



## RESEARCH ARTICLE

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## Key Points:

- Bedrock channels are narrower and deeper than alluvial channels
- Bedrock channels with restricted sediment flux are anomalously wide and shallow
- Variation in sediment flux may prevent width-slope scaling in bedrock channels

## Supporting Information:

- Figures S1–S3

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## Substrate, sediment, and slope controls on bedrock channel geometry in postglacial streams

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**Abstract** The geometry of channels controls the erosion rate of rivers and the evolution of topography following environmental change. We examine how sediment, slope, and substrate interact to constrain the development of channels following deglaciation and test whether theoretical relationships derived from streams reacting to tectonic uplift apply in these settings. Using an extensive data set of channel geometry measurements from postglacial streams in the Scottish Highlands, we find that a power law width-drainage area scaling model accounts for 81% of the spatial variation in channel width. Substrate influences channel form at the reach scale, with bedrock channels found to be narrower and deeper than alluvial channels. Bedrock channel width does not covary with slope, which may be due to downstream variations in sediment flux. Bedrock channel width-to-depth ratios increase with discharge (or area) and sediment flux, consistent with increasing bed cover promoting lateral widening. We find steep, wide, and shallow bedrock channels immediately below lakes, which we interpret as the result of limited erosion due to a lack of sediment “tools.” Where sediment supply is sufficient to exceed transport capacity, alluvial channels develop wider, shallower geometries constrained primarily by flow hydraulics. Our results indicate that simple scaling models of channel width with drainage area are applicable at regional scale, but locally, channel width varies with substrate, and in the case of bedrock channels, with sediment flux.

## 1. Introduction

Bedrock channel geometry controls fluvial incision rates and the mechanisms by which hillslopes are affected by channel bed lowering, thereby influencing landscape evolution in response to environmental change [e.g., Hartshorn *et al.*, 2002; Whipple, 2004]. In many landscape evolution models, variations in channel geometry are approximated by scaling with discharge or drainage area [e.g., Howard *et al.*, 1994; Whipple and Tucker, 1999]. However, recent studies have shown that substrate [Allen *et al.*, 2013; Turowski *et al.*, 2007], slope [Amos and Burbank, 2007; Finnegan *et al.*, 2005; Whittaker *et al.*, 2007], and sediment flux [Finnegan *et al.*, 2007; Yanites and Tucker, 2010] may also influence the channel cross section, and debate continues over the factors that control bedrock channel form.

Previous studies of channel morphology have been largely focused on active orogenic terrains, in which channels are adjusting to variations in uplift rate [Amos and Burbank, 2007; Duvall *et al.*, 2004; Whittaker *et al.*, 2007; Yanites *et al.*, 2010]. The rapid uplift condition provides an important, and often implicit, component of proposed mechanisms for explaining channel geometry evolution [e.g., Turowski *et al.*, 2007; Wobus *et al.*, 2006a; Yanites and Tucker, 2010]. However, high uplift rates only occur across a small proportion of the global land area. Furthermore, over 20% of the current land area has been repeatedly glaciated during the Quaternary [cf., Ehlers and Gibbard, 2007], and postglacial adjustment is therefore an important aspect of landscape evolution [Ballantyne, 2002]. In areas deglaciated at the close of the last Quaternary cold stage (between ~20 and 10 ka), river channels have eroded rapidly into bedrock and glacially derived sediment at rates rivaling those of active orogenic mountain ranges, even in postorogenic terrains [Ballantyne, 2008; Meigs *et al.*, 2006; Jansen *et al.*, 2011; Tomkin, 2009; Valla *et al.*, 2010]. Although the evolution of channel geometry is intrinsically linked to the rate of channel erosion and the evolution of deglaciated terrains, channel geometry and the factors that control it have yet to be systematically assessed in postglacial streams.

We explore the controls on channel geometry in the postorogenic, postglacial Scottish Highlands. This setting provides an opportunity to assess the importance of substrate, slope, and sediment in controlling

channel morphology at reach scales under conditions of slow tectonic uplift. We measure channel width and depth in the field in 139 reaches along the 77 km of main stem channel in three upland catchments. With these data, we compare the geometries of bedrock and alluvial channels and derive empirical relationships for channel width and depth that demonstrate the relative importance of catchment area (a proxy for discharge), slope, and modeled estimates of sediment flux.

### 1.1. Background

Bedrock river channels form where the long-term capacity of the flow to transport sediment ( $Q_c$ ) exceeds the sediment available within the channel, i.e., the sediment flux ( $Q_s$ ), allowing only thin and discontinuous alluvial sediment cover to develop [e.g., Howard *et al.*, 1994; Whipple, 2004]. The energy available for erosion or sediment transport in an open channel can be formulated in terms of the total stream power ( $\omega$  in watts/m), unit stream power ( $\omega$  in watts/m<sup>2</sup>), or shear stress ( $\tau$  in N/m<sup>2</sup>):

$$\Omega = \gamma QS, \quad (1)$$

$$\omega = \gamma QS/W, \quad (2)$$

$$\tau = \gamma R_h S, \quad (3)$$

where  $Q$  is the discharge,  $S$  is the channel slope,  $W$  is the width,  $\gamma$  is the specific weight of water (9807 N/m<sup>3</sup>), and  $R_h$  is the hydraulic radius, given as  $R_h = WD/(2D + W)$  for a rectangular channel [Bagnold, 1966; Knighton, 1998]. A generalized bed load sediment transport equation has been defined from the results of numerous empirical studies [e.g., Huang and Nanson, 2002]:

$$Q_c \sim (\tau - \tau_{cr})^{1.5}, \quad (4)$$

where  $\tau_{cr}$  is a critical shear stress required to entrain bed load sediment of a given grain size. Scaling factors that relate discharge to catchment area and width to discharge (or area) are widely used to estimate shear stress and stream power in natural streams as a function of drainage area (which can easily be estimated from topographic data):

$$Q = k_q A^c, \quad (5)$$

$$W = k_w Q^b, \quad \text{and} \quad (6a)$$

$$W = k_w' A^{b'} \quad (\text{with } k_w' = k_w k_q^b), \quad (6b)$$

where  $k_q$ ,  $k_w$ , and  $k_w'$  are empirically determined dimensional constants and  $b$ ,  $b'$ , and  $c$  are dimensionless exponents [Howard *et al.*, 1994; Montgomery and Gran, 2001; Whipple and Tucker, 1999]. A value of  $b = 0.5$ , derived from studies of alluvial channels [e.g., Leopold and Maddock, 1953], is commonly applied and supported by field studies that report no consistent difference in the scaling of bedrock and alluvial channel width with discharge [Montgomery and Gran, 2001; Wohl and David, 2008].

Although differences in  $W$ - $A$  scaling for bedrock and alluvial channels have yet to be documented, studies in bedrock channels have shown that variations in erosional resistance influence channel geometry at reach [Jansen, 2006; Montgomery and Gran, 2001] and catchment scales [Allen *et al.*, 2013; Jansen *et al.*, 2010]. Resistant substrates are thought to give rise to narrower and deeper channels relative to those in weaker substrates because of the greater erosive force required for rock detachment [e.g., Montgomery and Gran, 2001; Sklar and Dietrich, 2001; Wohl and David, 2008]. Given that rock resistance influences the geometry of bedrock channels, the premise that alluvial and bedrock channels have similar  $W$ - $A$  scaling despite the contrasting erodibility of their boundaries is unexpected and merits further investigation.

