# Contrasting bedrock incision rates from snowmelt and flash floods in the Henry Mountains, Utah

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## ABSTRACT

**Hydrograph variability and channel morphology influence rates of fluvial bedrock incision, but little data exist on these controls in natural channels. Through field monitoring we demonstrate that (1) short-term bedrock channel incision can be rapid, (2) sustained floods with smaller peak discharges can be more erosive than flash floods with higher peak discharges, due to changes in bed alluviation, and (3) bedrock channel morphology varies with local bed slope and controls the spatial distribution of erosion. We present a three-year record of flow depths and bedrock erosion for a human-perturbed channel reach that drains the Henry Mountains of Utah, USA. Starting from a small and steep (~30% slope), engineered knickpoint in Navajo sandstone, erosion has cut a narrow, deep, and tortuous inner channel in ~35–40 years. Along the inner channel, we measured up to 1/2 m of vertical incision into Navajo sandstone over ~23 days, caused by the 2005 season of exceptional snowmelt flow. In contrast, flash floods caused little bedrock incision even when peak discharges were much higher than the peak snowmelt flow. Flash floods were net depositors of coarse sediment while snowmelt flow cleared alluvial cover. We document the formation of a pothole and interpret that it was abraded by bedload rather than fine suspended sediment. Finally, several slot canyons (Peek-a-boo, Spooky, and Coyote Gulch narrows) in the nearby Escalante River drainage basin have erosional morphologies similar to the monitored channel reach. Feedbacks between flow, sediment transport, and transient erosion provide a plausible explanation for the evolution of channel slope, width, and bed roughness of these natural bedrock channels.**

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## INTRODUCTION

Do large floods, with greater stream power and boundary shear stresses, cause more bedrock channel erosion than smaller flow events? Understanding how the magnitude of floods influences bedrock erosion rates and patterns is central to understanding and predicting channel and landscape erosion rates and their sensitivity to climate over short and long time scales. Magnitude-frequency tradeoffs in the erosion of bedrock channels (e.g., Baker, 1988; Wohl, 1993; Howard, 1998; Lague et al., 2005; Sklar and Dietrich, 2006) are likely different from alluvial channels (e.g., Mackin, 1948; Wolman and Leopold, 1957; Andrews, 1984; Emmet and Wolman, 2001). In alluvial channels, morphologic adjustment occurs through the entrainment and deposition of sediment. In incising bedrock channels, morphology changes not only through sediment redistribution but additionally requires the detachment of bedrock by a variety of erosion processes. In this paper we restrict our analysis to bedrock channels.

Previous work suggests that magnitude-frequency relations between flow and bedrock erosion may vary in different climatic, lithologic, and tectonic regimes, demonstrating the need to understand these controls in a variety of field settings. For example, Jansen (2006) calculated for a slowly incising bedrock gorge in Australia that a flood with a thousand-year recurrence interval may be needed to mobilize moderately coarse bed sediment (in this case, the $D_{84}$ size fraction), and suggested that the dominant erosive events are extremely rare in this landscape. Hartshorn et al. (2002) measured bedrock erosion in Taiwan, and found more bedrock lowering from a large typhoon flood compared to subsequent smaller floods. However, when normalized by the estimated recurrence interval (~20 yr) the large flood gave lower annual incision rates (1.7 mm/yr and 0.3 mm/yr in differing lithologies) than smaller floods with approximately yearly recurrence intervals (6 mm/yr and 2 mm/yr, respectively). They interpreted that in this part of Taiwan large floods with moderate recurrence intervals are less important to long-term bedrock river erosion than more frequent but smaller events.

Alluvial cover and sediment supply effects provide a physical mechanism by which bedrock incision may be a nonunique function of flood magnitude. In the saltation-abrasion model, Sklar and Dietrich (1998, 2004) proposed that bedload sediment enables incision by providing "tools" for abrasive impacts. However, at higher sediment loads ($Q_s$) relative to the transport capacity of the flow ($Q_t$), deposition may partially or completely cover the bed and inhibit impact wear and most other erosion mechanisms (the "cover effect"). Laboratory experiments have shown the importance of cover effects in inhibiting short-term incision (Sklar and Dietrich, 2001; Finnegan et al., 2007; Johnson, 2007; Chatanantavet and Parker, 2008). Turowski et al. (2007) interpreted that the rates and patterns of erosion in Taiwan observed by Hartshorn et al. (2002) were influenced by tools and cover effects with bed alluviation commonly preventing thalweg incision during large typhoon floods. Recent field studies document that cover effects can influence long-term rates of bedrock incision (Jansen, 2006; Cowie et al., 2008; Johnson et al., 2009).

Large floods will always have a higher sediment-transport capacity than small floods. However, depending on changes in local shear stress compared to the local sediment load supplied from immediately upstream, a large flood can cause local deposition rather than erosion. The erosive potential of a given flood likely depends on its ability to keep the channel bed exposed while transporting a high sediment load. The timing of water discharge relative to sediment discharge may determine whether a given

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reach aggrades or incises during any given flood event. In some field settings larger floods may be associated with huge inputs of sediment from hillslopes and channel banks, increasing the volume of sediment in the channel and limiting erosion from these events (Howard, 1998). Conversely, where thresholds of sediment mobilization are high, larger floods are expected to be the most erosive (e.g., Snyder et al., 2003; Tucker, 2004; Lague et al., 2005; Jansen 2006). We hypothesize that in field settings with differing relationships between coarse sediment supply and flood hydrology, the relative roles of semi-annual and extreme flood events can be markedly different. Thus an important question is whether the observations of Hartshorn et al. (2002) and Turowski et al. (2007) are generally true, or specific only to those channels in Taiwan, or even to that particular sequence of flood events.

In addition to magnitude-frequency relations, feedbacks between channel width, alluvial cover, bed roughness, longitudinal slope, and incision rate are poorly understood. Understanding how these morphological parameters mutually evolve in response to upstream discharge and sediment flux and downstream base-level change is important for predicting the response of channels and landscapes to tectonic, climatic, or land-use forcing. Channel morphology is a first-order control on the magnitude and distribution of shear stress along the channel boundary, which in turn controls sediment-transport capacity. The distribution of boundary shear stress is commonly used to model bedrock erosion in channels, both directly and indirectly through sediment entrainment, transport, and deposition (e.g., Howard and Kerby, 1983; Hancock et al., 1998; Howard, 1998; Sklar and Dietrich, 1998, 2004; Wohl and Ikeda, 1998; Wohl et al., 1999; Stark and Stark, 2001; Finnegan et al., 2005; Stark, 2006; Wobus et al., 2006; Johnson and Whipple, 2007; Finnegan et al., 2007; Whittaker et al., 2007).

In the present work, we directly measure the relationship between flood hydrographs and bedrock incision rate by monitoring flood events and erosion, albeit in a single channel reach over a short period of time (three years). Rapid bedrock incision has occurred at our monitoring site in response to human modification of the local channel reach. The hydrology of floods in our ephemeral channel is very different from the mountains of Taiwan, giving an opportunity to expand the parameter space in which flow and bedrock incision have been monitored (e.g., Hancock et al., 1998; Hartshorn et al., 2002; Sklar et al., 2005; Stock et al., 2005; Crosby, 2006). In the results section we contrast hydrographs from snowmelt runoff and flash floods, infer differences in sediment transport between runoff events, show how rates and patterns of bedrock erosion varied between events and as a function of local reach slope, and document the local incision of a pothole. In the discussion section we show how changes in channel width and slope resulted from feedbacks between localized bedrock erosion and sediment transport, interpret that cover effects were a first-order inhibitor of erosion, and hypothesize that differences in hydrograph shape were critical to setting the local sediment concentration and size distribution in active transport, which in turn controlled whether local alluvial deposition or bedrock erosion took place. Finally, we present a case study of the morphology of several natural slot canyons in the Escalante River drainage basin, Utah. We argue that the similarity in channel morphology between the monitored channel and these natural slot canyons suggests that the feedbacks we interpret between sediment transport, bedrock erosion, and channel morphology are commonplace.

## FIELD AREA

Swett Creek drains the southeast side of Mount Hillers (peak elevation 3273 m) in the Henry Mountains, Utah, USA. While constructing Highway 276 over Swett Creek (ca. 1970), the Utah Department of Transportation initiated an experiment in bedrock incision by filling a canyon and rerouting the natural channel through a culvert and blasted bedrock slot (Fig. 1). We have monitored bedrock erosion through this human-perturbed bedrock reach, providing an opportunity to observe the morphological evolution of a bedrock channel from a well-constrained initial geometry. Directly downstream of the culvert (diameter 4.3 m, length 70 m, slope 0.024), the vertical-walled, blasted bedrock channel diversion has an upper reach (the "flume," length ~80 m, slope ~0.022, width ~5 m) and a shorter downstream reach that steeply slopes into the original channel (the "flume mouth," length ~17 m, slope ~0.18) (Figs. 1 and 2). The channel elevation at this location is 1437 m a.s.l., and the drainage area is ~24 km$^2$. We measured bedrock erosion along the lower-slope flume and the steeper flume mouth, where a narrow inner channel has formed.

The local bedrock is Navajo sandstone, a Triassic–Jurassic aeolian sandstone that is relatively weak on the core scale (tensile strength ~0.2 MPa, Johnson et al., 2009) but is strong enough to form large cliffs in the region because of its massive, unjointed nature. Blasting of the flume may have locally increased the fracture density. However, smoothly sculpted erosional forms indicate that impact wear is the dominant incision mechanism, both at our site and generally where Navajo sandstone is exposed in the region. Abundant clasts are overwhelmingly composed of durable diorite (tensile strength ~13 MPa) originating from the igneous intrusions that form the peaks of the Henry Mountains (Fig. 1).

Officials contacted at the Utah Department of Transportation found no records of when the culvert and channel diversion were constructed, and so we do not have a precise age constraint on when fluvial incision started. The filling of the Lake Powell reservoir began in 1963 and continued until 1980, and Highway 276 ends at Bullfrog Marina on Lake Powell. Maps published in 1965 and 1970 do not show this highway (Utah Geological and Mineralogical Survey, 1965; Gerlach, 1970), but a 1972 map does (Rocky Mountain Association of Geologists, 1972). We assume that Highway 276 and the culvert were constructed around 1970 and that incision here has taken place since that time.

In the discussion section we introduce a second field site in southern Utah, where narrow slot canyons have formed along several tributaries of the Escalante River. These slot canyons share morphologic similarities to the inner channel that developed at the monitoring site.

## METHODS

At the Swett Creek field site, flow depth was monitored using a sonic distance sensor (SR50, Campbell Scientific) mounted on the top of the culvert interior (Figs. 1 and 2). The sensor measured the distance down to the water surface. Flow depths were calculated by differencing the known distance to the culvert bottom and the measured distance to the water surface. Our configuration could not measure very low flow depths: the sensor has a relatively broad 22° beam acceptance angle (manufacturer's specifications), and the curved bottom of the culvert led to measured distances somewhat less than the distance to the culvert bottom and other occasional spurious measurements when little or no water was present. During the initial monitoring period (26 October 2004–7 June 2005) the sensor was mounted ~4 m from the culvert bottom and the minimum measurable flow depth was ~0.07 m. On 7 June 2005, we lowered the distance sensor to ~2.5 m above the culvert bottom, which improved data quality and reduced the minimum measurable flow depth to 0.03–0.05 m. Flow depths were recorded once every 10 min and represent the median value of measurements taken every minute, allowing approximately seven months of data collection between downloads.

