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Transient response of bedrock channel networks to Pleistocene sea-level forcing in the Oregon Coast Range

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TRANSIENT RESPONSE OF BEDROCK CHANNEL NETWORKS TO PLEISTOCENE SEA-LEVEL FORCING IN THE OREGON COAST RANGE

A thesis submitted in partial satisfaction
of the requirements for the degree of

MASTER OF SCIENCE

in

EARTH SCIENCES

by

David J. Santaniello

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David J. Santaniello

Transient response of bedrock channel networks to Pleistocene sea-level forcing in the Oregon Coast Range

Abstract

Although sea level has fluctuated repeatedly over the Pleistocene by up to 120 meters, how (or whether) this cyclic base level forcing impacts the development of bedrock river profiles in tectonically active settings is poorly understood. A major reason for this uncertainty is that bedrock river channels in unglaciated locations are typically buried at their mouths under sediment resulting from the Holocene marine transgression, making direct observations impossible. Here I present a novel approach to constraining the influence of cyclical sea-level forcing on the development of bedrock river profiles in the Oregon Coast Range.

Using a 1 m LiDAR DEM, I estimate the depth to the buried bedrock longitudinal profile using measurements of valley width, elevation, wall slope, and present day river width for the Smith River and its tributaries where the river is currently buried by Holocene fluvial and marine sediments. Assuming the current river width is spatially and temporally constant, I calculate the depth to bedrock based on a trapezoidal geometry. In an effort to reduce the noise in our signal, I calculate linear regressions for valley and river width and hold the valley wall slope constant at its average value (~ 30 degrees) for all our calculations.

Several important observations stem from this analysis. First, the bedrock profile of the Smith River projects to the same elevation (~ -110 m) as the bench cut into the continental shelf by the sea-level low stand of the last glacial maximum, suggesting the bedrock river is currently graded to the low-stand elevation. Second, almost all of the tributary bedrock profiles plot tens of meters above the mainstem bedrock profile. One key exception is the significantly larger North Fork of the Smith River, which falls directly on the projected bedrock curve for the Smith. These observations suggest that sea-level low stands force incision of the entire bedrock river network, but that smaller tributaries, which are buried and/or flooded during highstands, cannot keep pace with the mainstem during these periods of brief down-cutting. Our results imply that difference in erosion rates between tributaries and the mainstem results in the formation of fluvial hanging valleys that are buried during marine transgressions.

Acknowledgements

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Introduction

The potential for rivers to grow steeper with higher rates of tectonically induced base-level fall implies that rates of river incision should evolve to match rates of rock uplift in actively uplifting ranges (Whipple and Tucker, 1999). This realization forms the basis for using patterns of river steepness (typically normalized for drainage area) to infer patterns in rates of rock uplift (e.g., Wobus et al., 2006). That said, many tectonically active rivers drain directly into the ocean, where sea-level has fluctuated by ~ 100 m over the Pleistocene. To the extent that the continental shelf slope is steeper than that of the river that drains onto it, sea-level falls will result in channel incision that propagates up from the coast during low stands (Schumm, 1993; Zaitlin et al., 1995). Alternatively, during periods of sea-level rise, bedrock channels must either aggrade to keep pace with sea-level rise or become estuaries. Although the effects of sea-level rise are widely recognized in coastal rivers where huge Holocene fills are commonly 10's of km long and 10's of m deep (Merritts et al., 1994), few studies have directly linked sea-level lowstands to pulses of river incision in bedrock channels (e.g., Pazzaglia and Gardner, 1993; Castillo et al., 2013). That said, flights of strath terraces in many coastal rivers (e.g., Personious, 1995; Merritts et al., 1994) testify to pulses of bedrock river incision over time that recur at approximately the same periods that sea-level varies over. Hence whether bedrock rivers incise in response to sea-level falls and, if so, how far this signal propagates upstream is still poorly understood in many settings. A key impediment to understanding the coupling between sea-level and bedrock incision is that during the present high-stand conditions, evidence for sea-level forcing of bedrock incision may be buried under thick Holocene deposits that blanket the seaward ends of large coastal rivers (e.g., Merritts et al., 1994; Reneau and Dietrich, 1991).

In order to investigate whether bedrock channels record pulses of incision due to sea level variations, I chose the Smith River in the Oregon Coast Range as my field site. This location is ideal to investigate this problem as the lower Smith River is currently aggraded in response to sea-level rise, but has extensive bedrock channel sections upstream where flights of terraces are widely recognized. In addition, the Smith River is covered by LiDAR topography data, and cuts through an area of homogenous bedrock, the Tyee formation. Below, I employ a high resolution LiDAR dataset of the study area to infer the bedrock channel elevation beneath the lower alluvial reach of the river using a simple geometric relationship. This geometric projection of the buried bedrock channel allows for an analysis of the complete bedrock channel profile of the mainstem and tributaries of the Smith River and is used to investigate whether the bedrock channel of the Smith River is graded to the last sea level lowstand, if the mainstem bedrock channel records pulses of incision in the form of knickpoints, and finally if there are any discrepancies between the mainstem and tributary bedrock channel profiles.

Geologic Background / Field Site

The Smith River, OR, is an ideal location to test whether transient signals of sea level forcing can be found in bedrock landscape for several reasons. First, the confluence of the Smith and Umpqua Rivers is approximately 20 river km from the present day coast and the lower Smith River exhibits clear evidence for aggradation in response to sea-level forcing that has been recognized in other nearby coastal rivers (e.g., Reneau and Dietrich, 1991). Second, this region of the Oregon Coast Range is unglaciated and has experienced steady tectonic forcing rates over the recent geologic past (Kobor and Roering, 2004). Long-term lowering rates for this region have been estimated to be on the order of 0.1 mm/yr (Heimsath et al., 2001; Personius, 1995; Reneau and Dietrich, 1991). In addition to a constant tectonic forcing, the Smith River and its tributaries are entirely contained in a single rock unit, the Tyee Formation (Kobor and Roering, 2004). The Tyee Fm is a well-indurated turbiditic sandstone. Because the bedrock of the study area is entirely comprised of the same formation, any signals recorded in the bedrock profiles most likely are not caused by lithologic heterogeneity. This study site is also well suited to studying sea level signals because the elevation of sea level during the last glacial maximum (LGM) has been modeled for this region (Mitrovica, personal communication).

The morphology of the Smith River itself also lends itself to be a good study site. The lower ~ 40 km of the river is an alluvial reach. The bedrock channel has been buried by sediments during the last marine transgression, as is typical in coastal Oregon and northern California rivers (Merritts et al., 1994; Reneau and Dietrich, 1991). The geomorphic signature of these reaches is a valley width that is much wider than the active river width (Reneau and Dietrich, 1991). This section of the river and its tributaries will be used to test whether sea level signals propagate into the landscape by estimating the depth to buried bedrock channel profiles. The upper ~65 km of the Smith River is an exposed bedrock reach and observations from this section are useful in visualizing the morphology of the lower Smith River during sea level lowstands. The marked difference in slope of the longitudinal profile also pinpoints the abrupt transition from a bedrock channel to an alluvial channel (figure 2). In an addition to an abrupt change in slope, the grain size transitions from cobbles to silt and mud.

Finally, the Smith River is an excellent study site because it is almost entirely covered by high resolution light detection and ranging (LiDAR) topography data acquired by the Oregon Department of Geology and Mineral Industries (DOGAMI). The 1 m^2 spatial resolution attained for the resulting digital elevation model (DEM) allows for precise measurement of the variables needed to constrain the subsurface bedrock topography.

Methods

In order to test whether the mainstem of the Smith River responds to sea level forcing, I infer the buried bedrock channel depth in the alluvial reach using valley morphology measurements based on a 10 m DEM for the lower 10 km of the reach and a 1 m LiDAR DEM for the remainder of the study site. I create 41 transects across the valley at approximately 1 km spacing for the 40 km long alluvial reach and measure hillslope angle, valley width and mean elevation, and channel width (figure 3). These variables are used to project a trapezoidal geometry into the sediment filled valley to find the elevation of the buried bedrock channel. I assume the hillslope angle is equivalent and constant on both valley walls, the channel is located in the center of the valley, and that the modern channel width is comparable with the paleo-channel width (figure 4). The buried bedrock channel elevation can be found with the equation:

$$z_{bedrock} = z_{valley} - \frac{1}{2}(w_{valley} - w_{channel})tan\theta \quad (1)$$

where $z_{bedrock}$ is the elevation of the buried bedrock channel, z_{valley} is the mean elevation of the valley, w_{valley} is the width of the valley, $w_{channel}$ is the width of the modern channel, and θ is the mean hillslope angle.

The hillslope angle is measured for both sides of the valley walls by finding the slope of a fitted line through the data for the 41 transects in the alluvial reach, and the mean value from this dataset is used as the constant hillslope angle for the buried bedrock channel calculations at each transect. In order to check whether this constant hillslope angle is a good approximation, I compare this mean value to a slope histogram for the 10 m DEM (figure 5). Using these transects, I also measure valley width, mean valley elevation, and channel width (figure 5).

These variables are then used to calculate the buried bedrock channel elevation at the location of each transect as a function of longitudinal distance from the Smith-Umpqua confluence. Using these points, I create a buried bedrock channel longitudinal profile for the Smith River. In addition, because the buried bedrock channel elevation at any given point depends strongly on the valley and channel width of its associated transect, I use a linear regression to monotonically increase these variables downstream and to reduce the noise in the bedrock profile. Additionally, I exclude river width measurements from the 10 m DEM because of the poor resolution of the channel in this section and extrapolate the river width values from the 1 m LiDAR DEM for this reach (figure 6).

This method for creating buried bedrock channel profiles is repeated for the nine tributaries that enter into the alluvial reach of the Smith River (figure 1). Transects are taken approximately every 200-400 m for the smaller tributaries and every 500 m for the North Fork of the Smith River (figure 3). Most of the tributaries are not covered by the 1 m LiDAR DEM, and no river widths were measured on the 10 m DEM. River width was measured for tributaries R5, L3, and the North Fork because they are fully covered by the 1 m DEM. However, a channel width of 10 m will only increase the buried bedrock elevation approximately 3 m relative to a river width of 0 m, and all the tributaries except for the North Fork have river widths on the order of 5 m. It is only necessary to apply the river width correction to the North Fork and the mainstem.