Studies of bedrock channels experiencing spatial or temporal variations in uplift or erosion rate, i.e., where channel slope is not simply related to catchment area, have shown that the channel cross section may be influenced by channel slope [Finnegan *et al.*, 2005; Whittaker *et al.*, 2007]. Revised scaling relations accounting for the effects of both slope and discharge have been proposed, in which  $W \sim Q^{5/8} S^{-3/16}$  [Finnegan *et al.*, 2005] or  $W \sim Q^{0.38} S^{-0.44}$  [Whittaker *et al.*, 2007]. Covariance of channel width and slope, as well as a constant  $W/D$  ratio, is commonly assumed or predicted for detachment-limited channels, where erosion is considered to be wholly dependent on the capacity of the flow to detach rock [e.g., Attal *et al.*, 2008; Finnegan *et al.*, 2005; Wobus *et al.*, 2008]. Under these conditions, steeper slopes may be associated with faster flow and a reduction in the channel cross-sectional area [Finnegan *et al.*, 2005].

Recent theoretical studies have suggested that where sediment is introduced into bedrock channels, the dependence of width on slope will tend to increase, and the  $W/D$  ratio may scale weakly with catchment area and slope [Turowski *et al.*, 2007; Yanites and Tucker, 2010]. Such predictions are based on the propositions that channels widen with increasing bed cover, due to a relative increase in lateral bank erosion, and that bed cover tends to increase downstream, i.e., with increasing  $A$  and decreasing  $S$ . The link between bed cover, or sediment flux, and channel geometry has been confirmed in experimental studies [Finnegan *et al.*, 2007; Johnson and Whipple, 2010] and has been postulated as the cause of regional variations in  $W$ - $S$  scaling [Yanites and Tucker, 2010]. However, analysis of the relationships between slope, sediment, and the channel cross section is hindered by a lack of sediment flux data for natural streams: a problem that is further complicated in tectonically active terrains by potential covariation with rate of uplift, base-level fall, or erosion.

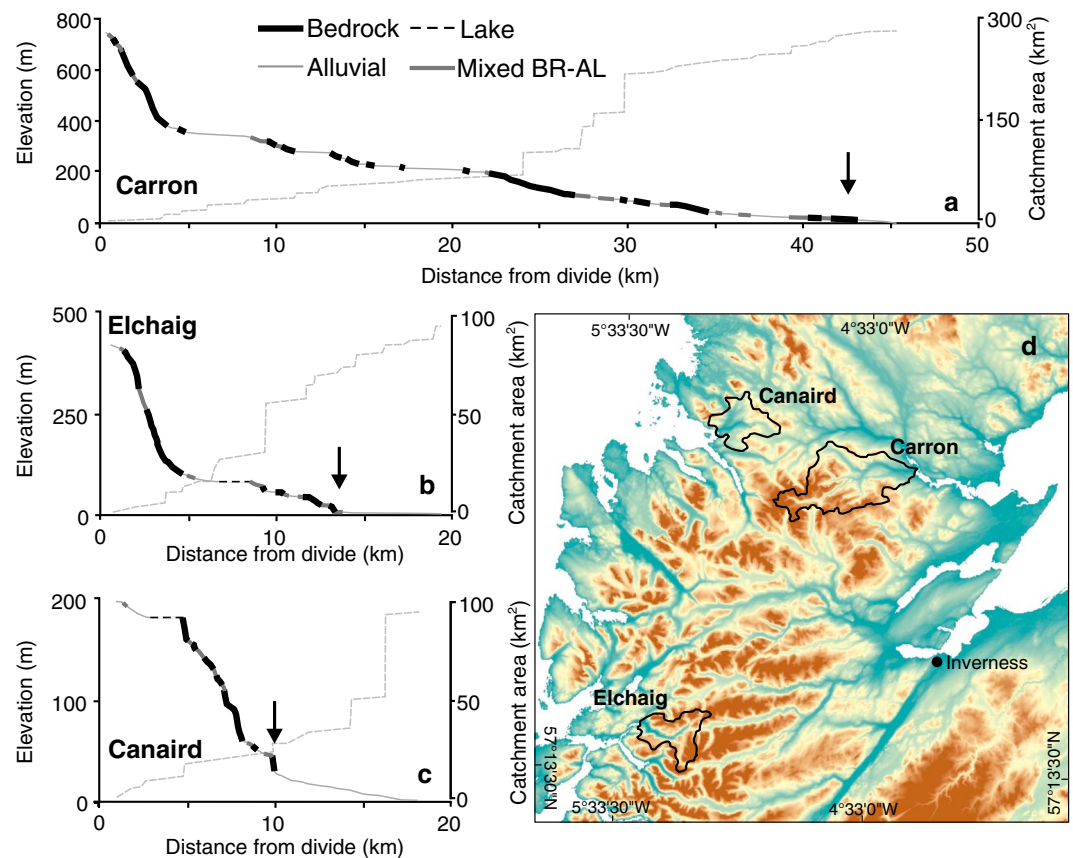
In areas of rapid uplift, channels are commonly considered to develop “optimum geometry” to carry the available water and sediment and maintain erosion around their boundaries at a rate that matches the uplift rate [Turowski *et al.*, 2007; Wobus *et al.*, 2006a; Yanites and Tucker, 2010]. This was demonstrated using analytical models in which both slope and width are considered to be freely adjustable, with channel slope renewed by tectonic uplift. However, in reality, channel width and slope take different lengths of time to adjust following perturbation because considerably more material needs to be eroded to affect channel slope [Knighton, 1998; Yanites *et al.*, 2010]. Furthermore, some observations of slope and width response to tectonic forcing suggest that narrowing alone may accommodate variations in uplift rate [Amos and Burbank, 2007; Lavé and Avouac, 2001; Yanites *et al.*, 2010] and/or that slope adjustment may be constrained by a threshold  $W/D$  [e.g., Allen *et al.*, 2013; Yanites *et al.*, 2010].

To investigate the relationships between sediment, slope, and width, we consider two aspects of natural streams. The first is the potential for downstream variations in sediment flux (and therefore bed cover) to influence channel geometry. Downstream variation in sediment flux may arise due to the distribution of sediment sources and the presence of sediment traps, such as lakes, along the channel network. The development of  $W$ - $S$  scaling may be restricted if local variations in bedcover influence channel geometry. The second aspect is the potential for channel slope to act as an independent control on channel width. The prolonged time scale for channel slope adjustment, particularly in tectonically quiescent terrains in which the capacity for steepening is limited, means that in many settings, “slope can be regarded as an imposed variable, or with valley slope imposed, as a variable adjustable over a limited range” [Knighton, 1998, p. 260]. Channels in which slope may be an independent control on width are important testing grounds for exploring the relationships between width, slope, and sediment. In such channels, steepening via uplift, the premise underpinning current theories of covariant channel dynamics, may be orders of magnitude slower than rates of channel erosion.

## 1.2. Hypothesis Testing for Postglacial Channels

The Scottish Highlands have been sculpted by glaciers during multiple phases of ice sheet and valley glaciation during the Quaternary, culminating with the British Irish Ice Sheet during the Late Devensian Glacial Maximum (26–21 ka) [Ballantyne, 2010]. A protracted deglaciation occurred from ~16 ka until the onset of the Holocene [Ballantyne, 2010]. The glacial legacy can be seen in the longitudinal profiles of Highland streams, which are commonly characterized by hyperconcave headwaters in glacial cirque basins and trunk valley segments containing numerous steep “steps” separated by valley “flats” or overdeepened rock-bound basins, in which lakes have formed (Figure 1) [e.g., Cotton, 1941; Jansen *et al.*, 2010]. The Highlands are underlain by resistant metasedimentary and igneous rocks, and their upland relief was established during a phase of tectonic uplift associated with continental rifting and passage of the Iceland plume in the early Cenozoic (65–50 Ma) [cf., Persano *et al.*, 2007; Hall and Bishop, 2002]. However, average Cenozoic denudation rates (since ~65 Ma) are only 0.02–0.03 m/kyr, implying that uplift rates are low, consistent with other postorogenic terrains [Persano *et al.*, 2007]. The glacial legacy and postorogenic setting of the Scottish Highlands make this region ideal for (1) comparing the geometry of bedrock and alluvial channels, (2) assessing whether slope acts as an independent control on bedrock channel width, and (3) examining the influence of spatially variable sediment flux on bedrock channel geometry.