Approximate channel discharge was calculated from the measured flow depths using

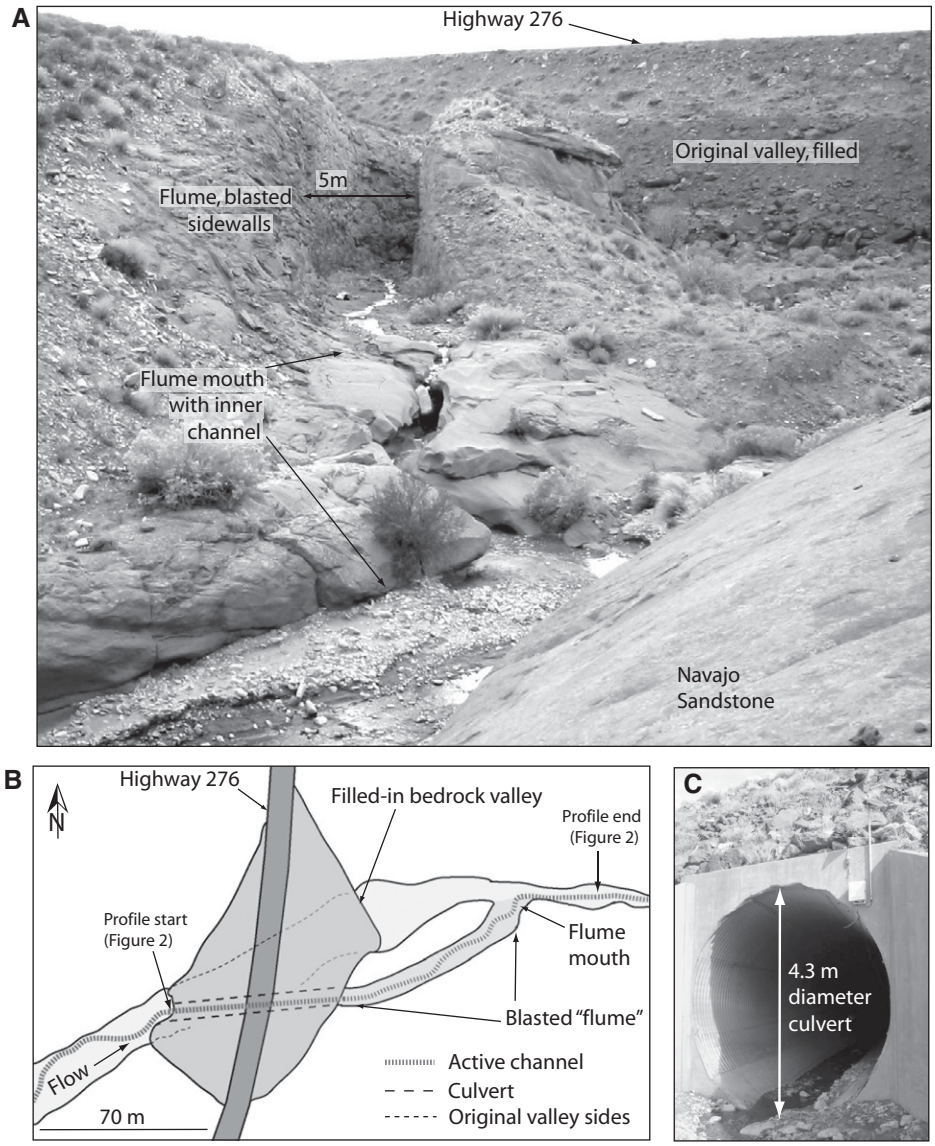


**Figure 1. (A) Photograph looking upstream (east) at the study site. We refer to the blasted slot as the "flume" (because of its similarity to laboratory flumes with rectangular cross sections) and the steep downstream bedrock surface with an incised inner channel as the "flume mouth." Clasts evident in the downstream natural channel (foreground) are primarily composed of diorite. (B) Sketch map view from above, showing the old valley filled in with rubble in order to construct the highway and the culvert and blasted slot through which the channel was diverted. Location is latitude 37.842N, longitude –110.578W. (C) Upstream end of the culvert. The bottom has a curved but smooth concrete lining. Note the solar panel and data logger mounted on the culvert face.**

Manning's equation. Based on surface flow velocities measured with floats, we calculate a range of Manning's $n$ values between 0.01 and 0.016 by assuming the cross-section averaged velocity is between 0.8 and 1 of the measured surface velocity. This Manning's $n$ range is consistent with values suggested for smooth concrete ($n$ = 0.012) to "normal" troweled concrete (0.013) (Ritter et al., 2002). We use $n$ = 0.012 in all discharge calculations. Hydraulic radius is calculated as a function of depth based on the culvert cross-sectional geometry. We assume that the water surface slope matched the culvert bottom slope.

Bedrock erosion was measured by comparing repeat surveys of local bedrock bed topography. Numerous four-inch concrete expansion bolts were tightened and epoxied into holes drilled into the Navajo sandstone bedrock to act as benchmarks (Stock et al., 2005; Crosby, 2006). We conducted repeat surveys over eight lines of bolts; results are later presented from a representative subset of three of these lines. Measurements of erosion were made between bolts to minimize the effects of local perturbations of flow and sediment transport around the bolts themselves. To do this, we initially used a profile gauge (also called a contour gauge), which consisted of parallel metal rods 1 mm in diameter, held next to each other in a housing that allows them to slide parallel to one another. The profile gauge can measure a linear cross section of topography 290 mm wide and 110 mm deep. Photographs of the profile gauge were taken on a gridded surface, and image processing was used to extract the topography at 1 mm spacing. This technique required bolt heads to be present in each measured profile in order to align profiles from different times. However, this proved to be a problem after the first season because, as will be shown, many bolts were removed by erosion. In places where bolts were lost, later surveys were conducted with a total station. When necessary, sediment was excavated from the bottom of the inner channel to expose the bed between bolts for resurveying. Sediment was then replaced, although clasts were more loosely packed following excavation.

We were unable to monitor sediment flux, despite its fundamental role in impact wear—the dominant erosion mechanism at this location. We have one direct measurement of coarse bedload transport during moderate snowmelt flow, demonstrating that high sediment fluxes do occur in this channel. We also interpret first-order differences in sediment transport between snowmelt runoff and flash floods from changes in deposition in the inner channel.

## RESULTS

### Flow Monitoring

Our monitored flow record shows that the ephemeral channel carries flow from snowmelt on Mount Hillers in some but not all years and also experiences flash floods from summer monsoon thunderstorms that occur most frequently from July through October. Figure 3 shows hydrographs for flow events that occurred between 26 October 2004 and 15 May 2007. Possible events shallower than 0.03–0.07 m depth and shorter in duration than ~40 min (four data points) were missed. The 10 min recording interval is too slow to capture the exact initial peak height of flash floods but nonetheless is sufficient to capture many hydrograph details.

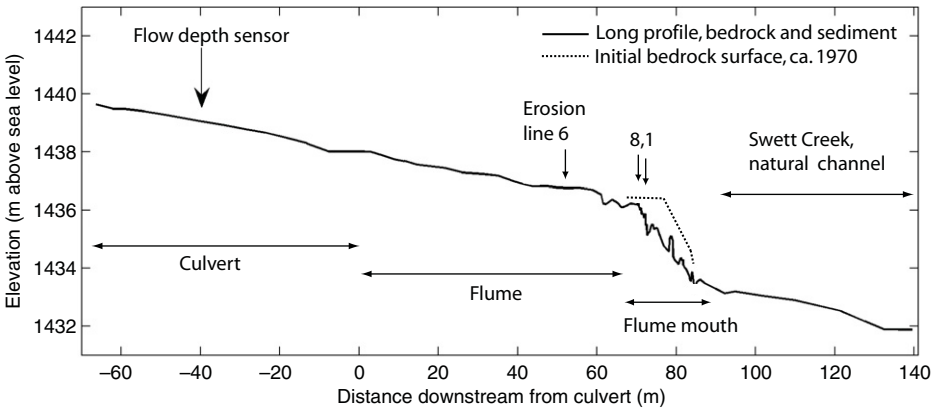


**Figure 2. Longitudinal channel profile for the section of Swett Creek diverted through the culvert, blasted flume, and over the steep and eroding flume mouth before returning to the original channel. The location in the culvert of the sonic flow-depth sensor is indicated, as well as the locations of surveyed topographic cross sections where erosion was measured.**

We are unaware of other comparable records of stage and discharge in the Henry Mountains vicinity.

Flow from snowmelt occurred for three weeks in April and May 2005, with a maximum calculated snowmelt discharge of ~1 $m^3/s$ (Fig. 3). The snowmelt hydrograph changes slowly and shows daily fluctuations. Winter 2004–2005 was an above-average snow year across much of the Colorado River basin, including here. The springtime increase in the Lake Powell reservoir water volume in 2005 was the largest since 1997 and the sixth largest springtime volume increase since 1969 (U.S. Bureau of Reclamation, 2007). In contrast, zero snowmelt flow was recorded in 2006 or 2007.

The other measured flow events were summer and fall flash floods. They show rapid increases followed by approximately exponential reductions in flow depth, with durations measured in hours, not days (Fig. 3). By far the largest short-term discharge event we captured occurred on 5 October 2006, with a measured flow depth in the culvert of ~0.7 m and a calculated maximum discharge >9 $m^3/s$. Until this event, the bottom of the culvert was bare of sediment except at the very upstream and downstream ends. Following the October 2006 flash floods, the entire culvert had a continuous layer of coarse sediment on its bottom, measured below the depth sensor to be 0.13 m thick.

Flow duration curves show the percentage of time that flow (measured at 10 min intervals) equaled or exceeded a given flow depth (or discharge) (Fig. 4) (e.g., Vogel and Fennessy, 1995). The curves are generated by first sorting the individual 10 min flow measurements from highest to lowest, and then calculating the percentage of total time (in this case, three years) represented by the cumulative number of 10 min measurements exceeding a certain flow depth (or discharge). Over the three-year monitoring period, snowmelt runoff exceeded 0.1 m depth ~2% of the time or ~22 days, while flash-flood flow exceeded 0.1 m depth ~0.07% of the time, or ~18 h. We present these results not only to contrast flow from flash floods and from snowmelt but also because there is relatively little data on ephemeral stream flow, making these data useful even though the time period of the flow record is limited (Reid et al., 1998). In this record, flash flood flow becomes dominant at discharges higher than ~1 $m^3/s$. By summing calculated channel discharges, we estimate that the recorded snowmelt flow transported approximately eight times more water volume than the flash-flood flow. The flash-flood events from October 2006 account for ~90% of the total flash-flood flow volume.

We observed in the field that at a discharge of ~0.5 $m^3/s$, all of the flow was contained in the flume-mouth inner channel (variable width, ~0.4–2 m), while immediately downstream along the lower-slope natural channel, the same discharge filled the entire width of the bedrock-walled canyon (width 4.2 ± 1.6 m, 1$\sigma$ surveyed variability). At peak snowmelt discharges, approximate stage indicators such as plant debris indicated that flow overtopped the flume-mouth inner channel.