Results

The mean hillslope angle calculated from the 41 transects is 30° (figure 5), which matches the mode of the 10 m DEM slope raster (figure 5). The linear regressions of both valley and channel width for the Smith River are shown in figure 6. Because the river width could not be measured on the 10 m DEM, the linear regression is extrapolated from the 1 m data (figure 6b). The effect of regressing both valley and channel slope on the longitudinal profile of the buried bedrock channel is shown in figure 7. The scatter present in the raw profiles (figure 7a) is strongly reduced. Additionally, the difference between the river width modified bedrock elevation (red line) and river width ignored elevation (blue line) highlight the importance of measuring river width on the mainstem (figure 7b). The buried bedrock profiles of the nine tributaries that enter the alluvial reach of the Smith River are shown in figure 8. In order to compare the elevation of the tributaries at the confluence with the mainstem, the elevations of the tributary buried bedrock channel profiles are plotted on the Smith River bedrock channel profile (figure 9, Table 1).

Table 1: Tributary Data

Tributary	**Confluence River km**	**Drainage Area (km^2)**	**Bedrock Channel Elevation (m asl)**	**Hanging Valley Height (m)**
Trib R1	2.5	8.5	-22	83
Trib R2	3.7	10.4	-72	26
Trib L1	5.1	17.1	-70	25
Trib R3	8.0	12.2	-32	54
Trib L2	10.2	7.1	-31	53
Trib R4	10.4	11.6	-32	52
Trib R5	13.3	11.6	-31	45
North Fork	24.0	130.2	-45	11
Trib L3	24.9	9.4	-19	38

Table 1: The tributary data listed above are plotted in figures 9 and 16.

Discussion

The validity of the methods presented above depends on four key assumptions: (1) the hillslope angle is constant for the study area, (2) the buried bedrock channel is located in the center of the valley, (3) the modern channel width is comparative to the width of the buried bedrock channel, and (4) the valley and modern channel width can be modeled accurately with linear regressions. Below I discuss these assumptions and the validity of the method described above. With a valid method, I then discuss the shape of the buried bedrock channel longitudinal profile, the tributary profiles, and how the combination of these two demonstrates the transient signals of sea level change in the landscape.

Assumption 1: Hillslope Angle

I assume the valley walls can be modeled with a characteristic planar hillslope angle that is representative of much of the Oregon Coast Range (e.g., Roering et al. 2007). The hillslope angle in equation 1 is held constant at 30° for all buried bedrock elevation calculations. The representative cross section in figure 5a qualitatively shows the valley walls can be approximated with this constant angle. This angle is also compared to the slope histogram for the 10 m DEM (figure 5b). The peak of the histogram is also at an angle of 30°. When the slope raster is viewed in map view, it is clear the valley walls on both sides of the river canyon slope at approximately 30°.

Assumption 2: Buried Bedrock Channel Location

Another key assumption of this method is that the buried bedrock channel exists in the center of the modern alluvial valley. This assumption allows a simple geometric relationship between the buried bedrock channel elevation and the variables of hillslope angle, valley elevation, and valley and channel widths (figure 4), and equation 1 is derived from this geometry.

Bedrock rivers in this setting do meander and migrate laterally as well as incise vertically (Johnson and Finnegan, in press). The strath terraces preserved on the inside of meander bends in the bedrock reach of the Smith River attest to the ability of this river to erode both horizontally and vertically (Personius, 1995). In order to reduce the uncertainty of the lateral migration of the bedrock channel, the 41 transects were selected to mostly avoid the inside of meander bends where there is typically a strong asymmetry in hillslope angles (figure 3). That said, some of the noise in the raw valley depth measurements may be related to the sinuous valley geometry here.

Assumption 3: Modern vs Paleo-channel Width

Another assumption of the valley geometry is the width of the buried bedrock channel. I assume the modern alluvial channel width is representative of the bedrock channel width buried beneath it at any given transect. This assumption will most likely not be globally valid; however, locally at this field site it is reasonable. The width of the channel does not change drastically at the transition from bedrock to alluvial (figure 10). Thus, I expect the buried bedrock channel to continue to increase in width downstream, just as the present bedrock reach channel width increases downstream. Because the alluvial channel widths seem to widen at the same rate as the upstream bedrock channel widths, I assume the alluvial channel width is a good indicator of the width of the bedrock channel buried beneath it.

Assumption 4: Linear Regressions of Valley and Channel Widths

The elevation of the buried bedrock channel depends heavily on the valley and channel width (equation 1). However, both the valley and channel can widen or narrow due to local factors. In order to reduce the noise in the buried bedrock channel longitudinal profile, I model the valley and alluvial channel width data with linear regressions (figure 6). The linear regression forces the valley and channel widths to monotonically increase downstream and allow the larger trend of the bedrock channel profile to be interpreted (figure 7). Defining the valley width as linear function of downstream distance is a fairly simple interpretation. However, channel width is normally modeled as a power law (Leopold and Maddock, 1953**)** such that $w = aQ^b$ or aA^b, where $b=0.4$ for bedrock channels (Whipple, 2004), and Q is a representative discharge whereas A is drainage area. For the alluvial section of the Smith River, both the linear regression and power law best fit describe the data with the same significance (figure 6). The major difference between these two fits occurs during the extrapolation of the widths measured on the 1 m LiDAR DEM to the lowermost 10 km of the Smith River not covered by LiDAR. For this extrapolation, the linear regression gives much more realistic values while the power law returns unreasonably high values for the channel width. Therefore, I model the channel width as a linear function of distance instead of a traditional power law.

Buried Bedrock Channel Longitudinal Profile

The modern profile of the Smith River is graded to current sea level (figure 2). As sea level changes, the alluvial reach of the river will adjust its longitudinal profile to base level by either excavating or depositing sediment. The sea level signal will not be transferred to the bedrock paleo-channel buried beneath the alluvial reach. However, this buried bedrock channel should be graded to some sea level lowstand before the valley fill was deposited. During the last glacial maximum (LGM), sea level was approximately 120 m below present level (Mitrovica, personal communication). The bathymetric profile extracted off the coast of the Umpqua River shows a bench cut into the shelf at approximately -120 m. I interpret this bench as the base level during the LGM. The predicted bedrock channel profile for the Smith River also grades to approximately -120 m at its mouth (figure 9). This similarity between the bedrock channel profile and eustatic sea level record suggests the buried bedrock channel is graded to the last lowstand sea level elevation.

Though the buried bedrock channel seems to be graded to -120 m, I do not suggest the river incised 120 m in 18,000 years. Instead, I suggest pulses of incision occurred as knickpoints were generated during previous sea level lowstands as well as during the LGM. It is widely recognized that as the period of glaciations increased over the last half of the Pleistocene, the magnitude of sea level fluctuations increased (Rohling et al., 2014). In fact, Rohling et al. (2014) estimate that sea level fluctuations of -100 m have only occurred during the last 1 Ma. Each sea level lowstand would generate a knickpoint and a pulse of incision. Due to the convergence of the Juan de Fuca and North America tectonic plates in the Cascadia Subduction Zone, steady uplift of the study region has occurred. Because the slope of the shelf below ~-120 m is steeper than the profile of the river upstream, uplift permits each successive sea level lowstand to form a knickpoint capable of propagating upstream. Depending on knickpoint retreat rates, length of lowstands, and the rate of the subsequent sea level rise, a knickpoint could be buried by either water or sediment and would become inactive if it had not propagated out of the zone affected by sea level rise. During the successive lowstand, this knickpoint would be excavated and reactivated, continuing a pulse of incision upstream. This hypothesis of a series of knickpoints propagating upstream will be further discussed in the hanging valleys section of the discussion.

Tributary Profiles

In addition to the mainstem, the nine tributaries that enter the alluvial reach of the Smith River also record important evidence of transient signals propagating through the bedrock channel network. Yet more caution is required when interpreting these profiles compared to the mainstem. Below I highlight three examples that demonstrate this need to understand the factors that can alter the buried bedrock profiles.

The lowermost tributary of the Smith River is R1, the first tributary from the confluence of the Smith and Umpqua that enters on the river right side of the Smith River (figure 1). In detail, valley width of the first kilometer of the profile is much wider than the upstream reach of the tributary (figure 11). This lowermost kilometer extends into the floodplain of the Smith River, and the valley width measured here is most likely strongly modified by erosion due to the mainstem. Therefore, I discount this portion of the profile and begin measurements for tributary R1 at the 1 km mark.

Tributary L1, the lowermost tributary on the river left side, is also affected by the mainstem (figure 1). A detailed map shows the lowermost 1.8 km of the tributary are significantly wider the upper 4 km (figure 12). This over-widened area is fully contained in the tributary valley, and lateral erosion from the current mainstem cannot fully explain the valley morphology. However, there are many examples of meander cutoffs throughout the study site (figure 13). These cutoffs vary in size and shape, but two examples from the upstream bedrock reach provide a good visible analog to describe the widened lower valley of tributary L1 (figure 14). These smaller, tight meander cutoffs significantly widen the valley beyond what the tributaries that feed into them could accomplish. Tributary L1 also has a step function change in valley width that cannot be explained by lateral migration of the tributary. I interpret this widening as a meander cutoff of the mainstem, and I do not include the lower 1.8 km into my buried bedrock channel analysis.

The final tributary I use as an example of the difficulties in performing this analysis is the North Fork of the Smith River. The North Fork is an order of magnitude larger than the other tributaries and has an alluvial reach of 14 km as well as a bedrock reach that extends past the 1 m LiDAR coverage area. The other smaller tributaries range from 1.5 – 7 km long and are fully filled with sediment. The North Fork enters the mainstem at the outside of a meander bend and the first approximately 2 km of the tributary are affected by both the influence of the mainstem and a meander cutoff (figure 15). Therefore these data points are excluded from the bedrock analysis (figure 15). Additionally, 4 km of the river from kilometers 5 – 8 are also excluded from the dataset. This area represents a lobe of over-widened valley due to meander cutoffs, and if included would cause a large depression in the bedrock profile. Therefore, it is difficult to interpret the bedrock profile of the North Fork. However, this analysis can still provide a general profile of the buried bedrock channel that is sufficient to compare with values from the mainstem.

Smith River Hanging Valley Profiles

After examining the tributary profiles and correcting for effects due to the mainstem, such as over-widening the lower portion of the tributary valleys, it is clear most of the tributary buried bedrock profiles project between -20 and -40 m at the confluence with the Smith River (figure 8). If the tributaries lower at the same rate as the mainstem, these bedrock channels should project to the same elevation; however, when the bedrock channel elevations of the tributaries at the confluence are plotted along the Smith River profile, the bedrock channels of the tributaries plot well above the mainstem bedrock channel (figure 9). Most tributaries plot over 30 m higher than the mainstem bedrock channel at their confluence, with the North Fork being a notable exception. The data suggest there are hanging valleys at the junction of the tributary bedrock channels that are presently buried by 20 – 70 m of sediment (figure 9).