The downstream variations in channel slope in Highland streams give rise to multiple downstream alternations between exposed bedrock (where  $Q_c > Q_s$ ) and alluvial channels (where  $Q_c \leq Q_s$ ), the latter



**Figure 1.** Longitudinal profiles for (a) the River Carron (maximum catchment area 300 km<sup>2</sup>), (b) River Elchaig (maximum catchment area 95 km<sup>2</sup>), and (c) River Canaird (maximum catchment area 95 km<sup>2</sup>), showing the distribution of reach types (bedrock, alluvial, and mixed bedrock-alluvial (BR-AL)) and the downstream increase in catchment area (dashed light grey line). (d) The location of the catchments in northern Scotland, with elevation from the 50 m grid NextMap DTM (© Intermap Technologies): high ground in brown (maximum elevation ~1400 m) and low ground in blue (minimum elevation = sea level). The arrows highlight the knickpoints that mark the limit of the glacio-isostatic uplift response; all reaches above this limit were surveyed. Note the different vertical scales in Figures 1a–1c.

being developed in unconsolidated postglacial alluvial or glacial/glaciofluvial sediment (Figure 1). The inherited glacially eroded valley floors mean that channel slope is unrelated to uplift and only weakly related to lithology [Jansen *et al.*, 2010; K. Whitbread, Postglacial evolution of bedrock rivers in postorogenic terrains: the NW Scottish Highlands, University of Glasgow, unpublished PhD Thesis, 2012]. Furthermore, generally low erosion rates mean that the response of the channels to base-level fall caused by glacio-isostatic rebound has affected the lower 10% of catchments only [cf., Bishop *et al.*, 2005; Jansen *et al.*, 2010; K. Whitbread, unpublished PhD Thesis, 2012]. The persistence of glacially conditioned knickpoints in the fluvial profiles indicates that postglacial erosion has been insufficient to recondition channel slopes since deglaciation [cf., Brardinoni and Hassan, 2007; Cotton, 1941]. At reach scale, initial channel slope may be considered as an “imposed variable” with limited slope adjustment having occurred during postglacial channel development. The irregular postglacial valley floors are also likely to influence the downstream flow of sediment in Highland catchments, particularly through the presence of overdeepened rock-bound basins/lakes, which frequently occur along main stem channels (cf., Figure 1). These lakes act as sediment traps and impose discontinuities in the channel sediment connectivity. Direct measurements of sediment flux at reach scales in natural streams are difficult to obtain over time scales relevant to fluvial incision (10<sup>2</sup>–10<sup>5</sup> years), but the presence of lakes in the Highland streams offers the opportunity to derive modeled estimates of sediment flux that account for the effect of sediment trapping. These streams are well suited for investigating the hypotheses that bedrock channels are consistently narrower than alluvial channels for a given discharge or area (hypothesis 1) and/or that the rate of change

in width with discharge or area may be lower for bedrock channels (hypothesis 2) [cf., *Montgomery and Gran*, 2001; *Wohl and David*, 2008]. We also test the hypotheses that steeper channels will be narrower than channels at lower slopes even in the absence of sediment (hypothesis 3) [cf., *Finnegan et al.*, 2005; *Wobus et al.*, 2006a] and that bedrock channels in areas with higher sediment supply will be wider and shallower than those with restricted sediment supply (downstream of lakes), due to increased bed cover promoting lateral widening during postglacial erosion (hypothesis 4) [cf., *Finnegan et al.*, 2007; *Turowski et al.*, 2007; *Yanites and Tucker*, 2010].

## 2. The Scottish Highlands Study Area

Glaciation of the Highlands has produced characteristic “U”-shaped glacial valleys with wide floors, steep sides, and irregular valley-floor profiles due to down-valley variations in the rate of glacial erosion [cf., *Brocklehurst and Whipple*, 2007; *MacGregor et al.*, 2000]. In these valleys, bedrock is mantled by a widespread but relatively thin cover of glacial and glaciofluvial deposits. Inner bedrock gorges, formed either before glaciation (during a preceding interglacial) and preserved under ice [*Montgomery and Korup*, 2011], excavated by subglacial meltwater [*Holtedahl*, 1967; *Jansen et al.*, 2014; *Tricart*, 1970], or formed by postglacial river erosion [*McEwen et al.*, 2002] are another feature of postglacial terrains in the Scottish Highlands [*Benn and Evans*, 1998; *Werritty and McEwen*, 1997; K. Whitbread, unpublished PhD Thesis, 2012].

Sediment supply rates to Highland streams are generally low due to the resistant metamorphic bedrock, the stabilizing effect of peat cover and vegetation on hillslopes, and the disconnection between hillslopes and channels in the wide U-shaped valleys [e.g., *Ballantyne*, 2008]. The low sediment supply means that bedrock reaches occur in steep channel segments (where  $Q_c > Q_s$ ) and in some lower slope segments of trunk channels where they are commonly associated with inner gorges (Figure 1). Debris flow activity is minimal due to the resistant rocks and thin sediment cover and affects only a few gullies and steep tributaries in the headwalls of glacial cirques and on steep valley sides. Owing to the wide valley floors, debris flow deposits rarely reach trunk channels.

Glacio-isostatic surface uplift following ice decay has caused relative base-level fall during the Holocene, with maximum uplift rates of ~10–30 m/kyr during deglaciation declining to ~0.5–2 m/kyr by the late Holocene [*Firth and Stewart*, 2000; *Shennan et al.*, 2000]. The duration of glacio-isostatic surface uplift has been too short to increase relief substantially [*Jansen et al.*, 2010], but the accompanying fall in relative sea level has led to steepening and erosion along many bedrock channel outlets at the coast [*Bishop et al.*, 2005; *Castillo et al.*, 2013; *Jansen et al.*, 2011]. Quantitative assessments of the retreat of these base-level response knickpoints indicate that channel systems remain only partially adjusted in their lowermost reaches [*Bishop et al.*, 2005; *Jansen et al.*, 2011]. The transient reaches formed in response to glacio-isostatic surface uplift are excluded from analysis in this study (Figure 1).

## 3. Methods

### 3.1. Field Data

Channel geometry was surveyed along a total of 77 km of trunk channel in three Highland catchments: the River Carron (300 km<sup>2</sup>), River Elchaig (97 km<sup>2</sup>), and River Canaird (95 km<sup>2</sup>; Figure 1). All the catchments are underlain by resistant metamorphosed sandstone (including quartzite) and mudstone with minor outcrops of intrusive basaltic and granitic rocks. The catchments were selected for their absence of major hydroelectric power schemes and the limited extent of disturbance associated with recent forestry plantations.