## Sediment Transport

On 24 May 2005, we collected several bedload samples during snowmelt runoff. An improvised bedload trap was constructed using a fabric mesh bag (with holes of several mm diameter) and a rectangular opening 270 mm wide by 70 mm tall. Measurements were made in the lower-slope flume reach several meters downstream from bolt line 6 (Fig. 2), where flow most closely approximated steady and uniform conditions. At a calculated discharge of ~0.5 $m^3/s$, where the active channel was ~3 m wide (i.e., the trap width covered ~1/12 of the channel width), we collected a sample of 4.7 kg in one minute, with a maximum clast diameter of ~80 mm. The median diameter in transport was $D_{50}$ = 11 mm ($D_{16}$ = 5 mm, $D_{84}$ = 21 mm). These sediment sizes in active transport are somewhat smaller than the local bed surface alluvium ($D_{50}$ = 23 mm, $D_{16}$ = 8, $D_{84}$ = 45, measured by random-walk point counts), although apparent bed armoring by a coarser surface layer in this channel was not observed during periods without flow. Transport was vigorous but highly variable at the time scale of minutes, with no apparent changes in local flow (e.g., Bunte and Abt, 2005). While sampling was not sufficiently detailed in space or time to justify calculating a channel-wide bedload transport rate, this measurement demonstrates that abundant coarse-sediment transport occurs in this channel at moderate discharges.

We do not have any direct measurements of sediment transport during flash-flood events. However, repeat surveys of partial alluviation along the flume-mouth inner-channel bottom reveal that the flash floods tended to deposit sediment in the inner channel, while the recorded snowmelt event effectively cleared sediment from the inner channel (Fig. 5). Initially in spring 2004 the inner channel was quite clear of sediment. We did not survey the thalweg profile or sediment cover in fall 2004, which is unfortunate since sediment cover increased between spring 2004 and fall 2004, and then decreased as a result of the spring 2005 snowmelt. Based on photographs and field notes we estimate the inner-channel sediment cover present in fall 2004, which included an extensive sediment blockage in the upstream third of the inner channel (Fig. 5). There had clearly been flow during this time interval (before we installed the flow-depth sensor) which deposited a large amount of alluvium in the inner channel, including a boulder (long axis ~1 m) that became wedged between the inner-channel walls above the bed. Most of this sediment was then excavated by the spring 2005 snowmelt flow, although the large boulder moved downstream only slightly (to between meters 78 and 79; Fig. 5). Subsequently, sediment cover in the inner channel increased due to flash floods, and in particular the fall 2006 floods.

There is no clear evidence that road and culvert construction substantially perturbed the sediment-transport field, although we cannot

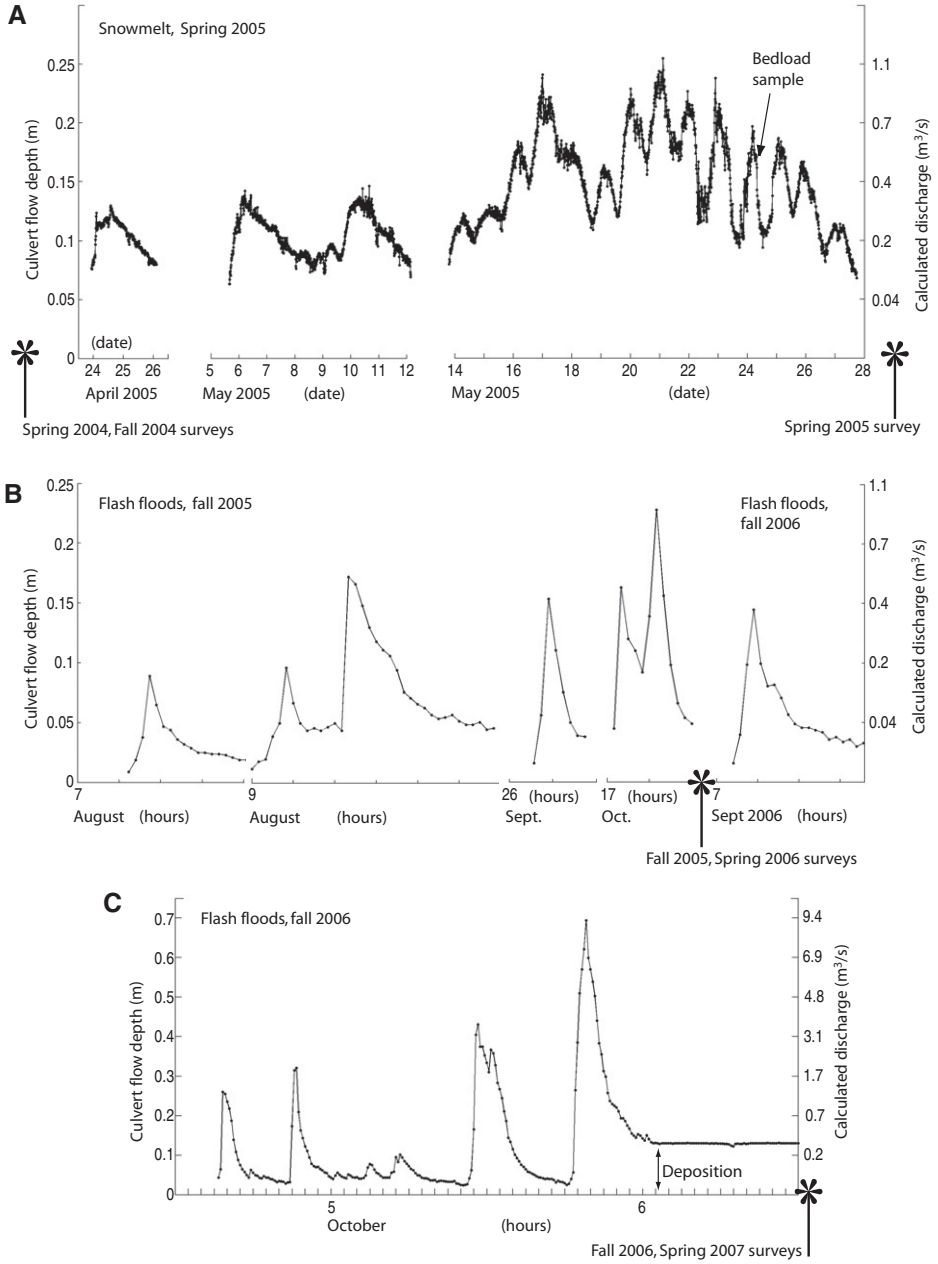


**Figure 3. Flow-depth and discharge hydrographs showing all of the flow events observed in this study. Data have not been smoothed, although some spikes were removed from the snowmelt record (A). Tick marks on the abscissa represent days (of the month) for the plots of spring 2005 snowmelt (A). Tick marks on the flash-flood plots represent hours (B and C). Note the different vertical scale of the fall 2006 flash floods (C). Each individual data point represents 10 min in all plots. The timing of our topographic surveys relative to the flow record is indicated.**

discount this possibility. Upstream of the culvert the active channel and the bedrock-walled valley contain abundant coarse sediment. Bedrock outcrops occasionally along the channel bed and banks, but most of the channel is alluvial with no apparent surface layer coarsening. We interpret that coarse-sediment supply is not limited along Swett Creek for kilometers upstream of the culvert. The upstream Swett channel has a similar morphology to Trail Canyon, located ~5 km north, which we also interpret to not be supply-limited in its sediment load (Johnson et al., 2009). For 50 m upstream of the culvert, the active channel narrows and is modestly incised (up to 1 m) into an alluvial surface that may represent the local channel floodplain prior to culvert construction, suggesting that the upstream culvert end was built slightly lower than the active channel and that local sediment transport may have initially increased. However, in a channel of this size the perturbation to the sediment-transport field would likely have been slight and the response fast.

From our three-year record of flow and deposition, we interpret that high sediment-transport rates occurred due to both snowmelt and flash-flood flow. However, flash floods increased deposition along the inner channel, while snowmelt flow exposed bedrock along the inner channel. We hypothesize possible explanations for the relations between hydrographs and sediment transport in the discussion section.

## Patterns of Erosion

### *Threshold of Detachment*

Qualitative field observations suggest that a negligible threshold of detachment is required to abrade bedrock (Fig. 6). The photograph shows the contact between an alluviated channel bed and bedrock sidewall, taken in Swett Creek ~1 km upstream from the culvert. The bedrock here is Entrada sandstone (tensile strength 0.7 MPa), a Jurassic aeolian unit broadly similar to Navajo sandstone in strength and its massive, unjointed nature. The close correspondence in shape between focused sidewall erosion and clasts as small as ~50 mm (intermediate diameter) suggests that the eroding flows were not large enough to move these bed clasts. Smaller granules and sand in transport presumably provided the abrasive tools. Beyond the threshold of sediment motion, we interpret that there is no minimum flow intensity needed to initiate impact wear in this relatively weak sandstone. The observation is consistent with disk mill abrasion experiments that found no evidence for abrasional detachment thresholds over a wide range of rock and sediment strengths (Sklar and Dietrich, 2001).

### *Erosion Line 6*

Bolt line 6 forms a channel cross section in the lower-slope flume ~20 m upstream of the flume mouth (Fig. 7A; location in Fig. 2). Note two incised longitudinal grooves, filled with moderate sediment cover. Repeat surveys of the cross-sectional channel topography show that up to ~100 mm of vertical incision into bedrock occurred as a result of snowmelt flow between fall 2004 and spring 2005, focused at the topographic lows of the groove bottoms (Fig. 7B). Little to no bedrock erosion occurred here from flash floods for the duration of our study.

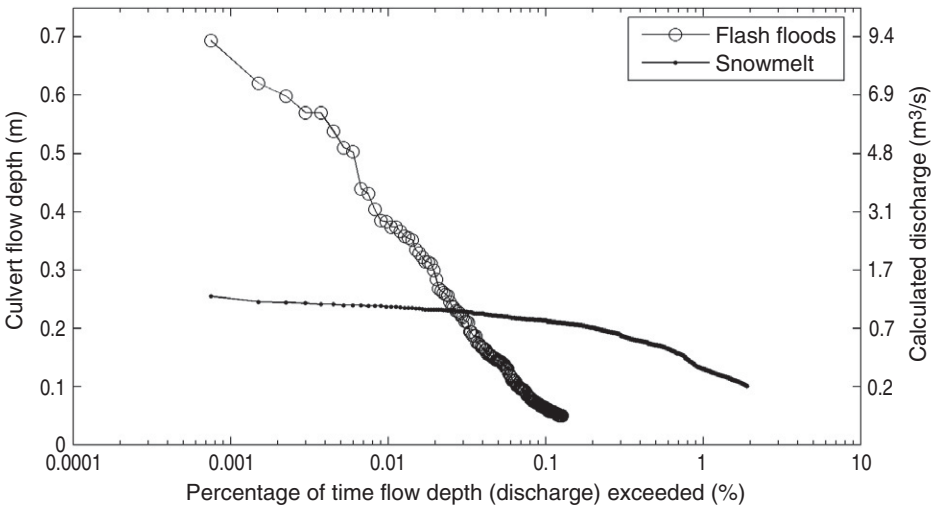


**Figure 4. Flow duration curves calculated for flash-flood flow and snowmelt.**

### *Flume-Mouth Longitudinal Profile*

Along the flume mouth, up to ~500 mm of vertical incision into bedrock occurred during spring 2005 snowmelt flow (Fig. 8). Photographs and comparisons to other erosion profiles demonstrate that inner-channel downcutting did not occur between the spring 2004 and fall 2004 surveys. Later surveys (fall 2005 to spring 2007, omitted for clarity) show no apparent bedrock erosion at this scale but do clearly show increasing alluvial deposition (Fig. 5). The transient incisional response has been for the initially steep bedrock surface to erode to a lower longitudinal slope, although a clear slope break has persisted between the flume and flume mouth (Figs. 2 and 8).