I compare both the tributary valley elevation and hanging valley height with their respective computed drainage areas (figure 16). The tributary confluence elevation shows no clear pattern with drainage area, though the absolute elevation of the tributary valleys should not be affected by the processes of differential incision between the mainstem and tributary channels. Instead, the tributary elevation normalized to the mainstem bedrock elevation at the confluence (the hanging valley height) is expected to have a relationship with this incision differential (figure 16b). For tributaries with a drainage area less than 20 km^2, no relationship between drainage area and hanging valley height is observed. However, the North Fork is an order of magnitude larger than the other tributaries and has a calculated drainage area of ~ 130 km^2 based on the 10 m DEM. The associated hanging height is 10 m, much lower than the other tributaries.

Because the lower 2 km of the North Fork were not included in the bedrock channel analysis, I use a best fit line to the data to extrapolate the elevation of the buried bedrock profile at the mouth of the North Fork. Using this value, -45 m, the hanging valley height is approximately 10 m. Due to the difficulty in calculating the bedrock channel profile of the North Fork, it is possible this hanging valley height could be even lower than 10 m.

The hanging valleys interpreted under the alluvial reach of the Smith River show that incision of the mainstem due to sea level forcing has not propagated up into the tributary landscape. For example, the height difference between the Smith River bedrock profile and its most downstream tributary, R1, is approximately 80 m. If the mainstem lowers at the long term erosion rate of ~0.1 mm/yr, then this hanging valley reflects 800 ka of differential incision, assuming no incision upstream of the buried knickpoint. Therefore, these hanging valleys signify that the incision associated with the last several and perhaps up to the last 8 sea level lowstands has not propagated into the tributary network.

Formation of Hanging Valleys

One way to allow the mainstem of a river to incise faster than its tributaries is through knickpoint migration. When the base level of a river is lowered, such as during a sea level lowstand, a knickpoint at the mouth of the river is created and the river responds by incising to the new base level. As stated before, the magnitude of sea level fluctuations have increased in the latter half of the Pleistocene and sea levels below -100 m may have only occurred in the last 1 Ma (Rohling et al., 2014). These base level drops would have produced a succession of knickpoints capable of propagating upstream and inducing waves of incision on the mainstem.

Knickpoint propagation is an active area of research in the geomorphology community (Berlin and Anderson, 2007; Berlin and Anderson, 2009; Crosby and Whipple, 2006; Lamb et al., 2007; Weissel and Seidl, 1998). The processes that control knickpoint retreat and retreat rates are vital to developing hanging valleys between the mainstem of a river and its tributaries, which are significantly smaller in terms of drainage area. This difference in drainage area significantly affects the knickpoint retreat rate. Berlin and Anderson (2007) and Crosby and Whipple (2006) both observe that knickpoint retreat rate is proportional to drainage area. Knickpoint retreat can also halt at low drainage areas (Crosby and Whipple, 2006) or at large contrasts in drainage area (Wobus et al., 2006). These observations suggest waves of incision from propagating knickpoints may be transmitted upstream in the mainstem but may stall and not continue up the tributary reaches, leading to the formation of hanging valleys (Crosby et al., 2007; Wobus et al., 2006).

The process by which drainage area controls knickpoint migration is not currently understood but there are at least three possible hypotheses. Crosby and Whipple (2006) suggest knickpoint migration is inhibited at a small drainage area due to the reduced capacity of the river to incise because of lack of sufficient sediment and water input into the channel. However, Stock and Dietrich (2003) show that bedrock terraces typically do not extend upstream to the portion of the river that experiences debris flows. Because bedrock terraces are commonly formed by retreating knickpoints (Finnegan, 2013), their data supports the hypothesis that knickpoints do not propagate into the debris flow regime of the river network, possibly because the presence of large immobile boulders prevents knickpoint retreat. A third hypothesis stems from the saltation abrasion model of Sklar and Dietrich (2004). If the slope increases at a tributary junction due to knickpoint retreat, the hop length of the abrading bedrock particles will increase, thereby reducing bedrock incision. Each of these hypotheses could explain how a propagating knickpoint on the mainstem Smith River could lead to hanging valleys at the tributary junctions.

However, hanging valleys are not observed in the upstream bedrock reach of the Smith River. Although these tributaries of are similar size to the streams in the alluvial reach, the tributaries grade into the bedrock mainstem. Bedrock terraces are observed throughout the bedrock reach of the Smith, therefore the stalling of knickpoints at tributary junctions cannot fully explain the hanging valleys observed in the alluvial reach.

Another factor that differentiates the tributaries in the bedrock and alluvial reaches is the amount of time they are sub-aerially exposed and therefore able to erode their beds. The tributaries in the alluvial reach are subject to burial by either water or sediment during sea level highstands, which is demonstrated by their present buried state. Thus for much of the Pleistocene, these tributaries are unable to incise. Moreover, before incision can occur during a lowstand, the evacuation of sediment in the buried valley must occur first. Because knickpoint propagation rate is proportional to drainage area, I interpret the presence of hanging valleys in tributary valleys (but not the mainstem) as a manifestation of the frequent burial and hence erosional inefficiency of these channels over the Pleistocene. Although the mainstem has experienced the same burial history, it requires much less time to propagate a base-level signal upstream, all else being equal. The fact that only the two largest channels (mainstem Smith and North Fork Smith) show no evidence for buried knickpoints supports this interpretation.

Conclusion

Other studies have attributed propagating waves of incision in bedrock rivers to eustatic sea level fall (Castillo et al., in press; Pazzaglia and Gardner, 1993). However, it can be difficult to show a relationship between sea level forcing and incision in current bedrock canyons because of the large distance between the paleo-shoreline and the canyon. In fact, it has been challenging for geomorphologists to assign evidence of intermittent channel incision to sea level forcing (Finnegan and Balco, 2013; Merritts et al., 1994). Selecting the Smith River in the Oregon Coast Range as the study site largely reduces the distance between the paleo-shoreline and bedrock canyon compared to river on the east coast of the United States. This distance is further reduced by combining high resolution topographic data obtained from LiDAR and a simple geometric approach to estimate the buried bedrock channel beneath the present day alluvial reach of the river.

When the elevations of the tributary bedrock channels are superimposed on the profile of the inferred mainstem Smith River buried bedrock profile, the hanging valleys separating the mainstem and tributary drainages are noticeable. Because the mainstem is graded to the lowstand during the last glacial maximum, this observation suggests the tributary drainages are not in equilibrium with the current imposed base level and long lived transient signals due to sea level fluctuation are recorded in the landscape. Further, the lack of hanging valleys in the bedrock reach of the Smith River suggest more than just drainage area affects the formation and preservation of hanging valleys. The opportunity of the knickpoint to erode, or exposure time, may also be a key factor in preserving hanging valleys. At present, the hanging valleys observed in the alluvial reach are all buried under water and sediment due to the damming effect of the current sea level highstand. These periods of burial during highstands significantly reduce the exposure time of the knickpoints in the alluvial reach compared to the bedrock tributaries well upstream of the backwater length and may have a significant role in maintaining hanging valleys on the lower Smith River.

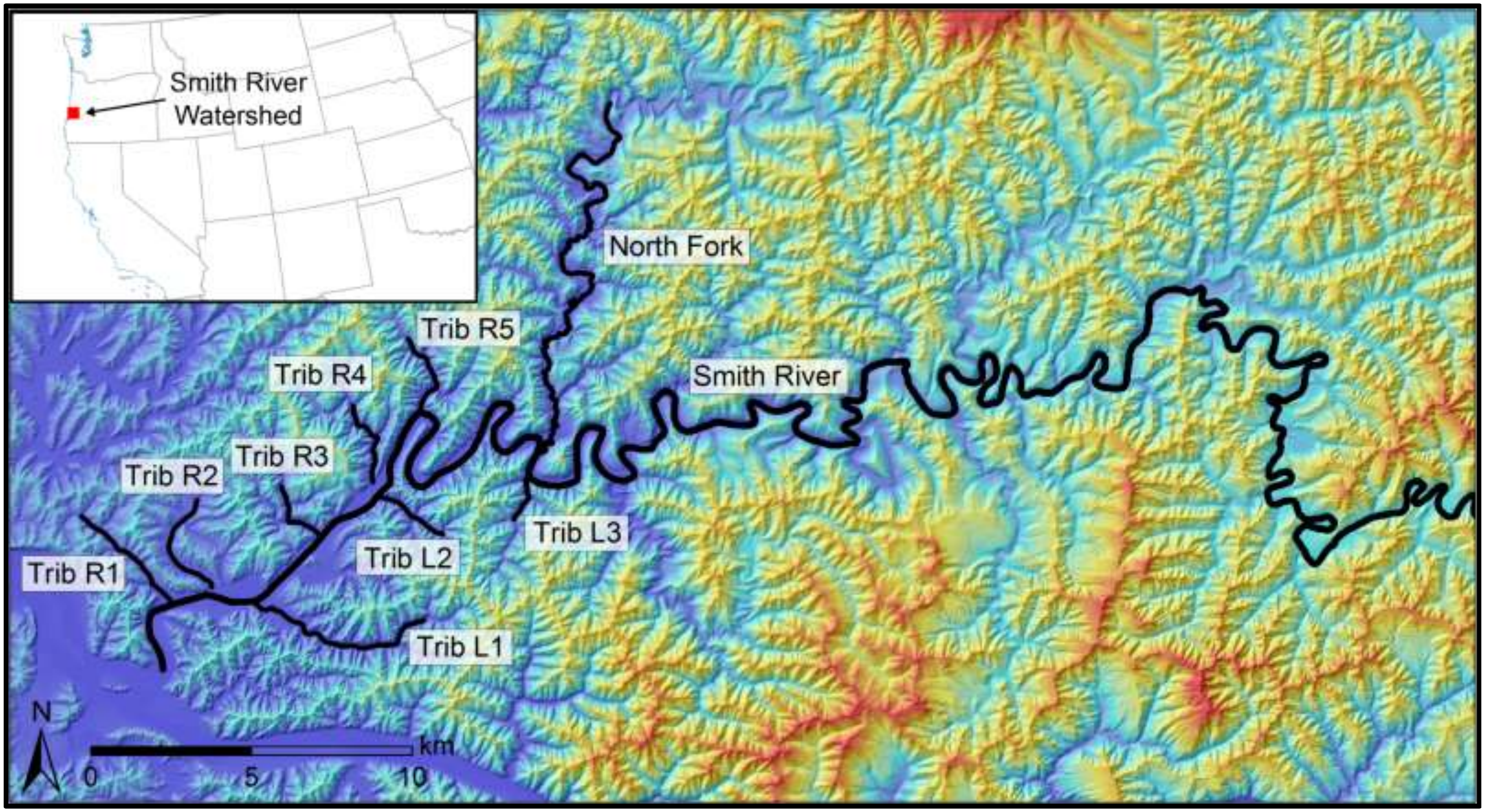

Figure 1: DEM of Smith River and tributary network. Inset map shows approximate location of the Smith River watershed in the Oregon Coast Range. The confluence with the Umpqua River in the lower left is approximately 20 km upstream of coast.