We surveyed 42 alluvial, 65 bedrock, and 32 mixed bedrock-alluvial reaches (Figure 2) ranging from 50 to 2000 m in length and located above the postglacial base-level fall knickpoints (Figure 1). A reach is defined as a channel segment exceeding five channel widths in length that displays a consistent morphology characterized by variation in the degree of bed cover [e.g., *Wohl and Merritt*, 2008]. In reaches with partial bedrock exposure, bed cover was surveyed at 10 m intervals by visually estimating the proportion of bedrock exposure in the banks and bed in a 2 m wide swath across the channel. An overall fraction of bed cover was derived as the average cover along the length of the reach. Reaches with bedrock banks and less than 30% bed cover were classified as bedrock reaches. Where sediment cover (including the channel





**Figure 2.** Field photographs of example reach types in the Scottish Highlands. (a) A confined bedrock reach in metasandstone in the River Carron (5% bed cover). (b) An unconfined bedrock reach in metasandstone in the River Elchaig (<5% bed cover). (c) A mixed bedrock-alluvial reach in the upper River Carron (70% bed cover). (d) An upland alluvial reach in the upper River Elchaig (100% bedcover). The blue arrows indicate the flow direction.

banks) exceeded 99%, reaches were classified as alluvial, and mixed bedrock-alluvial reaches were designated where bed cover fell between 99 and 30%. Reaches where one or both banks were formed in unconsolidated sediment were also classified as mixed bedrock-alluvial even if the proportion of bed cover was lower than 30%. In order to allow for greater discrimination of potential substrate effects on channel geometry, this study applied a 30% bed cover (i.e., 70% rock exposure) threshold for defining bedrock channels, somewhat lower than the 50% cover applied in previous studies [e.g., *Montgomery and Gran*, 2001; *Wohl and David*, 2008].

A total of 450 measurements of channel width and depth were carried out at 2–10 points along each reach, using a laser rangefinder (width  $\pm 10$  mm accuracy) and 1 cm graduated measuring pole (depth). Width and depth were measured at bankfull, defined by the break of slope between the channel and floodplain in alluvial reaches, and at the bankfull equivalent defined by debris wash lines, vegetation scour area, and bank or gorge morphology in bedrock reaches [*Montgomery and Gran*, 2001; *Wohl and David*, 2008]. Debris lines from a recent bankfull event on the River Carron, traceable through many bedrock and alluvial reaches, were used to “calibrate” the bankfull-equivalent stage in bedrock reaches. Bankfull flow relates approximately to the 1–2 year return interval flood [e.g., *Jansen et al.*, 2010; *Knighton*, 1998; *Wolman and Miller*, 1960]. The distribution and number of measurements per reach were selected to account proportionally for subreach-scale variations in channel morphology [*Montgomery and Buffington*, 1997; *Wohl and Merritt*, 2008]. In alluvial channels, measurements sites were selected to account for the frequency and relative length of pools and riffles. In bedrock reaches, which were commonly more variable in morphology, the relative lengths of pools, chutes, and cascades were accounted for in the spacing and increased frequency of measurements.

Sediment in both alluvial and bedrock channels consists mainly of subrounded gravel, cobbles, and boulders of metasandstone and quartzite with rare granite and basalt and subordinate sand. The alluvial and

**Table 1.** Lithology Data<sup>a</sup>

Lithology	Schmidt Hammer $R^b$	Average Fracture Spacing (m) <sup>c</sup>	No. of Bedrock Reaches <sup>d</sup>
Fault affected	$46 \pm 1.3$	0.1*	8
Metamudstone	$38 \pm 1.4$	$0.6 \pm 0.04$	6
Metasandstone	$53 \pm 1.0$	$0.2 \pm 0.01$	33
Jointed metasandstone	$47 \pm 0.7$	0.1*	6
Gneiss	$50 \pm 0.8$	$0.36 \pm 0.10$	9

<sup>a</sup>Lithological resistance data for the main rock types in the study channels.

<sup>b</sup>Schmidt hammer  $R$  values are either averages from surveys of 30 readings from a single site (italics) or averages of several  $R$  values derived from multiple sites.

<sup>c</sup>Average fracture spacing was derived from surveys of 25 to 50 fractures along two perpendicular transects at multiple sites; note that the asterisk denotes estimated values for sites unsuitable for transects.

<sup>d</sup>Three bedrock reaches in metasiltstone were not included in this analysis due to the small sample size.

glaciofluvial terraces that comprise the alluvial channel banks generally consist of deposits of similar size and composition to those found in the active channel. Bedrock reaches were classified according to the dominant rock type exposed in the reach based on lithology and the degree of fracturing. For each rock type, estimates of compressive strength and average fracture spacing were derived at type localities from Schmidt hammer and fracture surveys conducted along perpendicular transects in the channel bed (Table 1).

### 3.2. Topographic Analysis

Streamlines, elevation profiles, channel slope, and upstream catchment area were obtained from the 5 m grid and 1 m vertical resolution NextMap digital terrain model (DTM), derived from interferometric synthetic aperture radar [Intermap Technologies, 2007]. DTM processing was conducted in ArcGIS, and derived streamlines and elevation profiles were manually corrected for errors using Ordnance Survey 1: 10,000 scale topographic maps and contours. Streamline errors affected less than 10% of reaches and were predominantly located in flat valley-floor segments, where meander loops were missed by the DTM processing.

Elevations (m), upstream catchment area ( $\text{km}^2$ ), and along stream distance (km) were derived at 5 to 7.5 m intervals along the main stem channel lines. For each reach, an average upstream catchment area was calculated as the arithmetic mean of points within the reach; no reach contains a substantial tributary junction. Reach slope ( $S$ ) was calculated as the elevation fall through the reach divided by the along-channel reach length. This method of deriving  $S$  was preferred over moving-window averaging [Wobus *et al.*, 2006b] because the latter method tended to underestimate the slope of bedrock reaches which are commonly short and preceded and succeeded by longer, low-slope alluvial reaches.

### 3.3. Flow, Stream Power, and Sediment Flux

The gauging data available for the study catchments are insufficient to account for the attenuating effect of lakes or provide an independent estimate of long-term average discharge ( $Q$ ) for each reach. Catchment area,  $A$ , is used as a proxy for discharge in much of the following analyses [Montgomery and Gran, 2001].

Sediment gauging records are not available for the study catchments, nor are there long-term records of sediment flux. However, sediment supply to channels may be considered to scale as a power function of catchment area [cf., Massong and Montgomery, 2000; Montgomery *et al.*, 1996]. Power law-scaling relationships were used to derive modeled sediment flux estimates (termed sediment indices,  $I_s$ ) according to

$$I_s = k_{qs} A_c^e, \quad (7)$$

where  $k_{qs}$  is a constant arbitrarily taken as 100 to give the index values a similar range to shear stress,  $e$  is defined according to the basin wide or linear scaling supply models described below, and  $A_c$  is the contributing catchment area. Lakes in the Elchaig and Canaird catchments (Figure 1) are acting as sediment traps which are likely to significantly influence sediment flux and thus potentially affect channel geometry. To account for this trapping effect,  $A_c$  is reset to zero at lake outlets. Approximately 30% of reaches are located downstream from lakes (i.e.,  $A_c < A$ ).

In the first model ( $I_{s1}$ ), a value of  $e = 0.85$  is used to reflect a system in which sediment is supplied from hillslopes and catchment headwaters source more sediment to streams than downstream areas due to steep slopes and narrow valley floors [Massong and Montgomery, 2000; Montgomery *et al.*, 1996]. In the second model ( $I_{s2}$ ), a

**Table 2.** Channel Width Statistics<sup>a</sup>

Value	Individual Measured Width (m)	Reach-Averaged Width (m)
Mean	18.8	18.7
Standard error	0.51	0.89
Standard deviation	11.2	10.8
Variance	126.1	116.0
Skewness	0.87	0.73
Kurtosis	0.74	0.08
n	491	139

<sup>a</sup>Comparison of descriptive statistics derived for all individual measured widths and all reach-averaged widths.

value of  $e = 0.6$  is used to reflect a system in which sediment is sourced largely via erosion of sediments and bedrock in the channel banks and bed. In this case, sediment supply is considered to be proportional to stream length, which has been found in many channels to scale as  $\sim A^{0.6}$  [Hack, 1957; Ijjasz-Vasquez et al., 1993].