### *Erosion Line 1*

A key element of the incisional response since the flume mouth was constructed and incision began (~1970) has been a dramatic narrowing of the zone of flow and sediment transport, forming the inner channel. In fall 2004, bolt line 1 was installed just upstream of the start of the inner channel, but ~450 mm of vertical erosion during the spring 2005 snowmelt season (Fig. 9A) led to the upstream migration of the start of the inner channel (Fig. 8). The width of focused incision was less than ~350 mm. Photographs taken in fall 2004 (before snowmelt runoff) and fall 2006 (after both snowmelt and flash floods) show bedrock incision, alluvial deposition, and bolt deformation by sediment impacts (Figs. 9B and 9C). Up to 20 mm of incision occurred on the upper bedrock surface near bolt f as a result of the fall 2006 flash floods.

Different surveyed profiles align poorly in places because of the change in surveying

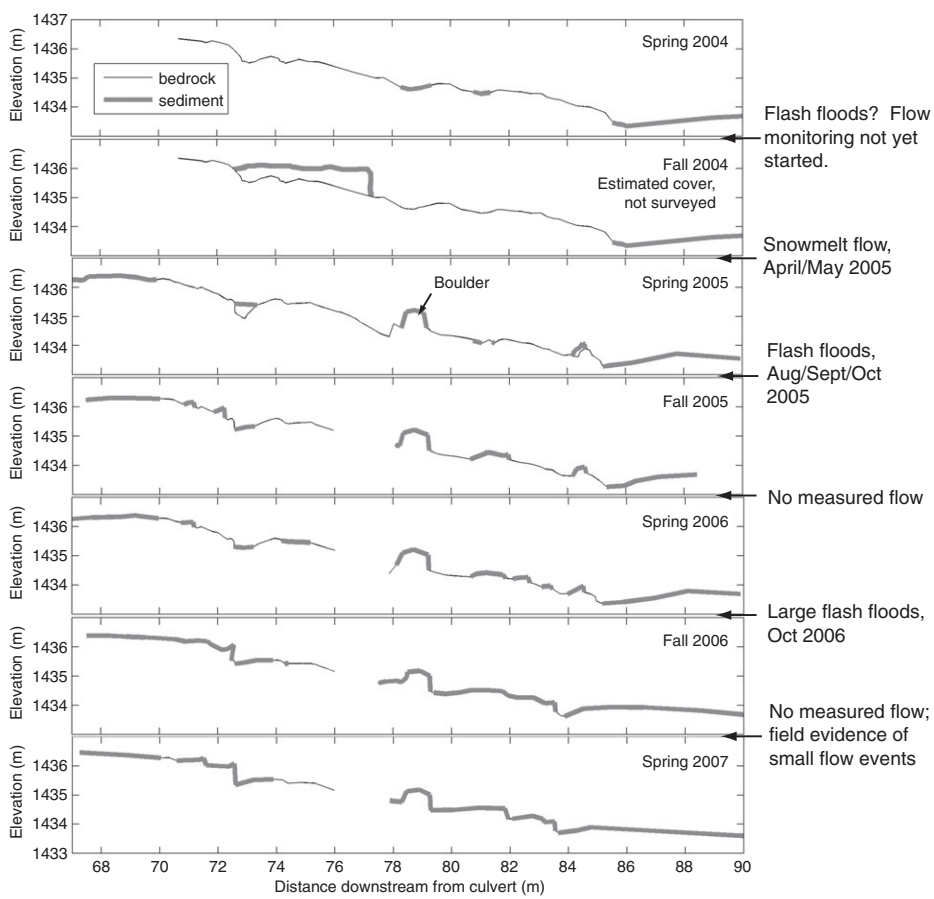


**Figure 5. Sediment deposition along the inner channel. The fall 2004 alluvial cover is estimated based on photographs and field notes, using the spring 2004 surveyed bedrock profile. Sediment deposition in the rest of the time steps was directly surveyed. Figure 3 shows when the surveys were done relative to the flow record. No flow was measured between fall 2005 and spring 2006; the differences in sediment cover between these time steps are relatively small, and either represent uncertainty in survey repeatability or that flow lower than what we can reliably measure occurred and modestly rearranged sediment in the channel. Between fall 2006 and spring 2007 low flow did occur (based on qualitative field observations of flow indicators) but was not recorded. The gap at 77 m is an undercut section of channel; bedrock bed elevations were measured in this reach in spring 2005 but not after. See text for transport history of boulder at ~78 m.**

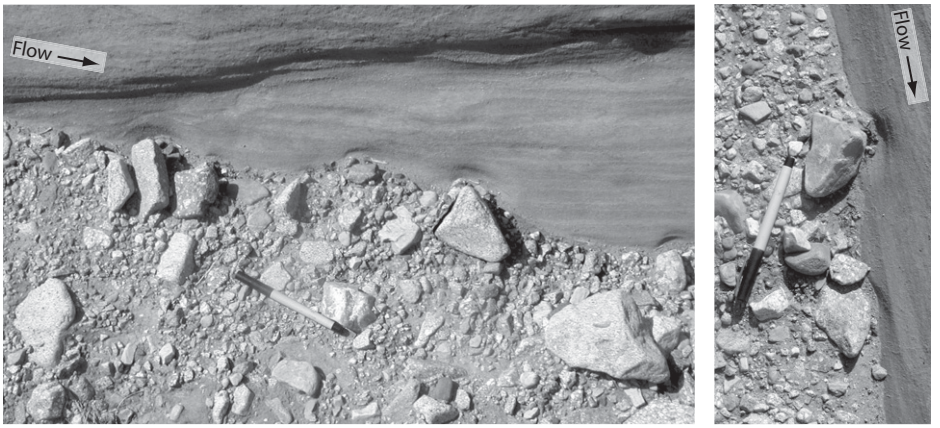


**Figure 6. Photographs of sidewall bedrock erosion around relatively small bed clasts in Swett Creek upstream of the culvert taken in spring 2004. We interpret that the sidewall erosion was accomplished by sand and granules in transport during flows too low to initiate motion of the adjacent bed clasts, and that the threshold to abrade this bedrock is negligible. The marker is 140 mm long.**

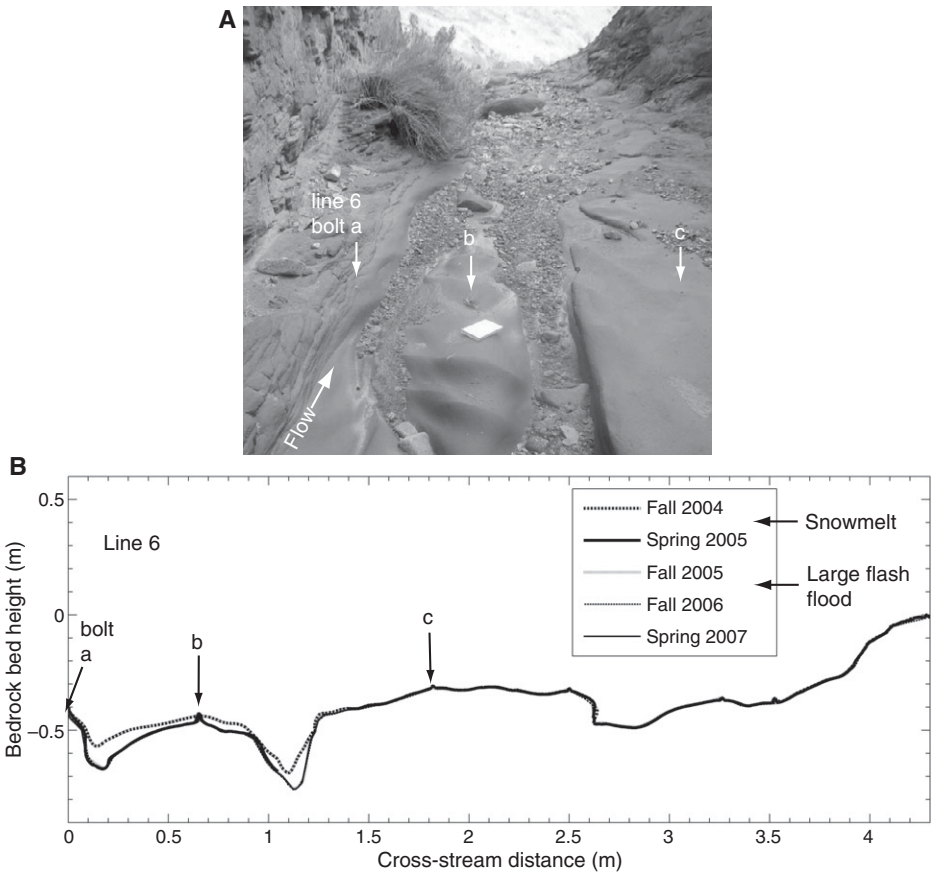


**Figure 7. (A) Photograph looking downstream at erosion line 6, located ~20 m upstream of the flume mouth (Fig. 2). Bolts a, b, and c are labeled. Photo taken in fall 2005; note the sediment filling both longitudinal grooves. (B) Line 6 repeat surveys, showing up to 100 mm of bedrock incision from spring 2005 snowmelt flow and negligible incision due to later flash floods. Surveyed lines are plotted in sequential order, and therefore later time steps cover the earlier ones. No bolts were lost due to bedrock erosion along this line; the initial bolt spacing was wider than the ~250 mm given in the methods section due to a limited supply of bolts at the time of installation.**

method from using a profile gauge to a total station after the alignment bolts in the center of the channel were removed by erosion (Fig. 9). It was difficult to accurately resurvey the same lines on steeply sloping surfaces, particularly when some of the benchmark bolt heads were bent or removed. Recording an accurate elevation at the base of the inner channel proved much more reliable (since the surveying rod tip rests solidly on the thalweg bottom rather than being held laterally against a steeply sloping surface), and the bottom elevation demonstrates negligible erosion following the spring 2005 snowmelt. Improving the resurveying technique would not change the first-order observation that little to no flash-flood erosion occurred compared to the snowmelt erosion.

### *Erosion Line 8*

Following the spring 2005 snowmelt and erosion event that removed many bolts along the flume-mouth inner channel, a new line of bolts was installed very close to the channel slope break that marked the new upstream start of the inner channel (Fig. 10; see Fig. 8 for cross-section location). The repeat surveys of this line were done using the higher-resolution profile gauge. Up to 10 mm of bedrock erosion occurred during subsequent flash floods. Erosion between spring 2005 and fall 2005 was broadly focused in topographic lows. Conversely, erosion between spring and fall 2006 primarily occurred on topographic highs. Figure 10C shows the bolt line in fall 2006, looking across the inner channel. Sediment filled the topographic low. The patterns of deposition and erosion suggest that local sediment cover was responsible for inhibiting incision.

Finally, measurements of erosion along bolt lines 2–5 and 7 are not presented but are consistent with the above results. Lines 2 and 7 were located on the flume-mouth surface outside of the inner channel, and show that minimal erosion occurred outside of the inner channel. Lines 3 and 4 were downstream inner-channel cross sections located at ~82.2 and 84 m downstream from the culvert (Fig. 8), and show similar results to line 1. Line 5 was located upstream from line 6 in the lower-slope flume, and consisted of bolts placed on a bedrock protuberance in the channel around which a horseshoe-shaped form had previously eroded. Figure 3d of Johnson and Whipple (2007) shows a photograph of bolt line 5.