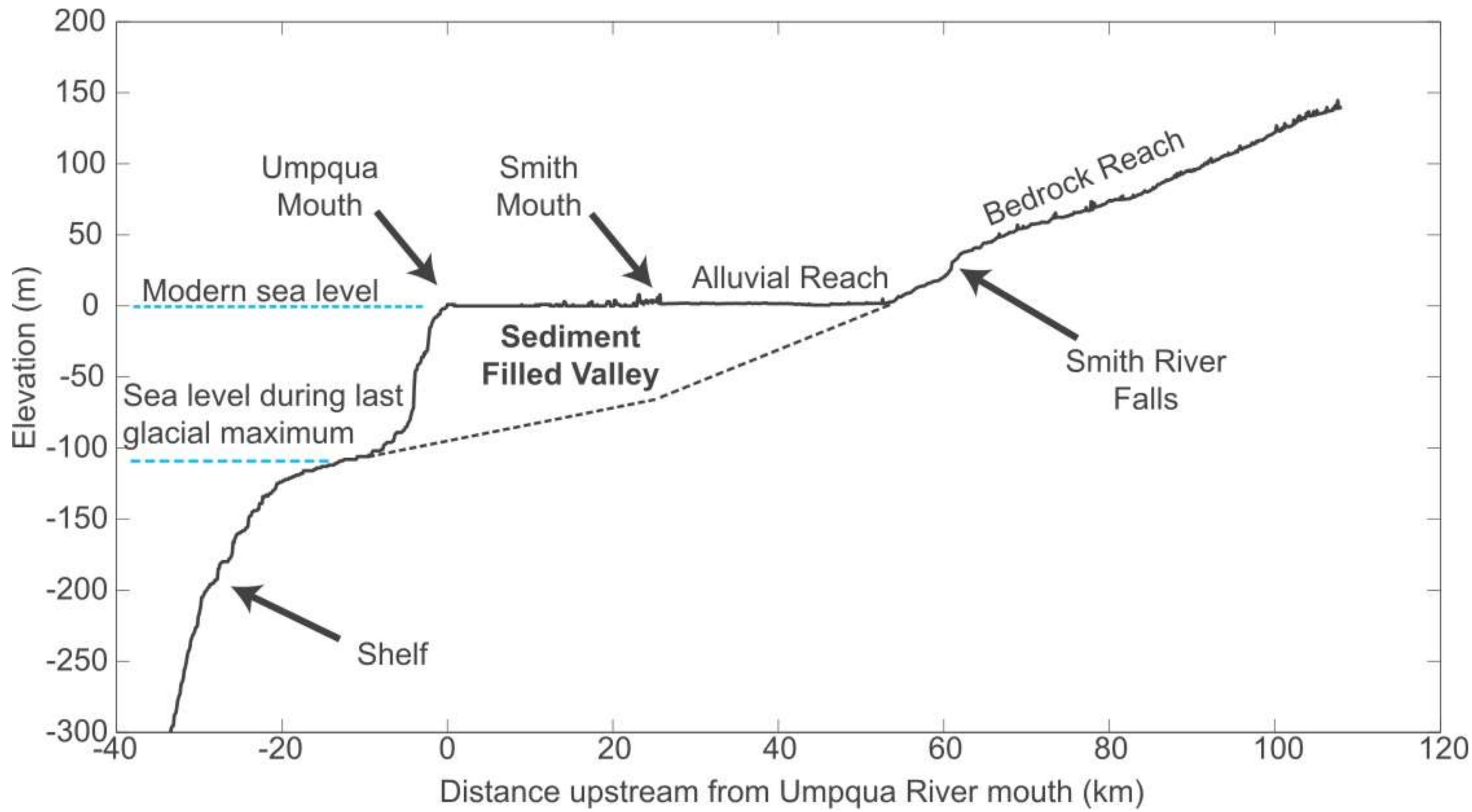

Figure 2: Longitudinal profiles of the Smith and Umpqua rivers, with added bathymetry of the outboard shelf. Dashed lines schematically represent the wedge of sediment that has been deposited in the present day alluvial section of the profile.

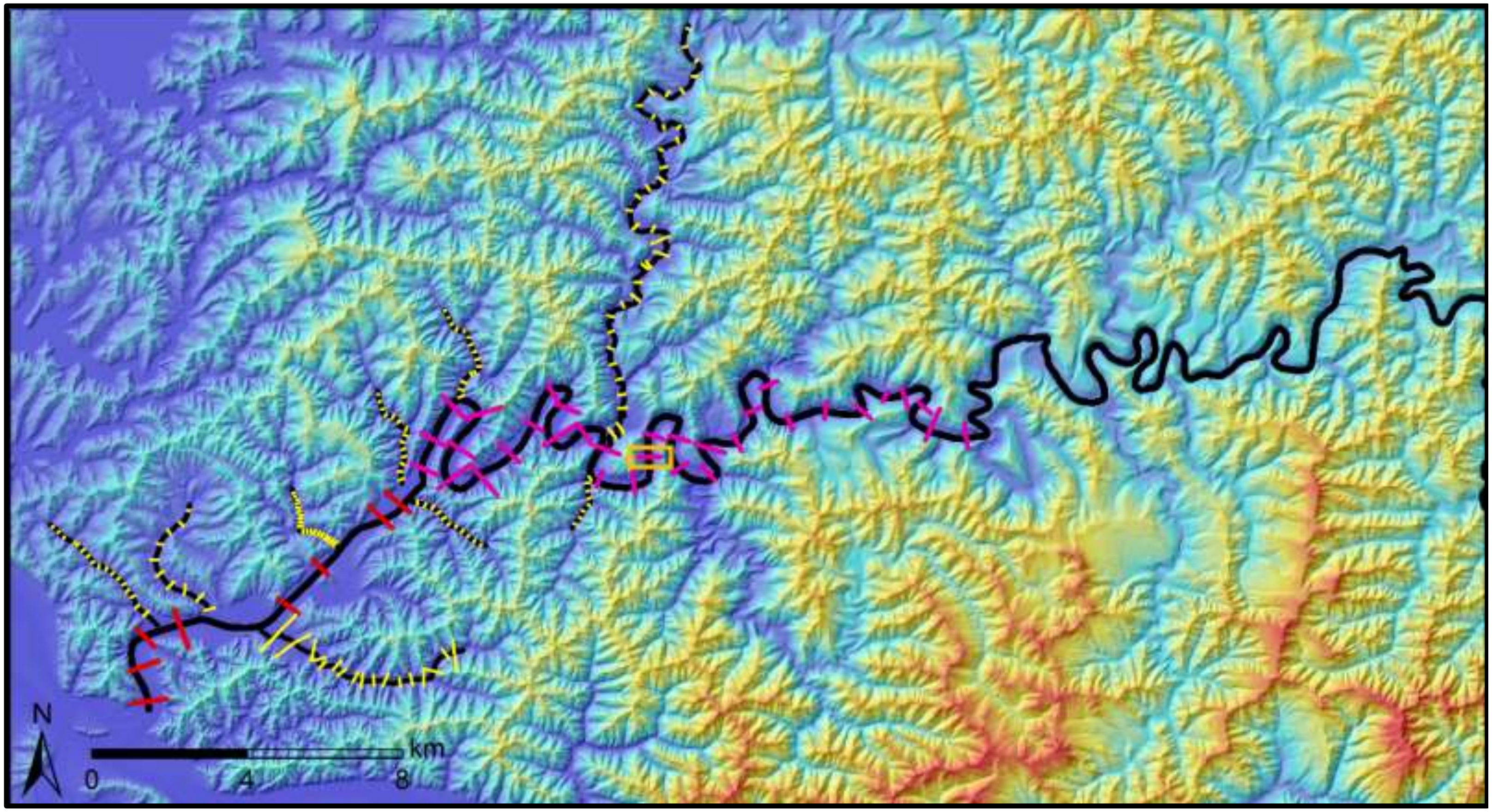

Figure 3: Base map of study area. Red and magenta lines denote transects where valley and river width was measured on the Smith River on the 10 m DEM and LiDAR dataset, respectively. Yellow lines similarly denote where valley and river width (if applicable) was measured on the tributaries. Orange box highlights Transect 25, which is used as a representative cross section in figure 5A.

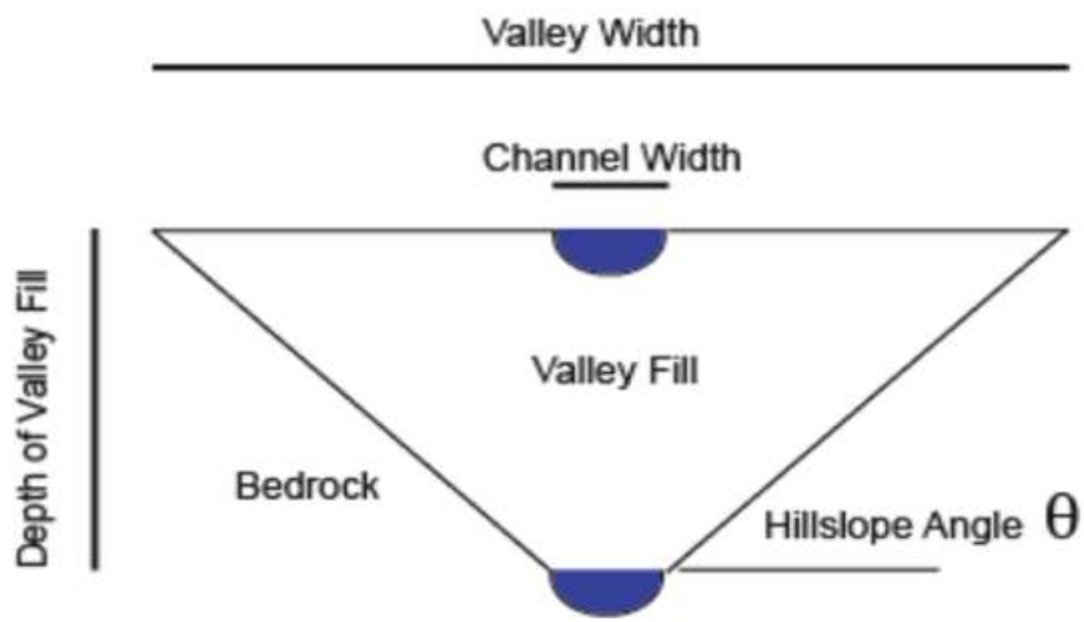

Figure 4: Cartoon detailing how valley geometry is used to infer the depth to buried bedrock. Hillslopes at a constant angle of 30 degrees are projected down from the valley walls to forma a trapezoid with the projected channel width. The channel is assumed to be in the center of the valley. This trapezoid is used to find the depth of the valley fill.

