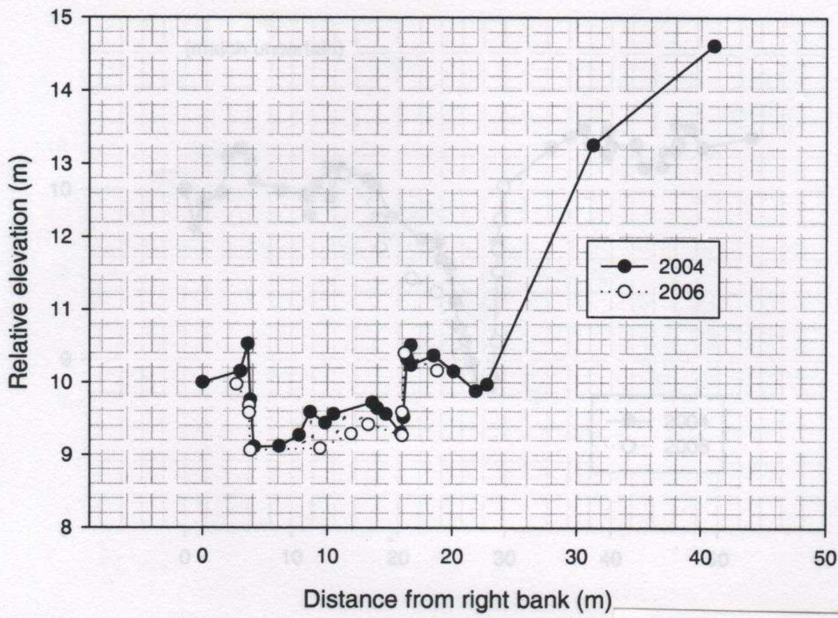
Xs 23


(match uncertain)
2004
2006
Relative elevation (m)
11
10
9
8
7
0
10
20
30
40
Distance from right bank (m)



Xs 24
Relative elevation (m)
11
10
9
8
7
6
5
0
20
40
60
80
100
120
Distance from right bank (m)
2004
2006

Xs 25


2004
2006
Relative elevation (m)
Distance from right bank (m)
15
14
13
12
11
10
9
8
0
10
20
30
40
50

APPENDIX II

AIR PHOTOS TAKEN ON 19 SEPTEMBER 2003 FROM LIGHT AIRCRAFT

OVER JÖKULHLAUP FLOOD ROUTE

View looking west to ice-dammed lake at the head of Grasshopper Glacier.

View looking down on northward-trending Grasshopper Glacier.

View looking downstream to the east of reach 1 on Grasshopper Creek.

View to south looking down on emergence of flood at treeline near the head of reach 2 on the Downs Fork.

View overlooking reach 3 in Downs Fork Meadows.

Overview looking toward east from Downs Fork Meadows to lower Dinwoody drainage.

View to south overlooking Mud Lake in reach 5.

Overview of Lower Dinwoody lakes in reach 5 and downvalley to Wind River floodplain.