### *Pothole Formation*

We fortuitously measured the erosion of a pothole along the inner channel. Erosional bedrock potholes are widespread, and their incision may be a dominant cause of channel

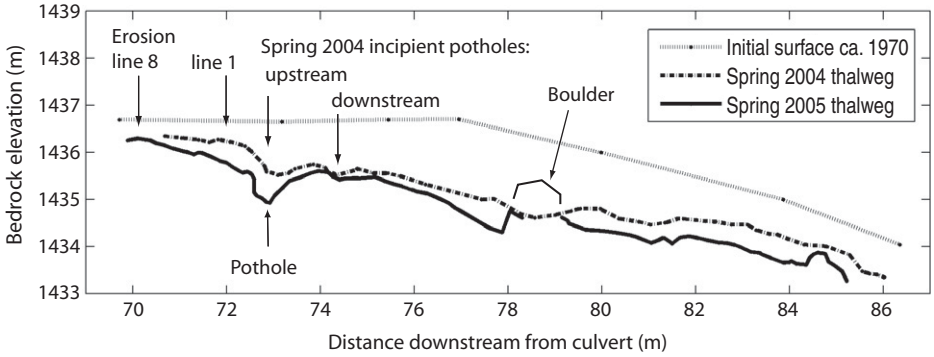


**Figure 8. Flume-mouth, inner-channel bedrock longitudinal profile, showing extensive bedrock erosion along its length due to spring 2005 snowmelt flow. An approximate initial bedrock surface prior to incision (ca. 1970) was reconstructed from surveys of the bedrock on either side of the inner channel. See text for transport history of boulder at ~78 m.**

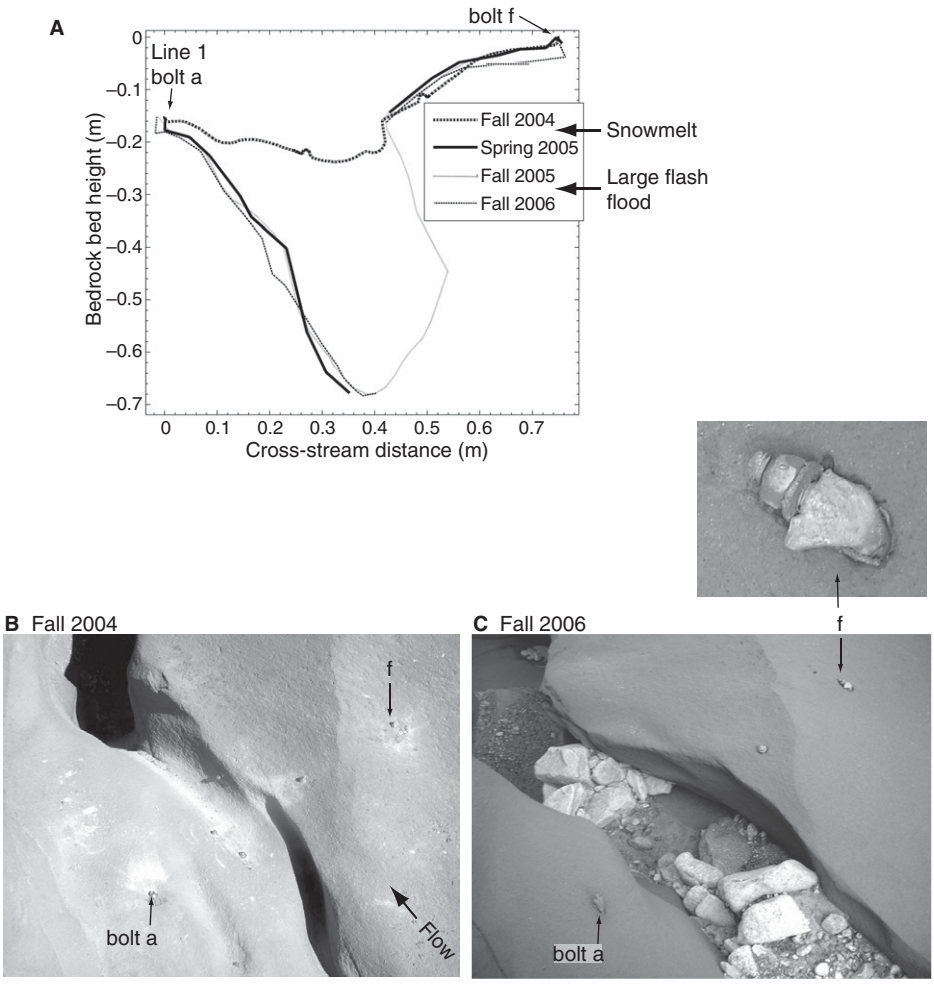


**Figure 9. Erosion line 1 (location in Fig. 8). (A) Repeat surveys of the topography between bolts. The fall 2005 survey was conducted with the total station in "reflectorless" mode from multiple viewpoints to accurately capture the undercut. No vertical exaggeration. (B and C) Photographs looking down from above on bolts in fall 2004 (immediately following installation) and in fall 2006. Distance between bolts a and f is ~750 mm. Bolt f was bent by sediment impacts during the fall 2006 flash floods. Flow toward upper left.**

lowering in some field setting, and yet surprisingly little is understood about their formation (e.g., Alexander, 1932; Whipple et al., 2000; Barnes et al., 2004; Richardson and Carling, 2005; Springer et al., 2005; Johnson and Whipple, 2007). In the initial profile there were two incipient potholes (depressions) of similar size (Fig. 11A). Both the upstream and downstream incipient potholes had complex erosional forms, but only the upstream depression grew in size due to subsequent bedrock erosion. After the spring 2005 snowmelt flow, the downstream incipient pothole was no longer identifiable as a separate form but was smoothly subsumed into the inner channel by erosion (Fig. 11B), while the enlarged upstream pothole was filled with coarse gravel and cobbles to a depth of 0.30–0.4 m.

Figure 8 shows inner-channel longitudinal profiles through the pothole. Figure 11C shows a pothole cross section approximately perpendicular to the inner channel. This cross section was surveyed in fall 2005, although longitudinal profile surveys demonstrate that lowering did not occur between spring 2005 and fall 2005. In fall 2004, this zone of the inner channel was completely alluviated (Fig. 5); we assume that all of the pothole incision occurred during the snowmelt season of spring 2005 and not between spring and fall 2004. The last pothole excavation and resurvey took place in fall 2005.

## DISCUSSION

In the following sections we explore bedrock channel morphology in the context of our flood hydrographs and corresponding erosion rates and patterns. First, the evolution of channel width and slope gives insights into controlling feedbacks. Second, our data show that partial alluvial cover can inhibit incision at short time scales. Third, the relationship between flood magnitude and bedrock erosion leads to hypotheses on how hydrograph shape influences sediment transport. Fourth, we interpret mechanisms of pothole incision and how pothole formation may influence reach-scale erosion. Finally, we discuss implications of our data for slot-canyon formation, by interpreting the incisional history of several natural canyons in Southern Utah.

In Swett Creek, the largest peak discharge events were not the most erosive. We do not expect this observation to be universal in other landscapes, or even in every channel in this landscape. Instead, we interpret that the erosivity of a given flood depends directly on local bedrock exposure and coarse sediment flux as suggested by recent models (e.g., Sklar and Dietrich, 2004). In our record, flash floods

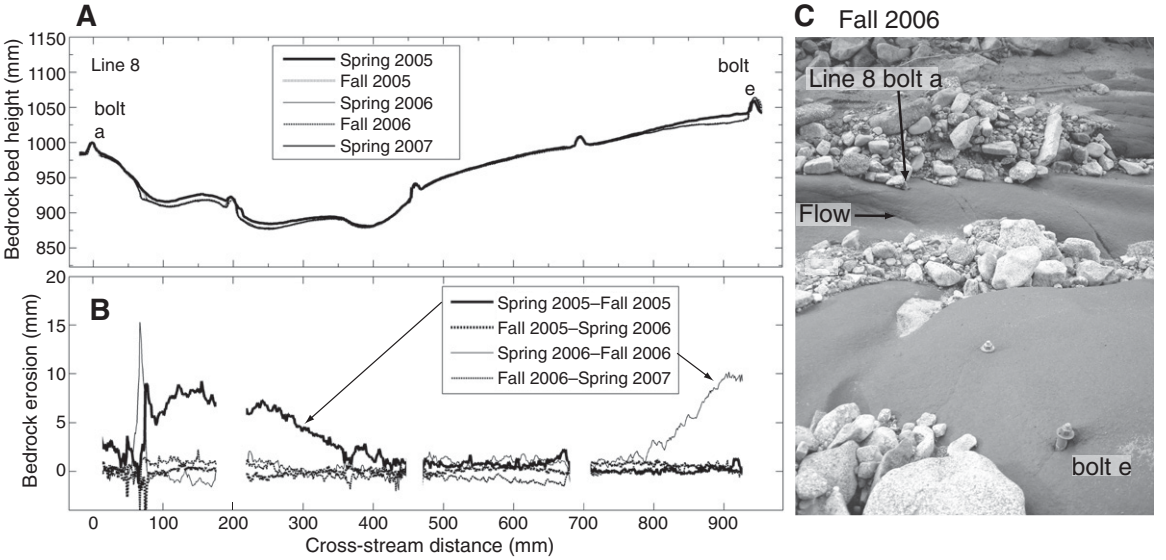


**Figure 10. (A) Cross-section line 8, just upstream of inner channel (location in Fig. 8). (B) Bedrock erosion (i.e., the difference between sequential cross sections). The measurement uncertainty is approximately ±2 mm. (C) Line 8 photograph looking across the inner channel from a low angle from right to left along the cross section. Note the bedrock erosion around the foreground bolt that occurred between spring and fall 2006.**

transported abundant coarse sediment but were net depositors in the inner channel and caused little to no erosion. In contrast, sustained but moderate-magnitude snowmelt discharge excavated coarse sediment out of the steep inner channel, increased bed exposure, and caused extensive bedrock erosion. As shown by erosion along the inner channel, local bedrock erosion can occur rapidly: we measured nearly 500 mm of vertical incision into bedrock along much of the inner channel over 23 days of snowmelt flow. This high incision rate is a result of the steep local channel slope, weak bedrock, a disequilibrated channel morphology, and high but not overwhelming sediment flux. Nonetheless, the rapid short-term erosion rate suggests an interesting and converse question: why are long-term fluvial bedrock incision rates so much lower? For example, Cook et al. (2009) reported long-term incision rates of ~0.4 mm/yr based on cosmogenic dating of alluvial terraces along a well-adjusted channel in the Henry Mountains, consistent with regional measurements of long-term incision (Garvin et al., 2005). Long-term landscape incision in areas of rapid tectonic uplift such as Taiwan can approach rates as high as ~10 mm/yr (e.g., Fuller et al., 2003).

One well-known factor is that bedrock-eroding floods only occur rarely. We have constrained, over a short time period in one channel, the frequencies of recurrence of floods that incise bedrock (Fig. 4). If local erosion was a deterministic function of shear stress, predicting bedrock erosion would mainly require accurate hydrologic measurements of flood magnitude-frequency relations (e.g., Hartshorn et al., 2002; Lague et al., 2005). However, our data also suggest that variability in sediment supply, transport, and deposition leads to a nonunique relation between flood size and bedrock incision, qualitatively consistent with the saltation-abrasion model (Sklar and Dietrich, 2004). Channel morphology also modulates these factors as explored below.