We use the variable  $I_s/\tau^{1.5}$  as a relative measure of the degree of oversupply or undersupply of sediment (i.e.,

variations in  $Q_s/Q_c$ ). Sediment transport is assumed to scale with shear stress (equation (5)), and the critical shear stress ( $\tau_{cr}$ ) is omitted on the basis that bankfull channel geometry is related to high flows with a 1–2 year return interval; during such flows, calculated values of  $\tau$  are typically greater than the  $\tau_{cr}$  required to mobilize sediment up to cobble size. Using this method, we find that the variability in  $I_s/\tau$  is strongly related to downstream variations in slope (i.e.,  $1/S$ ), with 10–16% of the variation arising from both the resetting of contributing area at lake outlets (affecting 30% of reaches) and changes in hydraulic radius (Figure S3 in the supporting information). Accounting for the effect of lakes on sediment supply means that a reach immediately below a lake will have a lower  $I_s/\tau^{1.5}$  than a reach of equivalent slope and hydraulic radius above the lake.

### 3.4. Data Analysis

Channel width ( $w$ ) and depths ( $d$ ) were averaged for each reach to remove the subreach-scale variations that are associated with pools and riffles in alluvial channels and with chutes and plunge pools in bedrock channels [Montgomery and Buffington, 1997; Montgomery and Gran, 2001]. A comparison of the individually measured widths with reach-averaged widths ( $W$ ) indicates that they have similar mean values and skewness, but reach-averaged widths have a lower standard deviation, indicating that averaging reduces the variability but preserves the pattern of the data (Table 2).

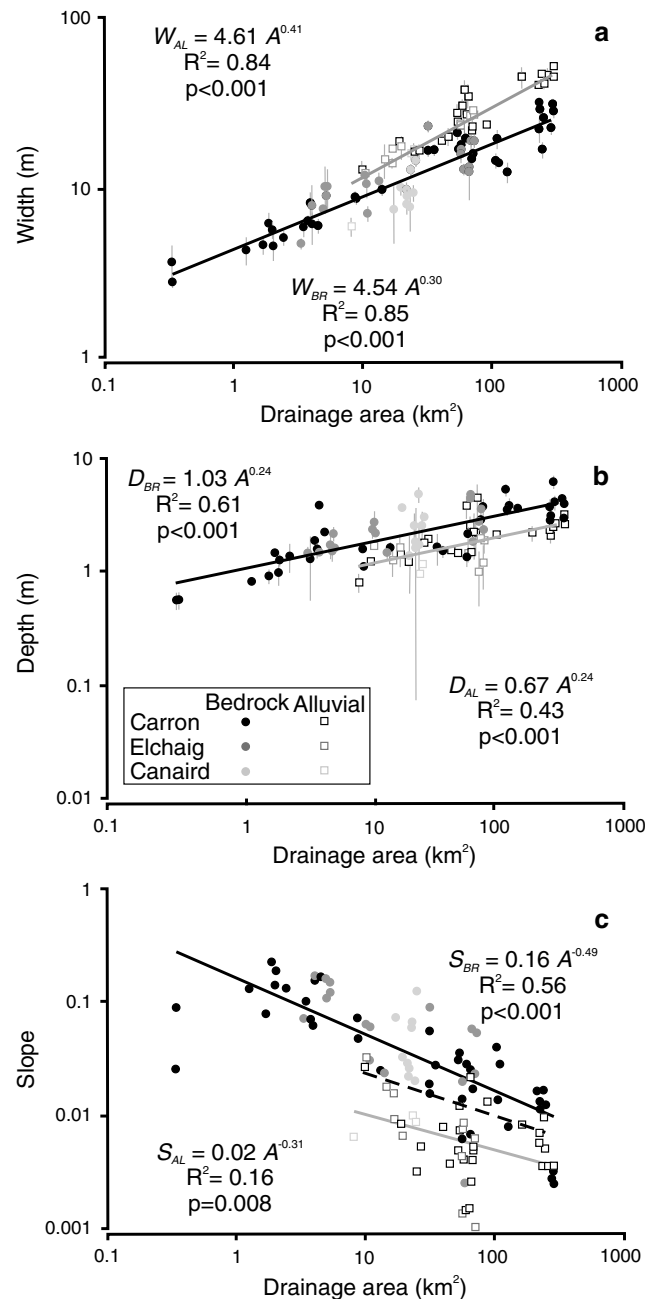
Uncertainty in  $w$  and  $d$  arises from measurement error and uncertainty in the reference discharge level. Due to high intrareach variability, the standard error of the reach-averaged width ( $W$ ), ranging from 1 to 25% (average 8%), is approximately an order of magnitude greater than the uncertainty on  $w$ .

**Table 3.** Regression Data for Channel Width and Depths<sup>a</sup>

	W-A			D-A		
	$k_w'$	$b'$	$R^2$	$k_d'$	$p'$	$R^2$
All data ( $n = 139$ )	$4.36 \pm 1.06$	$0.37 \pm 0.02$	0.81	$0.95 \pm 1.08$	$0.20 \pm 0.02$	0.42
Carron ( $n = 78$ )	$4.49 \pm 1.07$	$0.37 \pm 0.02$	0.86	$0.95 \pm 1.09$	$0.21 \pm 0.02$	0.61
Elchaig ( $n = 42$ )	$4.94 \pm 1.14$	$0.35 \pm 0.04$	0.68	-	-	-
Canaird ( $n = 19$ )	$1.30 \pm 2.01$	$0.71 \pm 0.23$	0.36	-	-	-
All bedrock ( $n = 65$ )	$4.57 \pm 1.06$	$0.30 \pm 0.02$	0.85	$1.03 \pm 1.09$	$0.24 \pm 0.02$	0.61
Carron ( $n = 36$ )	$4.46 \pm 1.06$	$0.31 \pm 0.02$	0.93	$0.96 \pm 1.10$	$0.24 \pm 0.03$	0.72
Elchaig ( $n = 20$ )	$5.37 \pm 1.17$	$0.28 \pm 0.05$	0.63	$1.22 \pm 1.21$	$0.19 \pm 0.06$	0.36
Canaird ( $n = 9$ )	-	-	-	-	-	-
All alluvial ( $n = 42$ )	$4.61 \pm 1.12$	$0.41 \pm 0.03$	0.84	$0.67 \pm 1.20$	$0.24 \pm 0.04$	0.43
Carron ( $n = 27$ )	$5.52 \pm 1.18$	$0.37 \pm 0.04$	0.81	$0.87 \pm 1.27$	$0.20 \pm 0.05$	0.36
Elchaig ( $n = 12$ )	$5.75 \pm 1.17$	$0.35 \pm 0.04$	0.87	-	-	-
Canaird ( $n = 3$ )	-	-	-	-	-	-
All mixed ( $n = 32$ )	$4.35 \pm 1.12$	$0.37 \pm 0.03$	0.86	$0.50 \pm 1.20$	$0.32 \pm 0.04$	0.67
Carron ( $n = 15$ )	$4.44 \pm 1.16$	$0.37 \pm 0.03$	0.93	$0.62 \pm 1.35$	$0.29 \pm 0.06$	0.65
Elchaig ( $n = 10$ )	$4.28 \pm 1.40$	$0.39 \pm 0.09$	0.71	-	-	-
Canaird ( $n = 7$ )	-	-	-	-	-	-

<sup>a</sup>Power law channel geometry scaling relations ( $W = k_w' A^{b'}$  and  $D = k_d' A^{p'}$ ) for combined data from the three study catchments and for individual catchments. Only correlations significant at  $P < 0.05$  are shown.