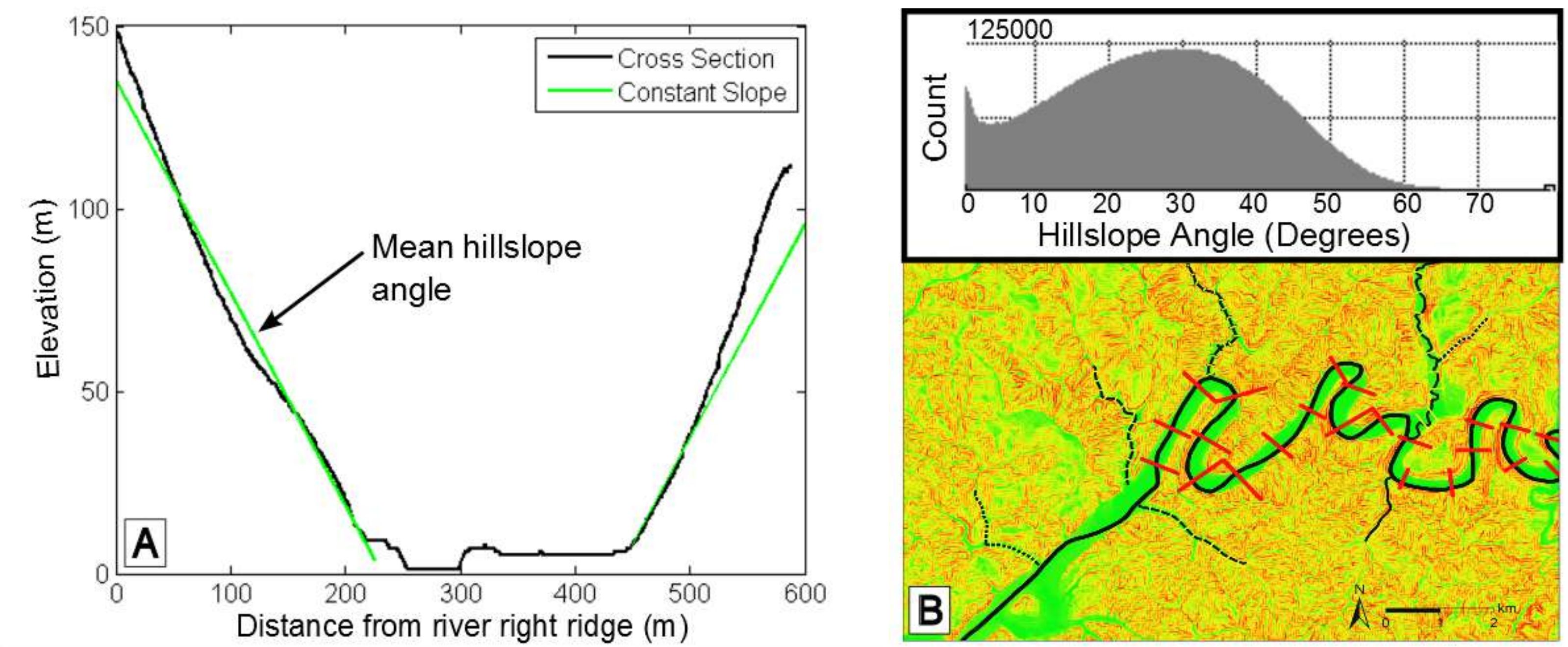

Figure 5: (A) Representative cross section of Smith River valley morphology. The hillslope angle for the bedrock depth calculation is held constant at the calculated mean of 30 degrees from LiDAR cross sections, which corresponds well with the histogram of the 10 m DEM slope raster of the study area (B).

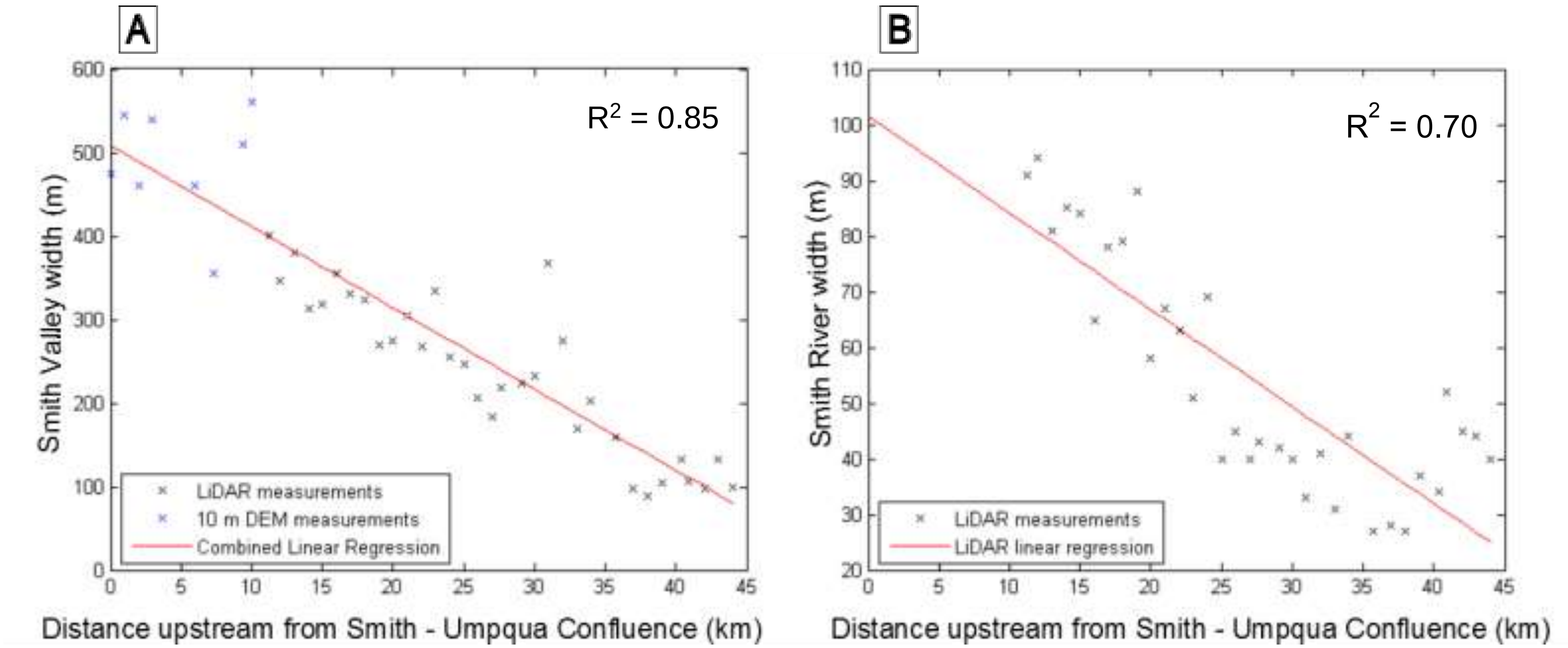

Figure 6: (A) Linear regression of mainstem valley width. (B) Linear regression of mainstem river width. Channel width could not be accurately measure on 10 m DEM, so the linear regression for 1 m LiDAR measurements has been extended in B.

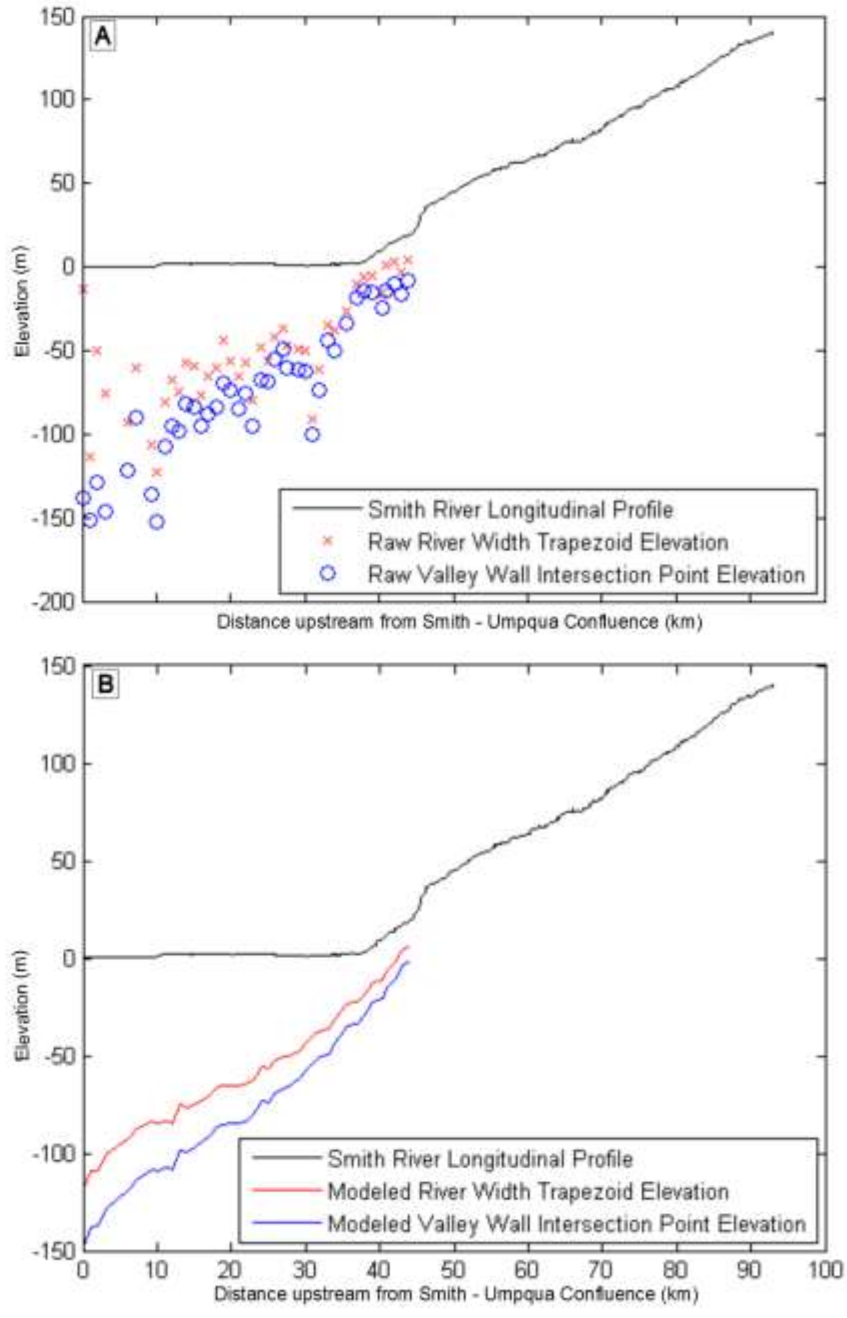

Figure 7: (A) Projected buried bedrock profile calculated with the raw data. Red x's denote the elevation of the supposed bedrock channel using a trapezoidal geometry. Blue circles ignore the channel width and project the valley walls to a point. (B) Employing the linear regressions smoothes the projected profile. The difference in elevation between the red and blue lines highlights the importance of channel width and a trapezoidal geometry.