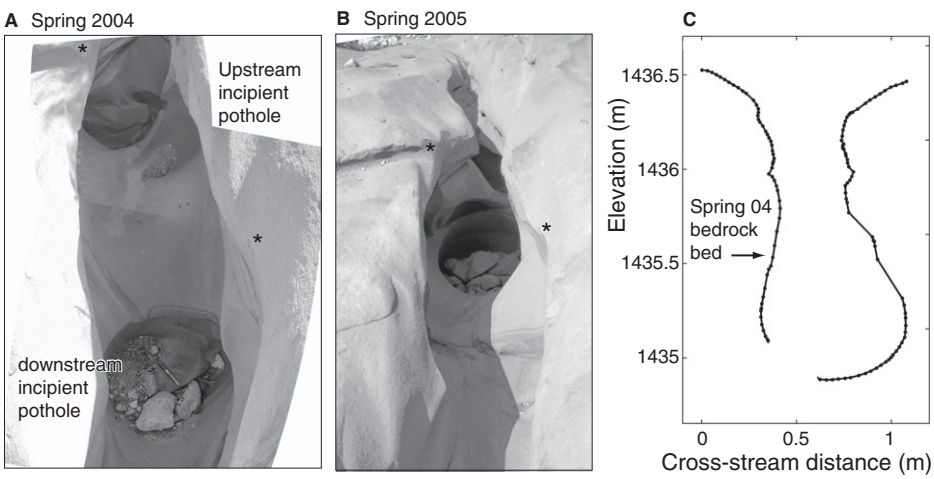


**Figure 11. Potholes. (A and B) Photographic comparison of the development of a pothole (spring 2004–2005). In spring 2004 two incipient potholes were present, but following spring 2005 snowmelt runoff, the downstream incipient pothole had eroded away while the upstream incipient form deepened ~500 mm to form a well-defined single pothole. Note somewhat different photograph scales and orientations. Asterisks mark matching locations in the two photographs. The spring 2004 image is a mosaic of two photographs. (C) Fall 2005 survey of the pothole cross section, measured with a total station in "reflectorless" mode, combining surveys from multiple locations to accurately capture undercuts. As discussed in the text, this detailed survey was completed in fall 2005, but the erosion that created the pothole resulted from spring 2005 snowmelt runoff.**

## Erosion-Driven Morphological Adjustment

The morphology of the Swett flume mouth is the result of transient channel incision from an initial condition of a steep, unchannelized bedrock slope (Figs. 1 and 8). Feedbacks between flow, sediment transport, and erosion lead to (1) width narrowing and the formation of an inner channel, (2) variable alluvial deposition along the inner channel, and (3) longitudinal slope relaxation. Channel morphology (slope, width, depth, roughness, and alluvial cover) is a degree of freedom that mediates landscape response to changes in boundary conditions and forcing. The evolution of channel morphology at the Swett flume mouth is consistent with and may help to explain previous field observations (e.g., Wohl, 1993; Wohl and Ikeda, 1998; Wohl et al., 1999), flume experiments (e.g., Shepherd and Schumm, 1974; Wohl and Ikeda, 1997; Finnegan et al., 2007; Johnson and Whipple, 2007), and numerical modeling (e.g., Johnson and Whipple, 2006; Stark, 2006; Wobus et al., 2006).

Morphological feedbacks observed in recent laboratory experiments (Johnson and Whipple, 2007) provide a useful context for interpreting our observations. These laboratory experiments started with a relatively steep and planar bed with minor roughness, made of erodible "bedrock" (weak concrete). Constant water and sediment fluxes were imposed. When considered over the entire laboratory flume bed area, the total sediment flux was much less than the total calculated sediment-transport capacity ($Q_s << Q_t$); however, below we make an important distinction between "channel-averaged" $Q_s$ and $Q_t$ (total values of each over the entire bed area) and "local" $Q_s^{loc}$ and $Q_t^{loc}$, determined at a single location on the bed. Under conditions of channel-averaged $Q_s << Q_t$, the bed eroded to form a narrow inner channel. Below we outline both positive and negative feedbacks among erosion rate, bed topography, sediment transport, and flow that collectively controlled incision rates and patterns. These feedbacks depended in turn on several straightforward relations between variables: (*a*) lateral bedload transport, in which sediment moves down lateral slopes into topographic lows due to gravity (Parker, 1984), (*b*) local tools effects, in which local impact wear increases with $Q_s^{loc}$, (*c*) local cover effects, in which increasing in $Q_s^{loc}$ relative to $Q_t^{loc}$ inhibits local impact wear, and (*d*) $Q_t^{loc}$ decreases due to increased local bed roughness and increased form drag.

Starting from a planar bed with minor surface roughness, (*a*) and (*b*) together led to an initially strong positive feedback where bedload was preferentially transported along subtle zones of interconnected low topography (*a*), which caused erosion and local bed lowering to be focused along the narrow zone of enhanced transport (*b*), which in turn increased lateral transport and local sediment concentration (*a*), increased the local incision rate over this now narrower zone (*b*), and so on. This positive feedback resulted in the incision of the inner channel. Following this morphological development, a negative erosion rate feedback occurred: the narrow inner-channel width and local bed roughness reduced the local transport capacity along the inner channel, while at the same time the local sediment flux increased because the deep inner channel captured all of the sediment in the experiments. Erosion rates along the inner channel then decreased due to cover effects (*c*), because narrow inner-channel width, increased local bed roughness, and increasing inner-channel depth increased flow drag and reduced $Q_t^{loc}$ along the inner channel (*d*), while at the same time lateral transport continued to increase $Q_s^{loc}$ (*a*). The overall outcome was a narrow inner channel along which cover effects inhibited incision. Furthermore, the inner-channel topography approached a quasi-equilibrium state adjusted such that the local transport capacity was balanced to just transport the sediment flux from upstream. Finnegan et al. (2007) observed similar feedbacks, although their experimental design was somewhat different.

### *Width Adjustment*

At the field site, we interpret that the inner channel at the Swett Creek flume mouth developed as a result of the positive and negative feedbacks described above. The 0.4–2 m width of the Swett Creek flume-mouth inner channel is much narrower than the active channel width just upstream in the lower-slope blasted flume (~5 m) or in the natural channel upstream and downstream of the culvert and study reach. Initially, the steep bedrock surface at the flume mouth was presumably unchannelized (Figs. 1 and 8). The transport capacity of the culvert and flume reach are sufficient to transport the sediment entering the culvert from upstream (as indicated by partial bedrock exposure observed along the flume reach), so sediment supply can be considered longitudinally uniform through the flume and flume mouth. In contrast, transport capacity varies with slope and width and bed roughness. At the flume mouth prior to inner-channel incision, accelerating flow would have greatly increased $Q_t$ relative to just upstream, leading initially to flume mouth $Q_s/Q_t<<1$. Sediment would have then been preferentially transported along local lows, focusing impact wear along a narrow zone and eroding the inner channel. At the same time, roughness of the inner-channel bed and sidewalls would have increased local form drag, reducing the inner-channel transport. While bed cover changed on a storm-by-storm basis, the pattern of increasing cover (Fig. 5) is at least consistent with the interpretation that the channel morphology may gradually approach cover-dominated conditions, as occurred in laboratory experiments (Finnegan et al., 2007; Johnson and Whipple, 2007).

### *Slope Adjustment*

The average inner-channel slope decreased from its initial condition while maintaining a rough but fairly linear profile (Fig. 8). The downstream elevation was set by the sediment fill level of the natural channel. The head of the inner channel migrated upstream ~2 m (the distance between bolt line 1 and line 8) due to local bedrock erosion during the spring 2005 snowmelt event. The incisional response is consistent with the relaxation of an initially steep slope, rather than a steep knickpoint that migrates back retaining its form (e.g., Gardner, 1983; Whipple and Tucker, 2002; Crosby and Whipple, 2006).

A sharp slope break occurs at the transition point between the lower-slope channel upstream and the inner channel downstream, rather than a gradual change in slope as is predicted by purely diffusive models of erosion (e.g., Begin, 1988; Whipple and Tucker, 2002). The inhibition of erosion by sediment cover may maintain this sharp slope break by protecting the bed from erosion immediately upstream of the inner channel (Fig. 5, distance downstream 67–70 m). Seidl et al. (1994) suggested a similar cover mechanism for the preservation of migrating bedrock knickpoints in Hawaii, where boulders covered the bed above and below steep exposed bedrock reaches.

## Cover Effects

Flash floods were much less erosive than snowmelt flow, even though at least two floods on 5 October 2006 reached peak discharges calculated to be nearly an order of magnitude larger than the maximum snowmelt flow (Fig. 3). Based on the pattern of sediment entrainment during snowmelt flow but deposition during flash floods, alluvial deposition was the dominant mechanism that prevented local incision along the inner channel. These results further validate that alluvial cover is an important negative feedback on bedrock erosion over a range of time scales (Sklar and Dietrich, 2004; Turowski et al., 2007; Cowie et al., 2008; Turowski and Rickenmann, 2008; Johnson et al., 2009).

The differences in erosion cannot only be explained by cumulative discharge, even though we calculate that the total volume of discharge from snowmelt was approximately eight times larger than the flash-flood discharge. For the

sake of argument, one may assume that erosion scales linearly with integrated sediment flux (i.e., the tools effect of Sklar and Dietrich [2004], while ignoring the cover effect), and that sediment flux in turn scales linearly with averaged discharge. The latter relation is plausible: Barry et al. (2004) empirically proposed $Q_s \propto Q_w^{1.11}$ for armor-free gravel-bedded rivers, based on compiled field data. If vertical incision therefore scaled with discharge, one would predict ~50 mm of vertical incision along the inner channel associated with the flash floods, i.e., 1/8 of the ~400 mm of incision from snowmelt. This amount of cross-sectional lowering would have been easily measured in our surveys of the inner-channel bottom and the channel upstream. However, we observed negligible vertical incision from flash floods, particularly along the inner-channel bed (Figs. 7 and 9). Thus cover effects likely play a dominant role in inhibiting incision.

We did measure minor erosion from flash floods but only in locations where bedrock was exposed, consistent with cover effects. Bolt line 8 was located just upstream of the inner channel and was initially clear of cover (Fig. 10). The modest flash-flood season ending in fall 2005 caused up to 10 mm of broadly distributed bedrock incision, focused in topographic lows. Following these flash floods, the local bed was still essentially bare of sediment cover. However, following the large fall 2006 flash floods, the topographic low was filled with sediment (Fig. 10), and erosion had only occurred at higher elevations along the bolt line. Note that the fall 2006 floods had both maximum and integrated water discharges nearly an order of magnitude higher than the fall 2005 floods but caused similar amounts of total erosion along bolt line 8, demonstrating again that erosion rate and discharge are only indirectly related.

## Flood Hydrographs, Sediment Transport, and Erosion

Swett Creek has no apparent surface armoring, consistent with other ephemeral desert streams (e.g., Laronne et al., 1994). It has an abundant supply of both coarse and fine sediment available in its bed and banks, suggesting that sediment supply is not limited during floods. In addition, we measured high sediment-transport rates with a broad size distribution during moderate snowmelt flow. Why then did snowmelt and flash floods produce such different extents of alluvial cover and amounts of bedrock lowering? We next highlight limited previous work on bedload transport in ephemeral channels, hypothesize how flood hydrographs may control the timing of bedload sediment transport (i.e., sedigraphs) during floods, and discuss possible controlling factors including differences in the transported coarse size distribution and feedbacks with inner-channel morphology.