**Figure 3.** (a) Channel width, (b) depth, and (c) slope versus drainage area for bedrock and alluvial reaches from the River Carron, River Elchaig, and River Canaird (key in Figure 3b). The grey error bars in Figures 3a and 3b are the standard errors of the reach-averaged width and depth values. Plots are shown with logarithmic scales and power law regressions for all bedrock (black line) and all alluvial (grey line) reaches. The critical slope ( $S_c$ ) for the transition from alluvial to bedrock channels (Figure 3c, dashed black line), as determined by discriminant analysis, is  $S_c = 0.05 A^{-0.36}$  [cf., Massong and Montgomery, 2000]. Data for mixed bedrock-alluvial reaches are not shown for clarity but are intermediate between bedrock and alluvial channels (Table 3).

area (Figure 3 and Table 3). Comparison of bedrock and alluvial channel scaling relations for the Rivers Carron and Elchaig via analysis of covariance (ANCOVA) demonstrates that for each catchment, the

Uncertainty on  $d$  is greater than for  $w$  due to the measurement method used, and the standard error of the reach-averaged depth ( $D$ ) is only marginally greater than the uncertainty on individual  $d$  values, ranging from 1 to 40% (average 12%).

Channel width and slope tend to depend on catchment area ( $A$ ) [e.g., Montgomery and Gran, 2001; Wohl and David, 2008]. In order to assess the role of substrate, sediment, and slope in setting channel width, the effect of variations in  $A$  on width was accounted for through the calculation of reference channel widths ( $W_R$ ), analogous to the channel steepness index. These values were derived by dividing  $W$  by the predicted width obtained from the  $W$ - $A$  regression so that

$$W_R = \frac{W}{k_w' A^{b'}}, \quad (8)$$

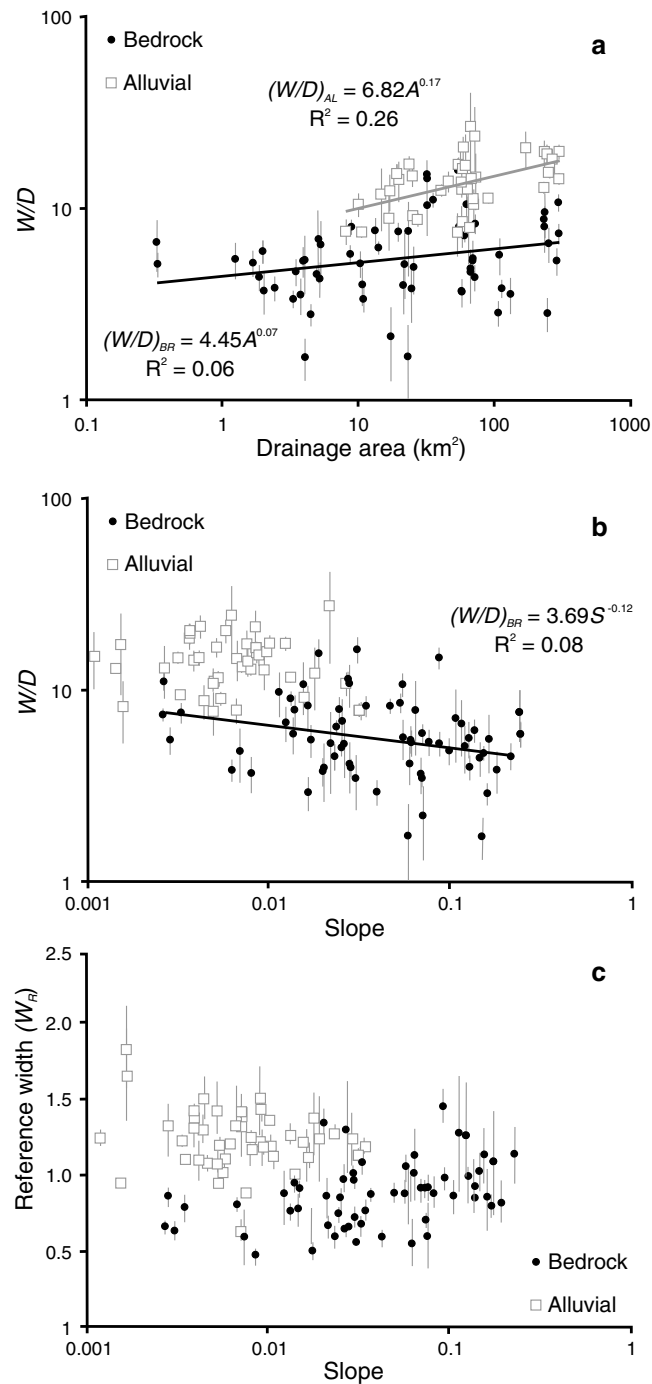
with values of  $k_w'$  and  $b'$  empirically derived for the relevant subsets of the data [e.g., Allen et al., 2013; Yanites and Tucker, 2010].

## 4. Results

### 4.1. Bedrock Versus Alluvial Channel Morphology

In both bedrock and alluvial settings, reach-averaged channel width ( $W$ ) is positively scaled with drainage area ( $A$ ), accounting for 36 to 86% of the variation in channel width in the study catchments (Table 3). Significant scaling of reach-averaged channel depth with  $A$  is only seen in the River Carron, reflecting greater spatial variability in channel depths than channel widths (Table 3). Analysis of the combined data from all three study catchments demonstrates that  $A$  accounts for 81% of the variation in  $W$  and 42% of the variation in  $D$  (Figure 3 and Table 3).

Grouping reaches by channel type indicate that bedrock channels are consistently narrower and deeper than alluvial channels for a given drainage



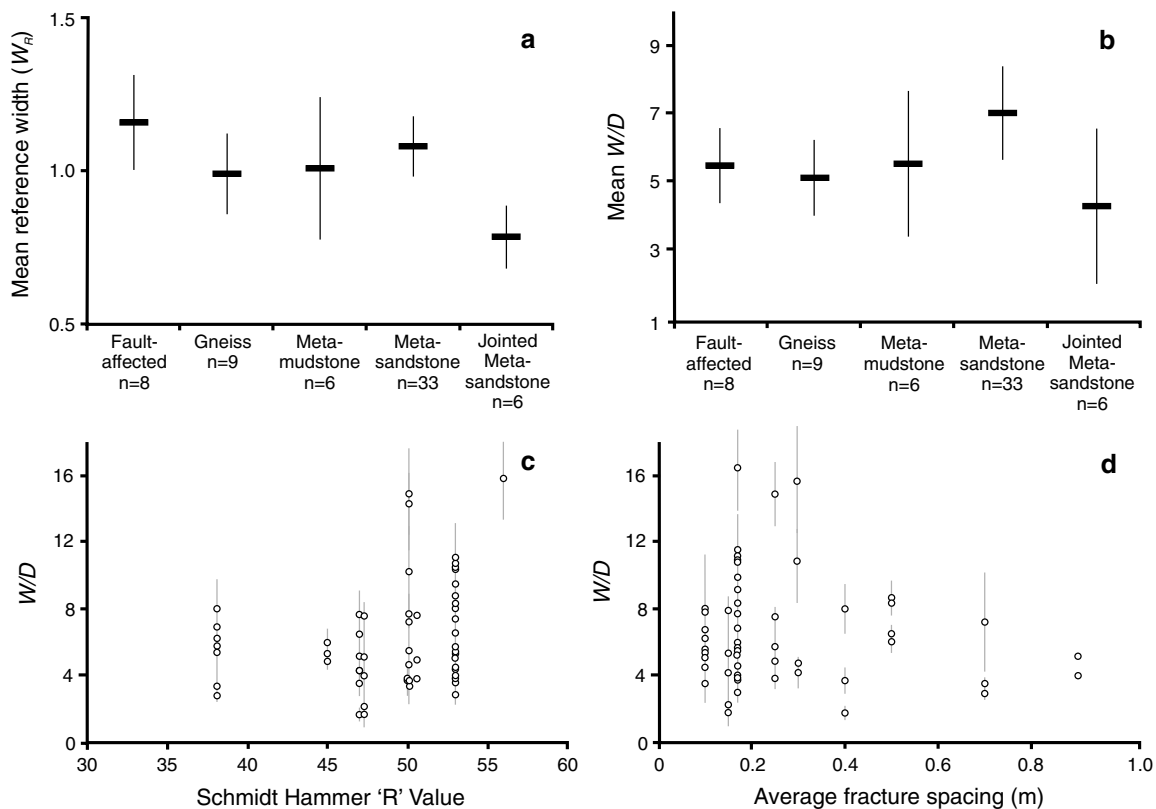
**Figure 4.** The width-to-depth ratio versus (a) drainage area and (b) slope for bedrock and alluvial reaches. Plots have logarithmic scales, and power law regressions are significant ( $P < 0.05$ ). The grey error bars show  $1\sigma$  uncertainty on  $W/D$  derived from the standard errors of the reach-averaged widths and depths. (c) Reference width ( $W_R$ ) versus slope;  $W_R$  was calculated from equation (8) using values of  $k_w' = 4.36$  and  $b' = 0.37$  from the  $W$ - $A$  regression for all data (Table 3).