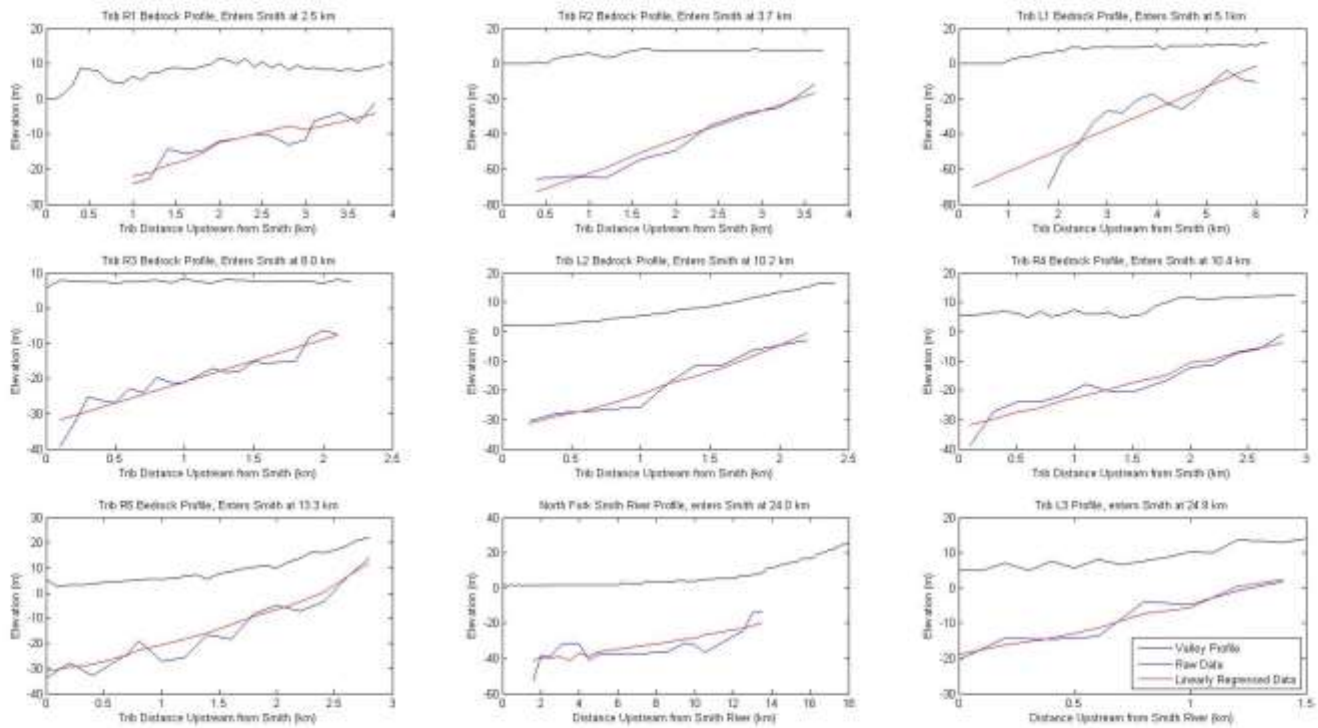

Figure 8: Buried bedrock profiles of alluvial tributaries. Blues lines represent the raw data and red lines represent the linearly regressed data (the linearly regressed data is used in further analysis of hanging valleys). Because measurements on the lower 2 km of Trib L1 were not credible, the linear regression of the data is extended to the confluence with the Smith. The buried bedrock profiles of the other eight tributaries are not extended to the confluence.

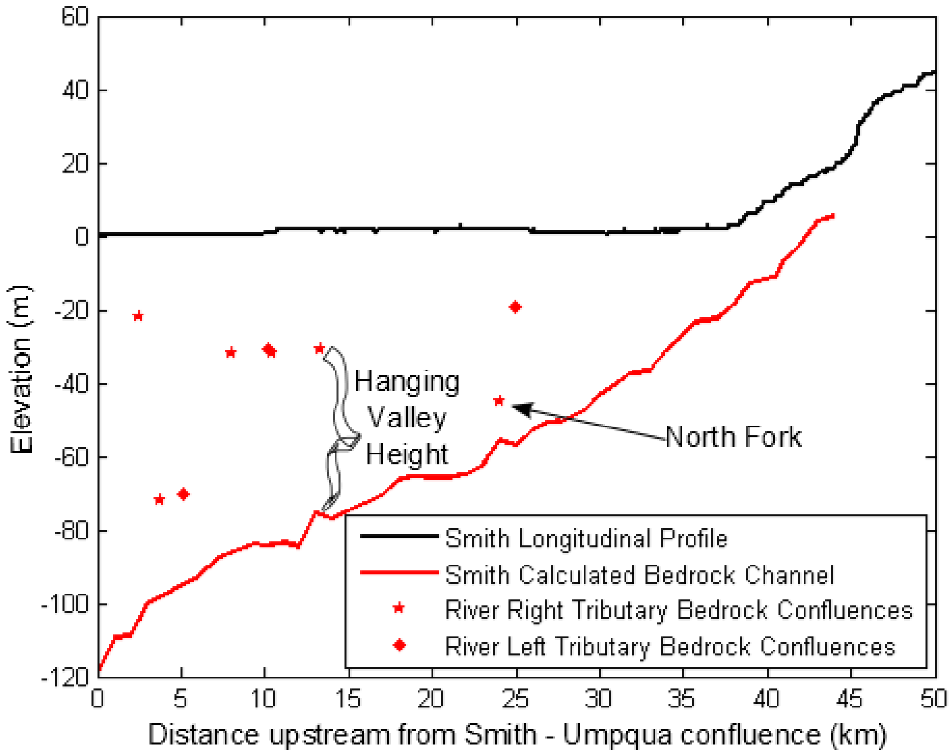

Figure 9: Smith River buried bedrock profile and projected tributary confluence bedrock elevations. The difference between the mainstem profile and the tributary confluence elevations represents a hanging valley that is presently buried in alluvium. All tributaries, except for North Fork plot well above the mainstem bedrock elevation.

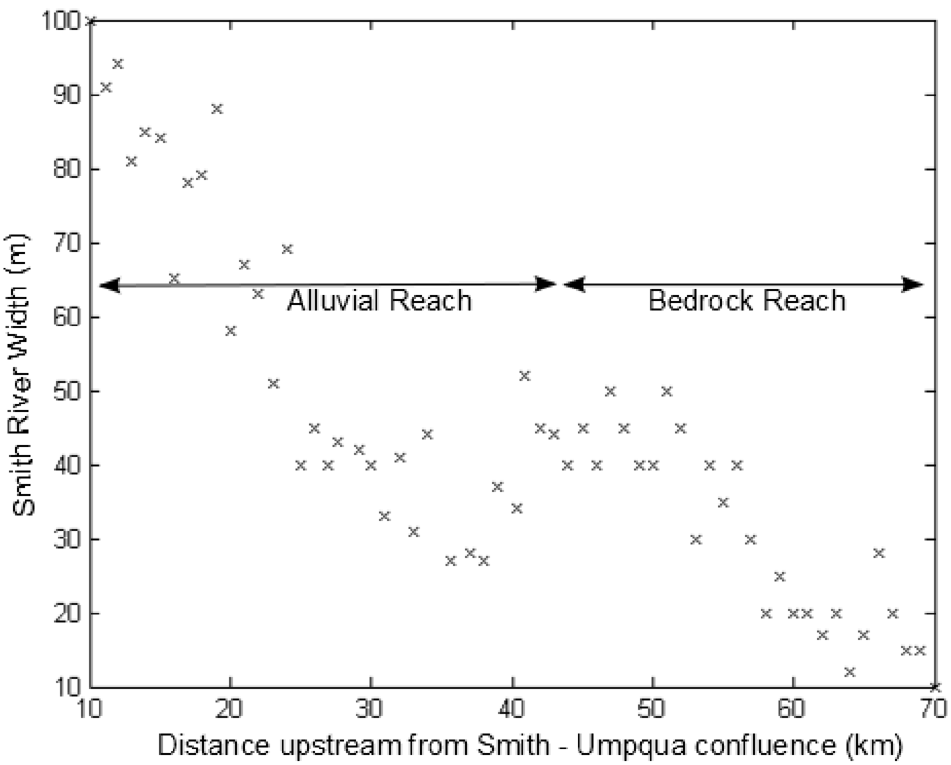

Figure 10: The plot above documents the river width along the Smith River across the transition from a bedrock to an alluvial reach. The river width generally increases downstream but there is no large jump in width across the transition.