Pioneering studies on flash-flood sediment transport demonstrate dramatic differences between transport in ephemeral and perennial channels but nonetheless give an incomplete picture of bedload transport during flash floods. Laronne and Reid (1993) demonstrated that bedload transport by flash floods in two unarmored ephemeral channels (Nahal Yatir and Nahal Eshtemoa, Israel) was much more efficient than bedload transport in a perennial gravel-bedded channel with a well-developed coarse surface layer (Oak Creek, Oregon, USA). Reid et al. (1998) found that bedload transport increased rapidly with shear stress during the rising limbs of flash floods. They found a more consistent relation between shear stress and sediment transport than for perennially flowing rivers. They interpret that a lack of surface coarsening (bed armoring) in ephemeral channels enabled extremely high transport rates compared to channels in more humid environments (e.g., Laronne and Reid, 1993; Laronne et al., 1994; Cohen and Laronne, 2005). However, limitations of their measurements demonstrate the difficulty of monitoring bedload transport in ephemeral channels. For example, the bedload monitoring system used by Reid et al. (1998) consists of buried traps that catch and weigh bedload (e.g., Bergman et al., 2007). Reliable bedload measurements with this system are typically not possible in the first "minute or so" following the passage of a flash-flood bore (i.e., the flow front) (Reid et al., 1998). In addition, their traps rapidly become filled and typically only measure bedload transport during a fraction of the rising stages of flash floods, particularly for larger events. Bedload measurements during the falling limbs of flows are rarely captured in their data. Other studies have found hysteresis between unsteady flow and sediment transport but have not typically recreated hydrographs that change as rapidly as flash floods (e.g., Kuhnle, 1992; Admiraal et al., 2000; Lee et al., 2004).

Consistent with Dunkerly and Brown (1999), Malmon et al. (2007) measured flash-flood suspended sediment concentrations (not bedload) and found that concentrations were consistently highest right at the flood bores. Sediment concentrations then decreased rapidly with time, even as water stage and discharge continued to increase up to a peak discharge, demonstrating a complex and nonmonotonic relationship between reach-averaged shear stress and suspended sediment concentration. While such a pattern does not appear to hold for bedload (Reid et al., 1998), it seems plausible that concentrations of bedload may be nonetheless elevated very early on during flash floods, perhaps due to high turbulence in the initial flood bore.

Although we have no direct measurements constraining sediment transport during our flash-flood events, we hypothesize that bedload flux was high at the flash-flood bores, while the slowly varying snowmelt flow had lower peak sediment concentrations and fluxes. One likely factor is that the intense rainfall events that caused flash floods also increased sediment supply into the channel from bank collapse and hillslope erosion, relative to snowmelt runoff. In addition, we hypothesize that flash floods in Swett Creek were depositional and not erosional because of very high initial sediment concentrations near the flood bore, leading almost immediately to early deposition during the rising limb of the hydrograph as well as during the falling limb.

If coarse sediment concentrations were elevated and/or maximum clast sizes were larger during the rising limbs of flash-flood hydrographs than during snowmelt flow, feedbacks with the narrow inner-channel morphology may have similarly enhanced bed alluviation. We do not know how the overall sediment size distribution transported in flash floods compared to snowmelt flow, but we do know that the flash floods transported large cobbles and boulders based on their presence in the inner channel, often as the keystones of sediment jams. Field observations and the surveyed patterns of cover (Fig. 3) from flash floods show that much cover along the inner channel occurred behind local cobble jams, in which one or more large clasts became wedged in the narrow and tortuous inner-channel bottom. Higher sediment concentrations and larger clasts would increase the chances of particles becoming interlocked. Initially, a local sediment jam would increase local form drag and decrease local transport capacity, causing deposition upstream of the jam up to the level where the sediment flux from upstream could be transmitted through the channel once again. During the large fall 2006 flash flood, these feedbacks led to the alluviation of most of the inner channel.

Why was snowmelt flow able to entrain existing alluvial cover, keep the inner channel largely free of alluvium, and extensively erode the bed? Entrainment and erosion were enabled by the slowly varying snowmelt hydrograph and the locally steep channel morphology. First, the gradually changing snowmelt hydrograph probably resulted in lower peak sediment concentrations and less variable fluxes relative to flash-flood sediment transport. Second, the steep channel slope through the flume mouth caused the local transport capacity ($Q_t$) of the

flow to increase relative to upstream, while the local sediment flux was limited by sediment supply from the lower-slope channel upstream. These factors combined to gradually entrain alluvium from the inner channel, pass sediment in transport through the inner channel without depositing, and erode the inner-channel bed. While flume-mouth $Q_t$ would also have increased during flash floods, we interpret that the inner-channel width and rough sidewalls were too constricting to pass the high flash-flood sediment concentrations and large clast sizes we infer, causing cobble jams to form and more than offsetting the effects of a locally higher $Q_t$.

The flume-mouth morphology of a narrow inner channel flanked by steep and broad bedrock surfaces may also cause differences between low- and high-discharge flow events due to a morphologic decoupling between flow and sediment transport. We observed in the field that at a calculated discharge of ~0.5 $m^3/s$, all flow was contained in the inner channel, but stage indicators such as vegetative debris suggested that higher snowmelt flow depths (peak discharge ~1 $m^3/s$) had overtopped the inner channel. With increasing discharge, inner channel $Q_s/Q_t$ may increase once flow overtops the inner channel: The sediment flux moving through the inner channel will continue to increase (because it is controlled by the supply from upstream), but the transport capacity through the inner channel likely remains relatively unchanged with increasing discharge because an increasing amount of flow goes over the bedrock flanks and does not substantially increase inner-channel flow depths or shear stress. However, it is unlikely that this sediment-flow decoupling is the only cause of differences between snowmelt and flash-flood deposition and erosion. A majority of flash-flood peak discharges were no greater than the peak snowmelt discharge, yet these smaller flash floods still increased deposition along the inner channel.

During flow events such as the large floods of October 2006 in which water did overtop the inner channel, the inferred low sediment flux over the higher bedrock surfaces suggests that the tools effect—in this case, a lack of local sediment—likely caused inefficient local bedrock erosion at high bed elevations. Simultaneously, along the inner channel we interpret that the cover effect caused inefficient bedrock erosion. Channel morphology can thus mediate rates and patterns of bedrock erosion in seemingly complex ways. Note that the existing inner-channel width and roughness evolved to its present state (at least over the time period of monitoring) as a result of erosional forcing from snowmelt flow, and was therefore at least approaching a bedrock morphology adjusted to transport sediment under snowmelt conditions but not flash-flood conditions. The morphology of the bedrock channel boundary was only influenced by snowmelt flow, while the channel boundary defined by both bedrock and alluvium developed in response to hydrologic forcing from both snowmelt and flash floods.

## Pothole Formation

In addition to the pothole that developed along the flume mouth (Fig. 11), all of the slot canyons that we discuss in the next section have extensive potholed reaches. A long-running debate in understanding pothole development has been whether the forms are dominantly eroded by suspended sediment or by "grinders," referring to large clasts that roll around (and around) the bottoms of potholes (e.g., Alexander, 1932; Hancock et al., 1998; Whipple et al., 2000; Barnes et al., 2004; Richardson and Carling, 2005; Springer et al., 2005). Based on laboratory experiments, Alexander (1932) interpreted that suspended sediment must be dominant because bedload becomes immoveable as potholes deepen, limiting the depth of incision. Interestingly, this is a statement of cover effects applied to a specific erosional form.

However, several arguments suggest that coarse gravel and cobbles, rather than suspended sediment, eroded the inner-channel pothole that developed during the spring 2005 snowmelt season. At a calculated discharge of about half the snowmelt maximum we measured vigorous coarse-sediment transport. The pothole is located along the channel thalweg, where coarse-sediment transport and erosion were most focused. Finally, following the cessation of snowmelt flow the pothole was full of coarse sediment. Sediment of this caliber would likely have been moving through the inner channel and in the pothole for the duration of snowmelt flow.

The pothole widened as it incised vertically, and widened more than the rest of the inner channel (Fig. 11, compare to Fig. 9). This is qualitatively consistent with results of Springer et al. (2005), who present a power-law scaling of pothole aspect ratios measured in the field and interpret from it that potholes widen laterally as they incise vertically. Centripetal acceleration would focus bedload impacts on the sidewalls and may account for the modest widening relative to downcutting we measured. In general, the vertical orientation of most potholes may also suggest that bedload rather than suspended sediment is commonly dominant in pothole erosion: impacts from coarse sediment would be more strongly focused on the pothole bottom (due to gravity) than abrasion by fine suspended sediment.

The question of whether potholes are eroded by suspended sediment or by bedload may be somewhat semantic, because it depends on where the criterion for suspension is defined. In order to exit a pothole, sediment must only be incipiently suspendable based on a combination of turbulent fluctuations and upward-directed flow leaving the pothole. In flume experiments, Johnson and Whipple (2007) documented the spontaneous formation of potholes that share morphological similarities with many observed in field settings. These potholes were eroded by a fine unimodal gravel that was primarily transported through the flume as energetic bedload, based on both direct observations and calculations of average flow conditions. However, the gravel was locally suspended by upward-directed flow exiting the pothole. We similarly interpret that the pothole in the Swett flume mouth was dominantly eroded by coarse sediment that was incipiently suspendable given local flow conditions but primarily transported as bedload along most of the channel.

For a pothole to initially develop, the erosion rate at the bottom must be greater than the lowering rate of the surrounding bedrock. However, locally higher erosion rates need not persist once the pothole is present. Vertical pothole incision (~0.5 m) was only slightly higher than incision along the inner channel directly upstream (Figs. 8 and 9). It is possible that pothole deepening became "limited" by the lowering of the upstream surface, resulting in similar vertical incision rates. A mechanism for this adjustment in rates could be cover in the bottom of the pothole: the pothole may have rapidly deepened to the point where the coarse sediment going in could no longer be incipiently suspended to exit the pothole. Further incision upstream would then have reduced the relative depth of the pothole, increasing the flow intensity inside the pothole and resulting in balanced erosion rates in the pothole and upstream. Our data are not sufficient to test this hypothesis.

If downcutting at the pothole bottom approximately equilibrated with the erosion rate upstream, then local bed roughness may have approached a relatively constant value. A natural limit on bed roughness is that it increases until the flow can just transport the local sediment load. Johnson and Whipple (2007) similarly argued that potholes represent locally high erosion rates, but that the overall effect of potholes may be to reduce the transport capacity of the flow (due to increased roughness), driving the channel toward alluviation and sediment load-dominated conditions. Potholes are distinct morphologically, but we interpret that pothole incision is governed by the same

basic feedbacks between morphology, sediment transport, and erosion that control channel incision in general.

### Implications for Natural Slot-Canyon Formation

Finally, we describe several natural bedrock canyons in southern Utah that have morphological similarities to our monitoring site. The comparison demonstrates two points. First, the pattern of transient incision seen at the flume mouth is a repeatable response to a "sudden" channel steepening. Second, channels in which flow is diverted over a steep bedrock slope occur naturally in this landscape, resulting in slot-canyon incision.

Several of the most dramatic slot canyons in southern Utah appear to have resulted from natural channel diversion events, starting from geometries broadly comparable to the initial condition at the Swett flume mouth. Figure 12A shows an aerial photograph of Coyote Gulch and tributaries, in particular Peek-a-boo and Spooky slot canyons (Wohl, 1998; Kelsey, 1999; Wohl et al., 1999) within the Escalante River drainage network. Both of these slot canyons, as well as the "narrows" just upstream along the mainstem Coyote Gulch, occur directly adjacent to wide but abandoned bedrock-walled valleys currently filled with windblown sand (Fig. 12A). Farther upstream these three active ephemeral channels all flow through wide bedrock valleys that are predominantly alluviated. A vegetated sand dune blocks the abandoned valley at the location where Spooky slot canyon begins (Fig. 12B). Downstream, channel width decreases through the slot canyon (Fig. 12C). The geometry suggests that the channels were diverted over steeply sloping Navajo sandstone slickrock surfaces. These slot canyons are examples of epigenetic gorges (Hewitt, 1998; Ouimet et al., 2007), which form when channels reincise through bedrock spurs after being diverted by a variety of natural processes including alluvial aggradation (James, 2004; Johnson et al., 2009), damming by landslides, and blocking by aeolian dunes as we interpret here.