channel  $W/D$ - $A$  scaling exponent ( $0.17 \pm 0.05$ ) is consistent with the difference between  $W$ - $A$  and  $D$ - $A$  scaling exponents ( $b'$  and  $p'$ , where  $D = k_d' A^{p'}$ ) for alluvial channels ( $0.41$  and  $0.24$ , respectively; Table 3). The  $W/D$ - $A$  scaling exponent ( $0.07 \pm 0.03$ ) for bedrock channels is also similar to the difference between  $b'$

intercept ( $k_w'$ ) for bedrock channel  $W$ - $A$  scaling is significantly lower than for alluvial reaches ( $P \leq 0.001$ ). Comparison between the Carron and Elchaig catchments indicates that there are no significant differences in  $W$ - $A$  scaling exponents for bedrock, alluvial, and mixed channels, and the  $D$ - $A$  scaling relation for bedrock channels, derived for each catchment (ANCOVA  $P$  values  $> 0.05$ ).  $W$ - $A$  and  $D$ - $A$  scaling relations for the different channel types in the Canaird catchment were not found to be significant due to limited data (Table 3). The lack of scaling precludes comparison with the Carron and Elchaig catchments, but all data from the Canaird lie within the range of values from the other two catchments (Figure 3) and have been included in the following analysis of the combined data set from all channels.

The  $W$ - $A$  scaling exponents ( $b'$ ) for bedrock ( $0.30$ ) and alluvial ( $0.41$ ) channels, derived from all the data, are significantly different when tested with ANCOVA ( $P < 0.001$ ), indicating that alluvial channel width increases more rapidly with  $A$  than does bedrock channel width. Comparison of the  $D$ - $A$  scaling relations indicates that only the intercepts in the regressions are significantly different, with alluvial channel depths approximately 60% of the bedrock channel depth for a given drainage area (Table 3). A reference bedrock channel at a drainage area of  $50 \text{ km}^2$  is predicted to be  $15 \text{ m}$  wide and  $3 \text{ m}$  deep compared to an alluvial channel  $24 \text{ m}$  wide and  $2 \text{ m}$  deep. Bedrock channels have correspondingly lower  $W/D$  ratios relative to alluvial channels, with average  $W/D = 5.9 \pm 0.4$ , compared to average alluvial channel  $W/D = 14.7 \pm 0.8$  (uncertainty is the standard error of the mean).

The  $W/D$  ratio of both alluvial and bedrock channels are weakly positively scaled with  $A$  (Figure 4). The alluvial



**Figure 5.** Lithology and bedrock channel geometry. (a) Mean reference width ( $W_R$ ) and (b) mean  $W/D$  for bedrock reaches grouped by rock type (Table 1). The black error bars show the 95% confidence interval for the mean. (c)  $W/D$  for each bedrock reach versus the Schmidt hammer  $R$  value and (d) average fracture spacing. The grey error bars show  $1\sigma$  uncertainty on  $W/D$  derived from the standard errors of the reach-averaged widths and depths. Note that three bedrock reaches in metasiltstone were not included in the rock strength assessment.

and  $p'$  (0.30 and 0.24, respectively; Table 3), highlighting that, for both alluvial and bedrock channels, the  $W/D$ - $A$  scaling is consistent with differences in the rate of change of channel width and depth with discharge (drainage area).

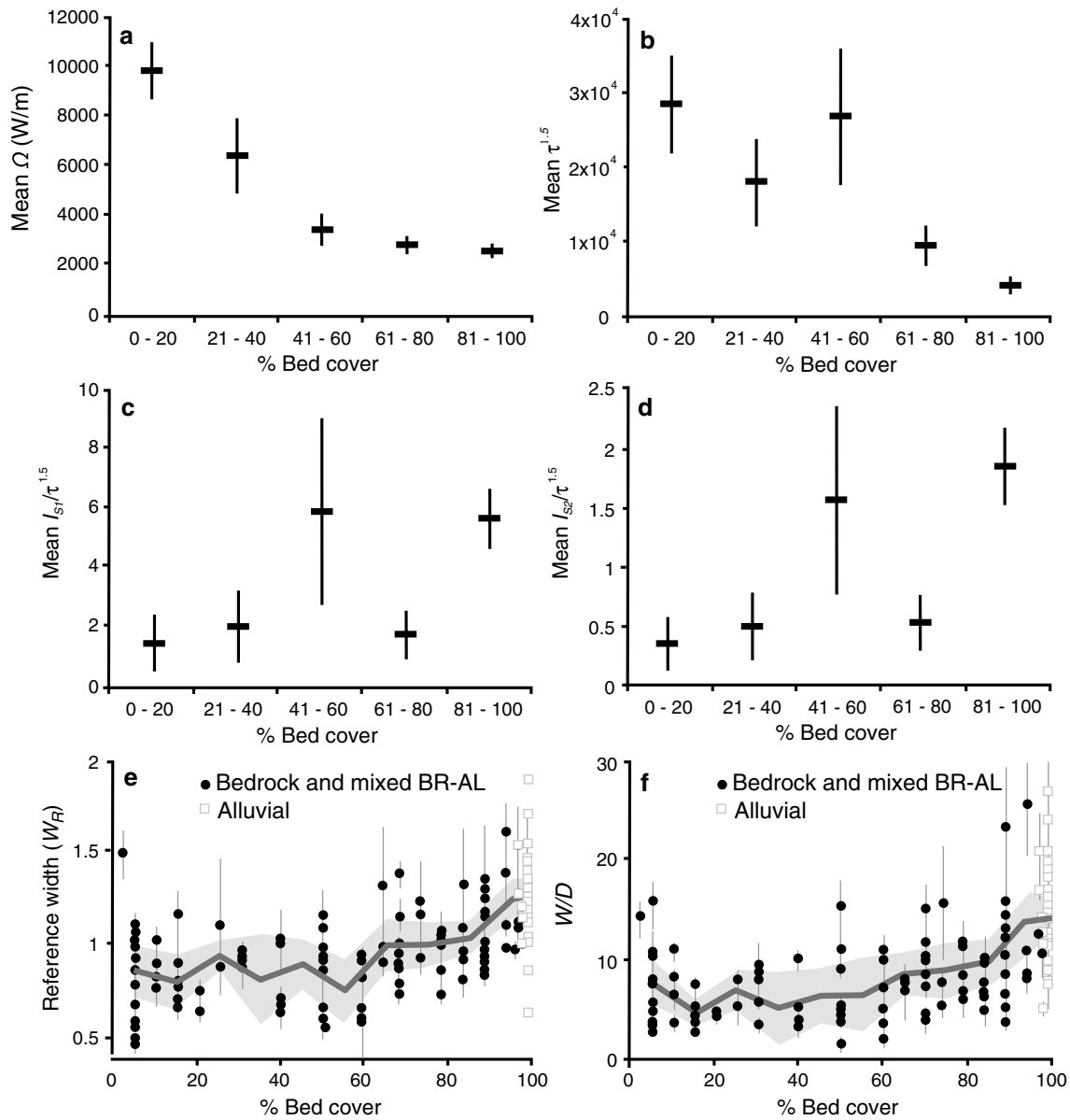
#### 4.2. Lithological Control on Bedrock Channels

Local changes in bedrock channel width at geological transitions have been observed in some settings [Allen *et al.*, 2013; Jansen, 2006; Montgomery and Gran, 2001], suggesting that variations in rock resistance associated with lithology or fracturing may influence  $W$  and  $W/D$ . The influence of lithology on bedrock channel geometry was assessed using the bedrock channel reference widths ( $W_R$ ; equation (8) with  $k_w' = 4.57$  and  $b' = 0.30$ ).