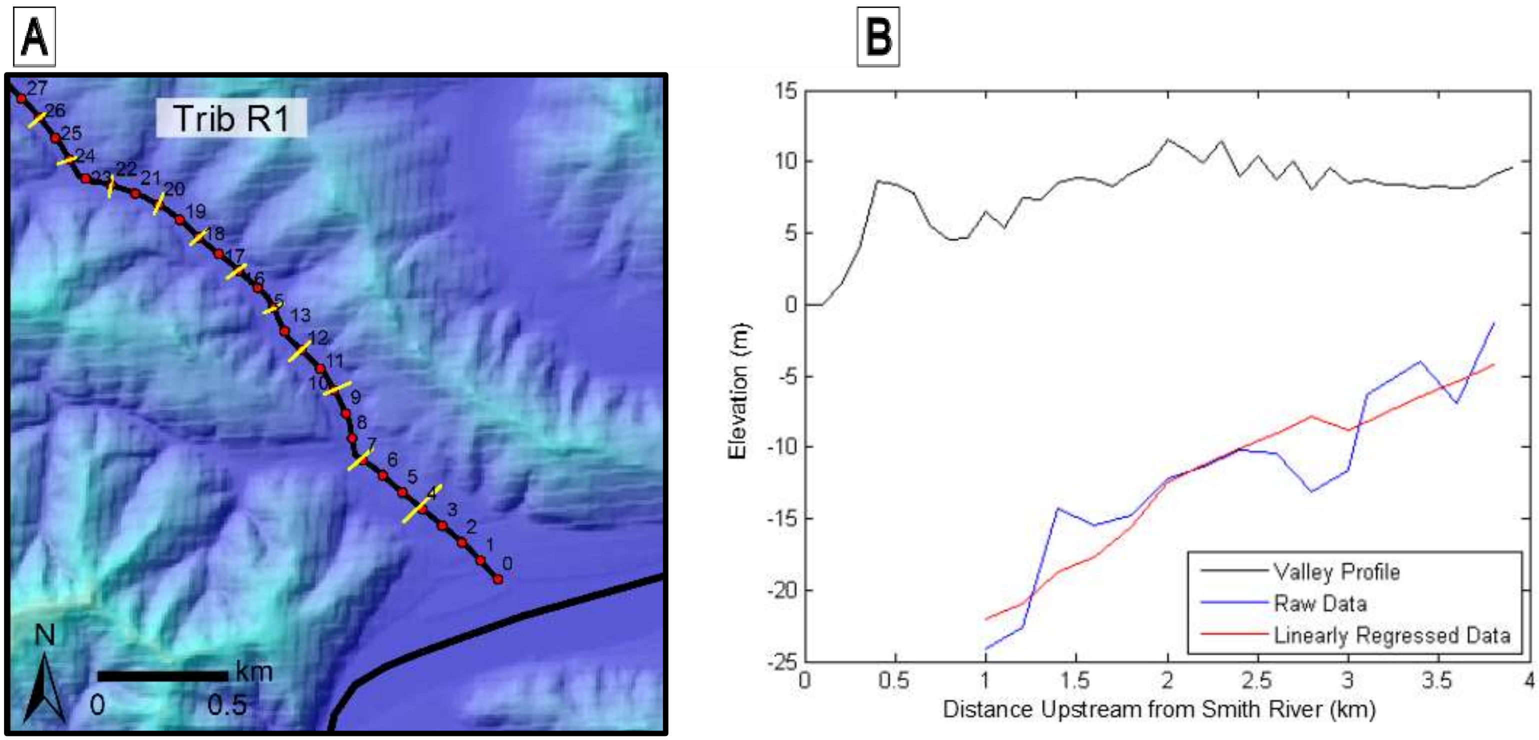

Figure 11: The base map in A shows the valley trace of Tributary R1 (black line) in increments of 100 m (red points) as well as the transects where valley width was measured (yellow lines). Note the widened valley in the lower most kilometer of the tributary (below point 10). Data collected in this kilometer is not representative of tributary processes and is not considered when calculating depth of the bedrock channel (B).

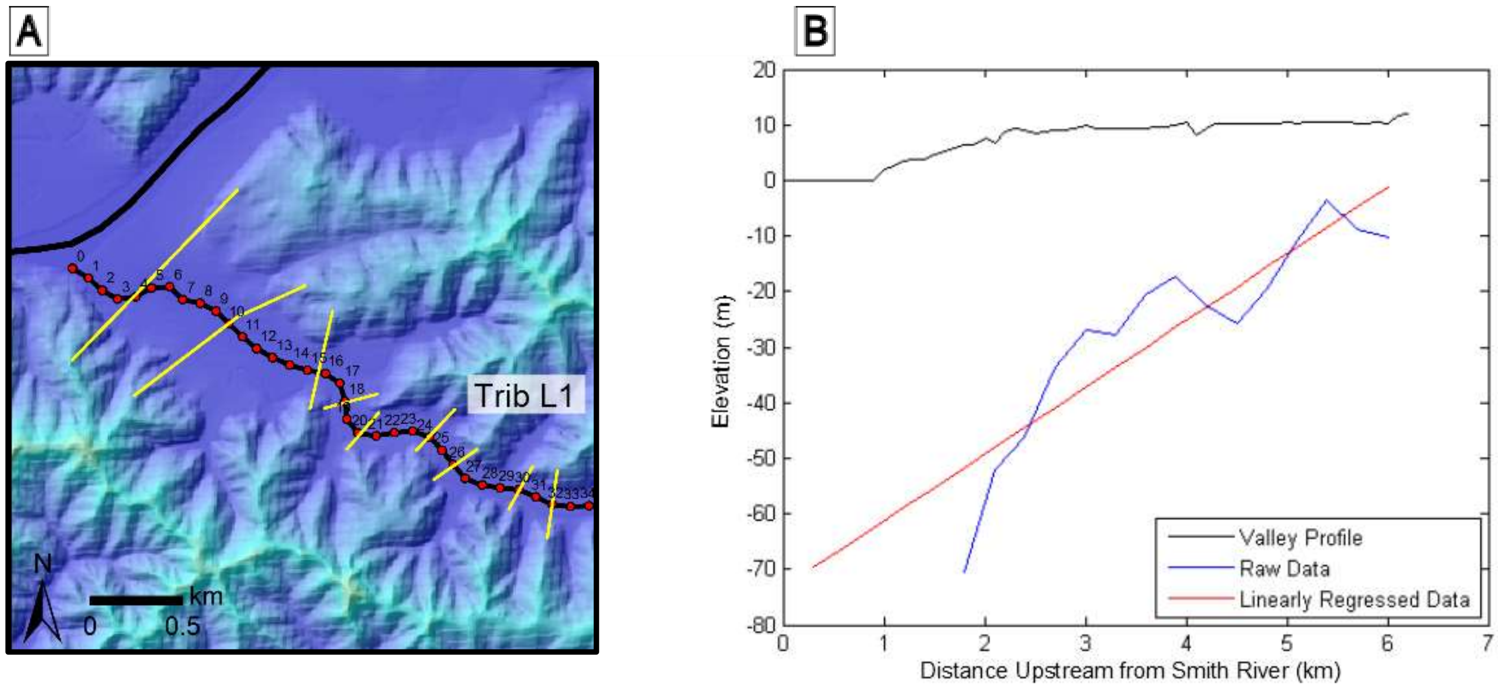

Figure 12: The base map (A) shows the valley trace of Tributary L1 (black line) divided into 100 m segments (red points) as well as transects where valley width was measured (yellow lines). The lower 1.8 km (below point 18) of the valley is significantly wider than upstream due to the Smith River, and not tributary processes. Therefore transects in this downstream reach are not considered when calculating the buried bedrock profile (B).

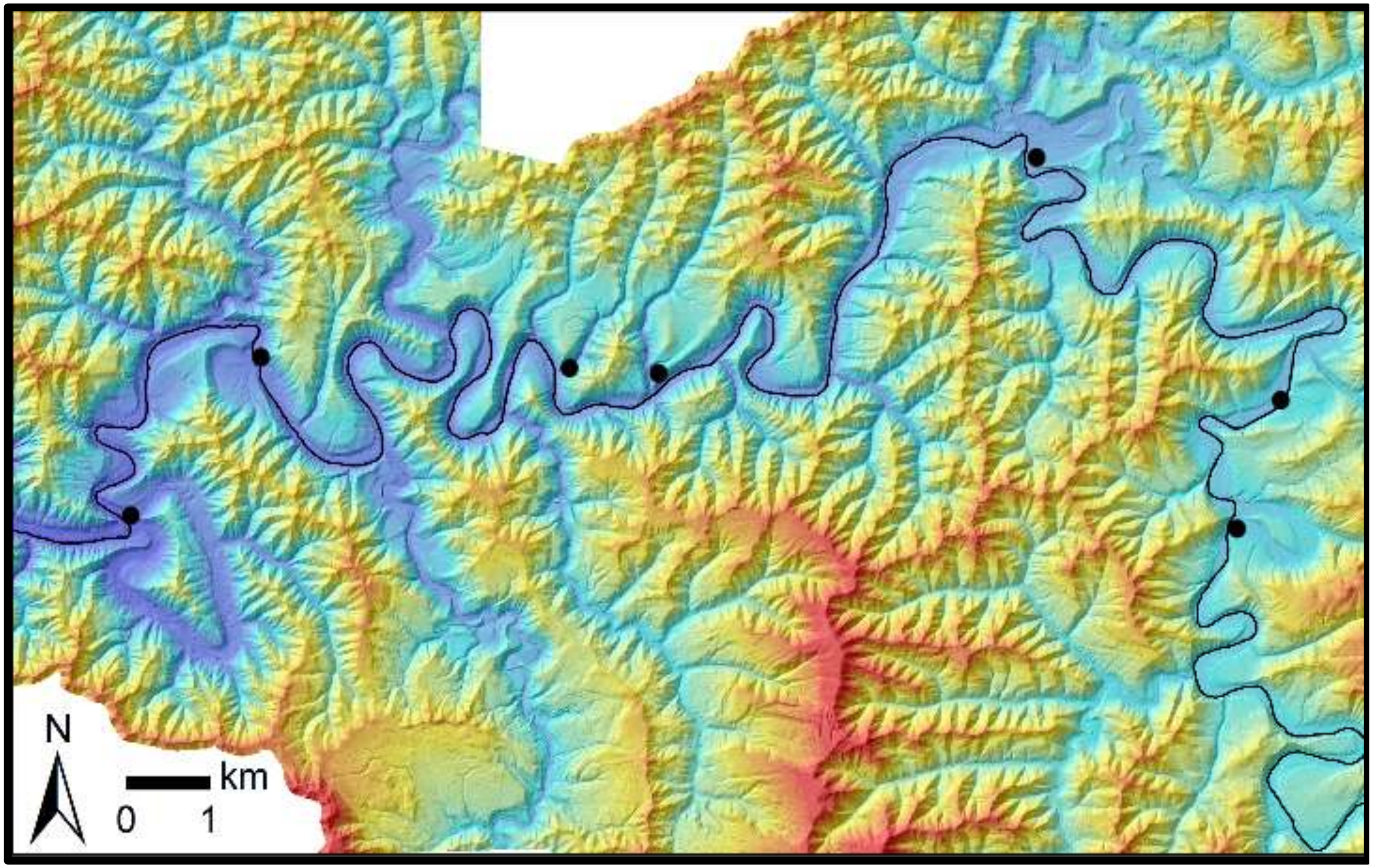

Figure 13: Basemap of the bedrock reach of the Smith River. The black line represents the thalweg of the Smith and the black circles note the upstream edge of meander bend cutoffs.