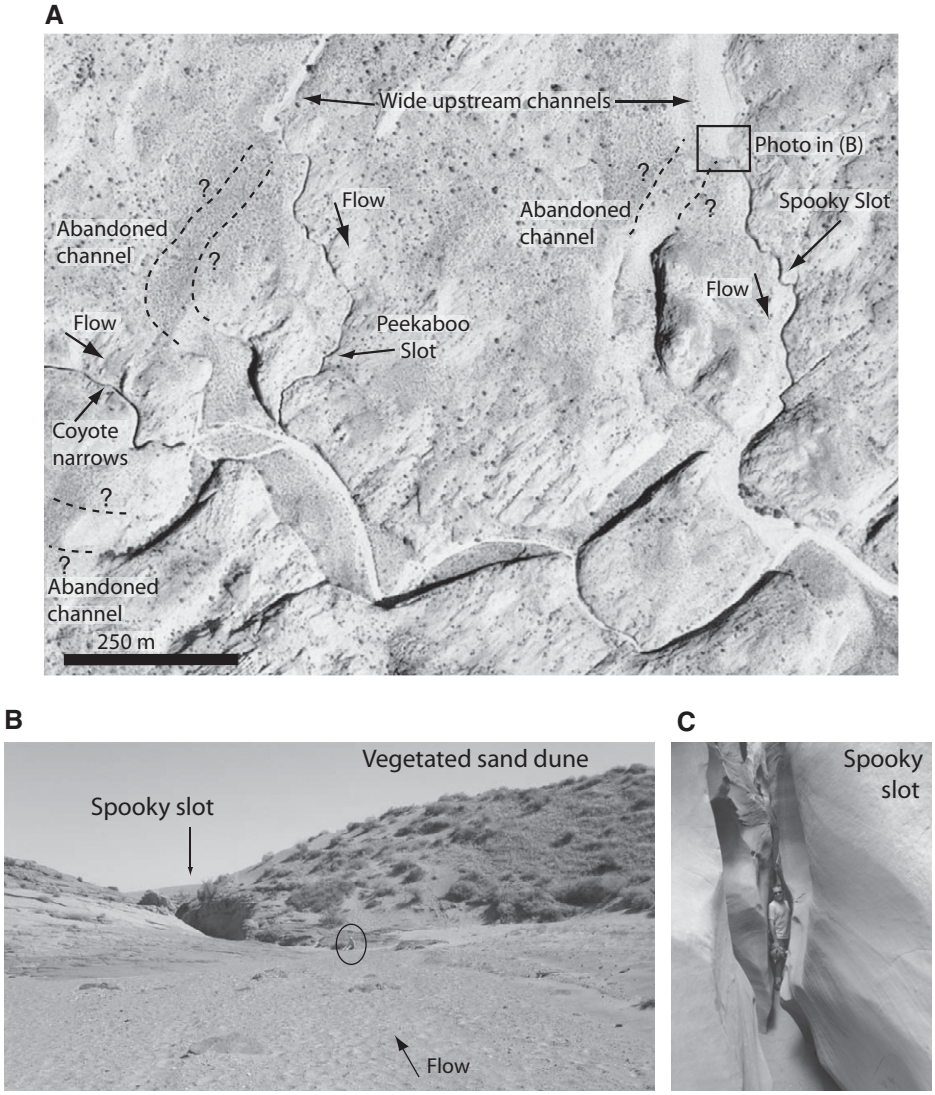


**Figure 12. (A) Escalante canyons, aerial photograph (crop of Big Hollow Wash, Utah, digital orthophoto quadrangle) of Coyote Gulch and tributaries including the slot canyons Peek-a-boo, Spooky, and Coyote Narrows (latitude 37.482N, longitude 111.216W). Rectangle shows approximate area of photo in 12B. Outlines of inferred abandoned channel segments are marked by dashed lines and question marks. (B) Photograph looking south of channel just upstream of Spooky slot canyon; note the start of the slot (shadowed bedrock). Sitting person is circled for scale. (C) A short distance downstream the active channel narrows greatly, forming Spooky slot canyon.**

A conceptual model for channel diversion and development at this location can be posed based on field observations. Initially, the mainstem and tributaries flowed in wide, presumably alluviated valleys. These valleys were then blocked, likely due to factors including the migration of aeolian sand dunes, alluviation behind sand-dune blockages, the intermittent nature of flow through these ephemeral channels, and subsurface water loss during flash floods. Eventually the blocked and aggrading channel overtopped its bedrock valley, and flow was diverted over a low point in the bare bedrock interfluve. Starting from this new initial condition of flow over a steep bedrock step, the channel incised a narrow and sculpted slot. These channel reaches are still in a transient state of adjustment, as evidenced by the dramatic differences in channel width and slope upstream and within the slot, similar to the Swett Creek flume mouth. Wohl et al. (1999) found no statistically significant lithologic controls on sidewall roughness and local width variations along Peek-a-boo slot canyon and the Coyote Gulch narrows, and interpreted that hydraulic processes controlled the morphologies of these channels. Similarly, we interpret that the morphology of these slot canyons reflects feedbacks outlined above between sediment transport, flow, erosion, and evolving channel morphology.

### Limitations and Future Work

Determining the generality of our relations between bedrock erosion and flood magnitude will await future data on flow, sediment transport, and bedrock incision from a variety of

landscapes and hydrologic settings. A limitation of our work is the short duration of the flow and erosion records and the small number of events captured, particularly for snowmelt flow. Another limitation is that the Swett Creek channel was initially perturbed anthropogenically, but this is also advantageous: the steep, artificial channel reach enabled rapid erosion and adjustment of channel morphology from natural forcing (flow and sediment transport). Furthermore, the slot-canyon case studies show that these patterns of erosion are reproducible and that they occur naturally in this landscape. Field areas that are locally perturbed by natural causes (e.g., wildfires, channel avulsions, and neotectonics) or humans (e.g., land use, construction, channel diversions, and reservoir base-level changes) are particularly useful for science-driven environmental monitoring, because signals will be larger, and the likelihood of observing changes will be much greater in a local landscape far from equilibrium. Moreover, transient landscapes capture more system dynamics than are observable under equilibrium conditions (e.g., Whipple, 2004).

Further monitoring of flow and sediment transport in semiarid and arid environments may provide unique opportunities for understanding basic mechanics of sediment transport. Our data suggest that hydrograph shape plays an important role in sediment-transport rates, timing of delivery, and patterns of entrainment and deposition. Previous work has shown that unsteady hydrographs affect bedload transport (e.g., Kuhnle, 1992; Laronne and Reid, 1993; Laronne et al., 1994; Lee et al., 2004). Monitoring sediment transport in a single natural channel reach that receives both snowmelt and flash-flood runoff would provide at least two benefits. First, the duration, shapes, and magnitudes of snowmelt and flash-flood hydrographs are close to being natural hydrograph end members, allowing effects of first-order hydrograph differences to be more clearly resolved. Second, monitoring the effects of different hydrographs within a single channel reach separates covarying controls of channel morphology that develop in response to long-term hydrological differences (e.g., bed armoring, pool-riffle sequences, and bedforms) and which are encountered when comparing separate channels from different climates.

## CONCLUSIONS

Rates of fluvial bedrock incision mimic rates of external landscape forcing (e.g., tectonic uplift and eustacy) when averaged over geological time scales, but local rates of channel downcutting into bedrock can be fast during the individual floods that actually drive bedrock incision: we measured up to 1/2 m of local vertical incision into bedrock over 23 days of snowmelt runoff (Fig. 9). Local channel morphology and high but not overwhelming rates of sediment transport enabled such a high local erosion rate. The local thalweg slope was high (~20%, Fig. 8), and the cross-sectional morphology of an inner channel focused flow and sediment transport over a narrow zone where almost all erosion occurred. While poorly constrained, field measurements demonstrated high rates of coarse-sediment transport. Additionally, preexisting inner-channel alluvium was entrained during this snowmelt runoff event, and so alluvial cover was not consistently present to mantle the inner-channel bed and inhibit bedrock erosion. Field observations also suggest that thresholds of detachment for abrading the local sandstone are negligible (Fig. 6).

Large floods can be much less erosive than small floods. In our channel, the peak discharge of the erosive but slowly varying snowmelt flow (~1 $m^3/s$) was nearly an order of magnitude smaller than the peak discharge of the largest flash flood (~9 $m^3/s$), which caused negligible bedrock erosion and extensive alluvial deposition (Figs. 3 and 5). Our observations demonstrate that bedrock erosion depends only indirectly on flood magnitude, and that floods with moderate peak magnitudes can cause more bedrock incision than floods with larger peak discharges.

Based on differences in erosion rates and surveyed changes in sediment deposition on the bed following flow events, we interpret that alluvial cover prevented local bedrock incision. Our field results further validate the "cover effect" at time scales of individual floods (Sklar and Dietrich, 2004; Turowski et al., 2007; Turowski and Rickenmann, 2008). Alluvial cover enables a negative feedback between erosion rate and increasing sediment flux, and is a mechanism by which flood magnitude is decoupled from erosion rate. The development of local alluvial cover was strongly influenced by the bedrock channel morphology as well as by the largest sizes of transported sediment, as much bed alluviation occurred behind interlocking cobbles that became wedged at the bottom of the inner channel. The inner-channel morphology may have also decoupled flow and sediment transport during higher discharge events. Bedload flux presumably increased with discharge, and the bedload would have been focused along the channel thalweg (in this case, the inner-channel bottom). However, basal shear stress along the inner channel would only minimally increase with increasing discharge once flow overtopped the inner channel. In this way, the inner-channel morphology would lead to an increase in local bedload flux relative to sediment transport capacity.

We did not directly monitor bedload transport, and so our data cannot explain why the snowmelt and flash-flood hydrographs caused such different erosional and depositional responses in this channel, which has an essentially unlimited sediment supply. We hypothesize that the flash floods were able to initially transport a higher concentration of large clasts (cobbles and some small boulders) due to turbulence associated with flash-flood bores and rapidly rising hydrographs. This relatively subtle factor in conjunction with the narrow inner-channel geometry may have been sufficient to cause sediment jams and local deposition during flash floods but entrainment during snowmelt runoff. Hydrographs may influence bedrock channel incision by influencing the timing of deposition and entrainment.

At our field site, bedrock channel incision occurred into a bedrock step created by humans during channel diversion, providing a well-constrained initial geometry of a steep, unchannelized bedrock slope. Erosion formed a narrow inner channel with rough sidewalls. This transient bedrock channel morphology is consistent with other natural slot canyons, in particular the Coyote Gulch narrows, Peek-a-boo slot, and Spooky slot canyons that are tributaries to the Escalante River in southern Utah. These canyons also incised from an initial condition of flow over steep, unchannelized bedrock slopes following channel diversions by sand dunes. The erosional topography in all of these cases is consistent with feedbacks between flow, sediment transport, and erosion observed in flume experiments (Finnegan et al., 2007; Johnson and Whipple, 2007).

Finally, our monitoring fortuitously captured the erosion of a bedrock pothole. We interpret that it formed by impact wear from coarse sediment rather than fine suspended load, although distinctions between bedload and suspended load may be less meaningful because localized incipient suspension of larger clasts is required to keep deposition from occurring inside the pothole.

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