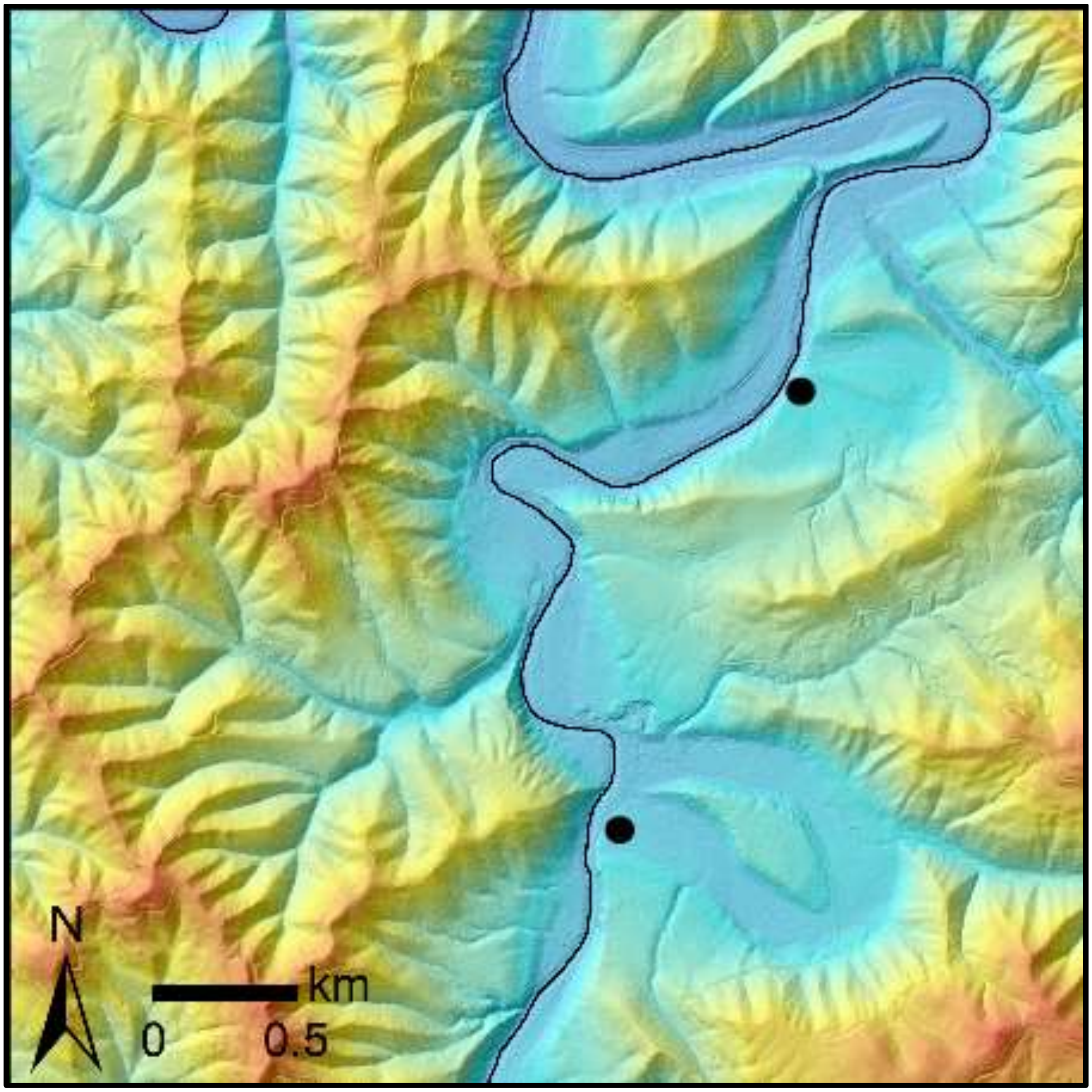

Figure 14: Detail of the bedrock reach of the Smith River from Figure 13. The two black points represent the two most upstream meander cutoff bends in Figure 13.

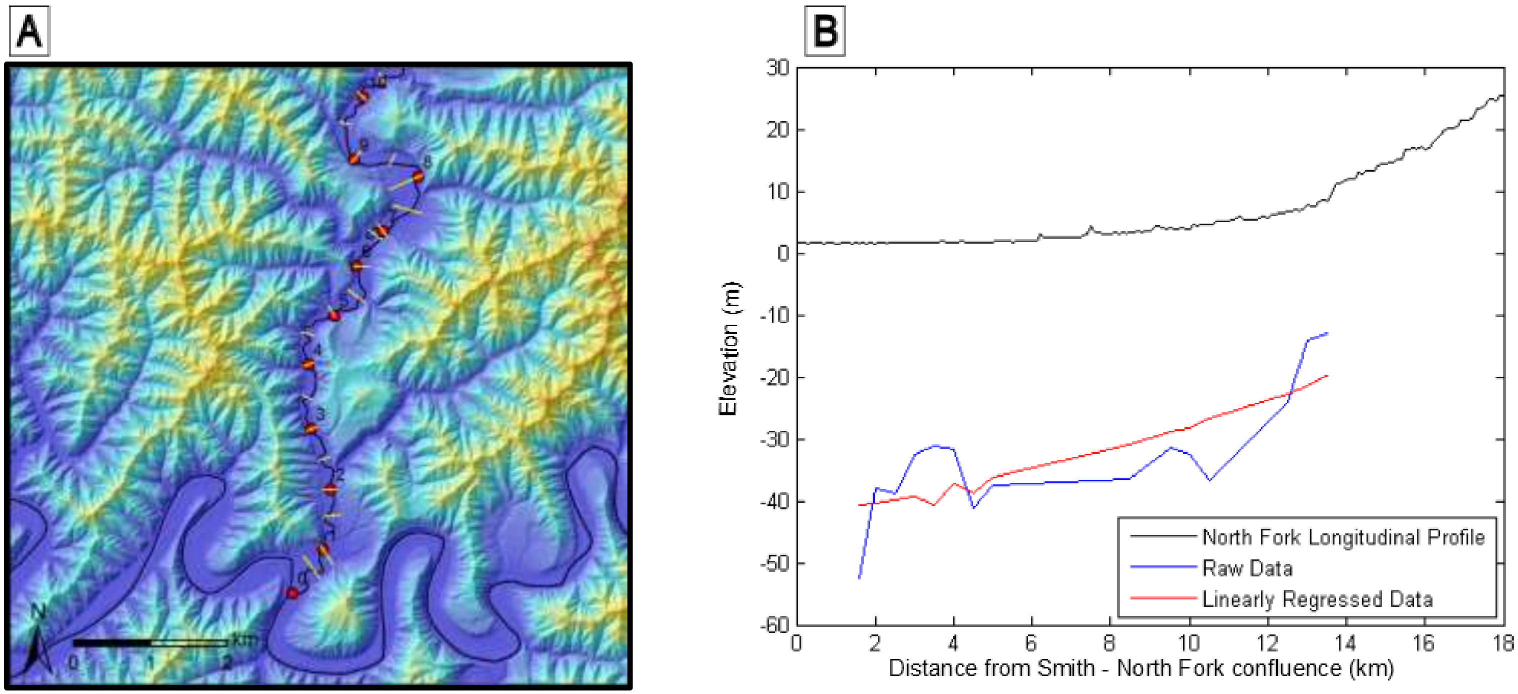

Figure 15: The basemap (A) shows the thalweg of the North Fork in 1 km sections (red points). Due to valley widening by the Smith River, transects (yellow lines) in the lower 1.6 km of the North Fork are excluded from the buried bedrock profile (B). Additionally, transects between river km 5-8 are also excluded to due to several meander bend cutoffs on the North Fork.

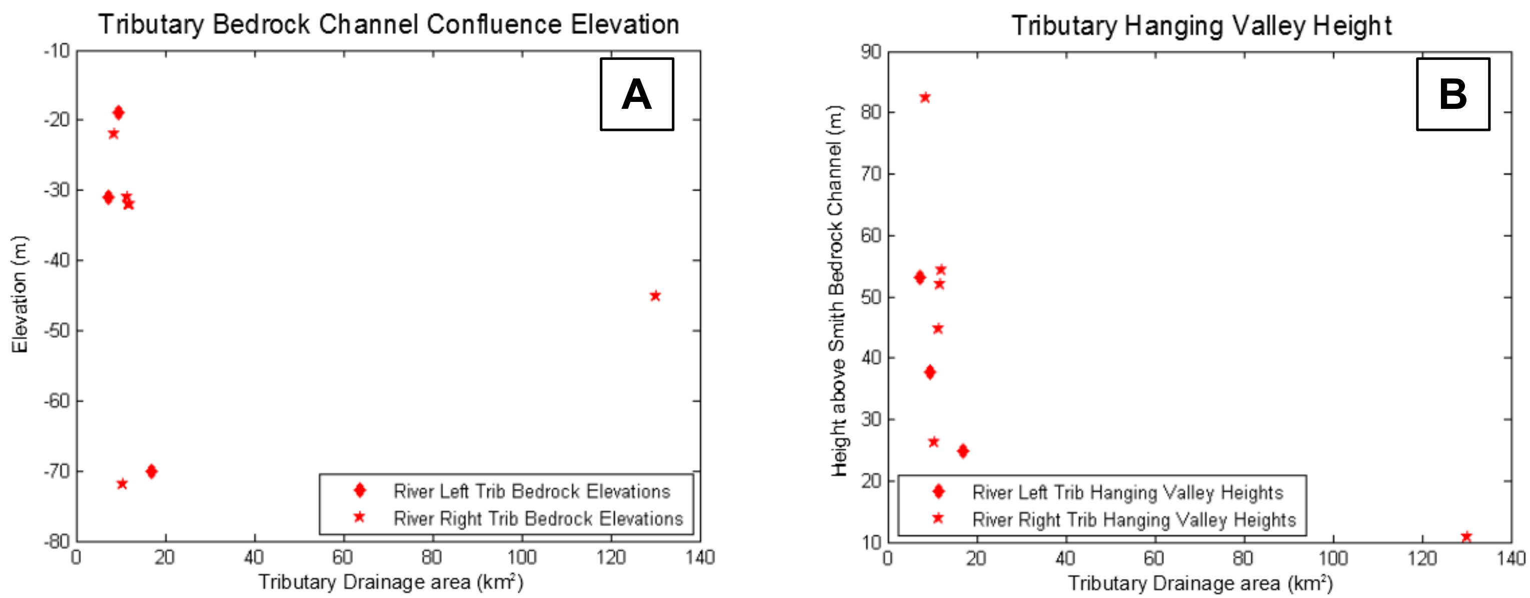

Figure 16: Both tributary bedrock confluence elevation (A) and hanging valley height (B) vary greatly for a given drainage area. Note the North Fork has a drainage area an order of magnitude greater than the other tributaries.

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