Mean values of bedrock channel  $W_R$  and  $W/D$  calculated for channels in gneiss, metamudstone, metasandstone, and fault-affected rocks are not significantly different ( $P > 0.05$ ; Figures 5a and 5b), but channels in metasandstone with strong vertical and horizontal jointing ("jointed metasandstone") are relatively narrow. Mean  $W_R$  for the jointed metasandstone reaches is significantly lower than for the fault-affected, gneiss, and metasandstone reaches. However, it should be noted that sample sizes for all groups except metasandstone are small (Table 1). The  $W/D$  ratio for each reach is not correlated with compressive rock strength estimated by the Schmidt hammer " $R$ " value or with the average fracture spacing (Figures 5c and 5d). However, the maximum  $W/D$  ratio appears to decrease with increasing fracture spacing, suggesting that only channels in densely jointed rocks maybe capable of maximizing their erosion potential.

#### 4.3. Slope and the Channel Cross Section

Bedrock channels form at steeper slopes for a given drainage area relative to alluvial channels (Figure 3c). The transition between alluvial and bedrock channels can be represented by a critical slope ( $S_c$ ), derived using



**Figure 6.** (a) Mean total stream power ( $\Omega$ ), (b) mean shear stress to the power 1.5 ( $\tau^{1.5}$ ), (c) mean  $I_{S1}/\tau^{1.5}$ , and (d) mean  $I_{S2}/\tau^{1.5}$  calculated for data grouped according to the degree of bed cover (20% bins). Sample numbers for the bins range from 12 (21–40%) to 64 (80–100%); the black error bars show the standard error of the means. Derivation of shear stress and sediment flux ( $I_S$ ) is discussed in the text. Stream power was derived from equation (1) using median annual flood discharge estimated from equation (5) using regional compilations of gauging data yielding:  $Q = 2 \times 10^{-7} A^{1.01}$  for the Rivers Canarid and Elchaig from data for western Scotland and  $Q = 8 \times 10^{-6} A^{0.79}$  for the River Carron based on data from the River Dee in eastern Scotland [cf., *Marsh and Hannaford, 2008*] (Figures S1 and S2 in the supporting information). (e) Reference widths ( $W_R$ ) and (f)  $W/D$  versus the percentage bed cover for each reach, with mean values for 10% bed cover bins (dark grey line) and the 95% confidence interval of the mean (light grey shading).

discriminant analysis, where  $S_c = 0.05 A^{-0.36}$ , which correctly discriminates over 93% of the bedrock and alluvial reaches [cf., *Massong and Montgomery, 2000; Montgomery et al., 1996*]. The critical slope reflects a hydraulic control on the distribution of channel types, leading to an inverse correlation of bedrock channel slope with area ( $R^2 = 0.56$ ,  $P < 0.001$ ) and a weak but significant  $S$ - $A$  scaling for alluvial channels ( $R^2 = 0.16$ ,  $P = 0.008$ ; Figure 3c).

To account for the covariance of  $S$  with  $A$ , the relationship between channel slope and width was assessed using reference widths calculated from the  $W$ - $A$  scaling relation for all data (equation (8) with  $k_w' = 4.36$  and  $b' = 0.37$ ). Both  $W_R$  and  $W/D$  values for alluvial channels are typically higher than bedrock channels for



a given slope (Figure 4), indicating that the differences in geometry of the bedrock and alluvial channels are due to substrate and not differences in channel slope. There is no significant scaling of  $W_R$  with slope for either bedrock or alluvial channels, but  $W/D$  for bedrock channels is weakly, but significantly, inversely scaled with slope ( $P < 0.05$ ). However, because  $S$  and  $A$  are strongly correlated in bedrock channels (Figure 3c), their apparent  $W/D$ - $S$  scaling may be an artifact of the downstream increase in  $W/D$  with  $A$ . This interpretation is supported by  $W/D$  scaling with  $A$  and not  $S$  in alluvial channels, which have a much weaker  $S$ - $A$  relationship (Figures 3c and 4).

#### 4.4. Bed Cover and Sediment Flux

In a given reach, the relationship between bed cover and long-term  $Q_s/Q_c$  may be affected by the local and/or short-term effects of sediment pulses and storm events. In the Scottish Highlands, infrequent debris flows and landslides rarely reach the main stem channels, and there is no evidence of their influence on the study streams. We postulate that bed cover is likely to be reflective of the long-term sediment flux through the reach in these channels, and thus, higher bed cover should be associated with higher values of  $Q_s/Q_c$ , as estimated here by the ratio of the sediment index to shear stress to the power 1.5.

Extensive bed cover is typically found in reaches with low stream power and low shear stress (Figures 6a and 6b), although there is considerable variability in shear stress for reaches with less bed cover as indicated by high standard errors. Bed cover also tends to increase with both  $I_{s1}/\tau^{1.5}$ , and  $I_{s2}/\tau^{1.5}$ ; reaches with less than 20% bed cover have a mean sediment flux index that is 4 times lower than that for reaches with over 80% bed cover (Figures 6c and 6d). Both  $W/D$  and reference widths for all the data (equation (8) with  $k_w' = 4.36$  and  $b' = 0.37$ ) tend to increase when bed cover exceeds 50% of the channel bed (Figures 6e and 6f) but are relatively constant for reaches with less bed cover. The considerable variability in bed cover fractions may reflect the influence of variations in bed roughness, grain size, or the spatial distribution of sediment sources, which were not investigated in detail.

When  $W/D$  is compared with modeled sediment fluxes, alluvial and mixed bedrock-alluvial channels are found to occur at higher values of  $I_{s1}/\tau^{1.5}$  and  $I_{s2}/\tau^{1.5}$ , and there is a weak positive scaling of  $W/D$  with  $I_{s1}/\tau^{1.5}$  for alluvial channels ( $P < 0.05$ ; Figure 7). Bedrock channels typically occur at lower  $I_{s1}/\tau^{1.5}$  values, and their  $W/D$  is weakly positively scaled with both  $I_{s1}/\tau^{1.5}$  and  $I_{s2}/\tau^{1.5}$  ( $P < 0.05$ ). We find no obvious difference in scaling between channels upstream and downstream of lakes, except for a few notable outliers with high  $W/D$  ratios occurring at low  $I_{s1}/\tau^{1.5}$  values (Figure 7). These high  $W/D$  values at low  $I_{s1}/\tau^{1.5}$  correspond to reaches located immediately downstream of lake outlets (Figure 8). Across the low range of  $I_{s1}/\tau^{1.5}$  ( $< 0.02$ ), the high  $W/D$  outliers yield a mean value of  $13.2 \pm 4.5$ , which is significantly greater than the mean  $W/D$  of the remaining data:  $4.6 \pm 0.6$  and  $4.8 \pm 0.5$  for  $I_{s1}/\tau^{1.5}$  and  $I_{s2}/\tau^{1.5}$  data subsets, respectively (two sample  $t$  tests  $P = 0.015$ , mean  $\pm$  90% confidence interval).

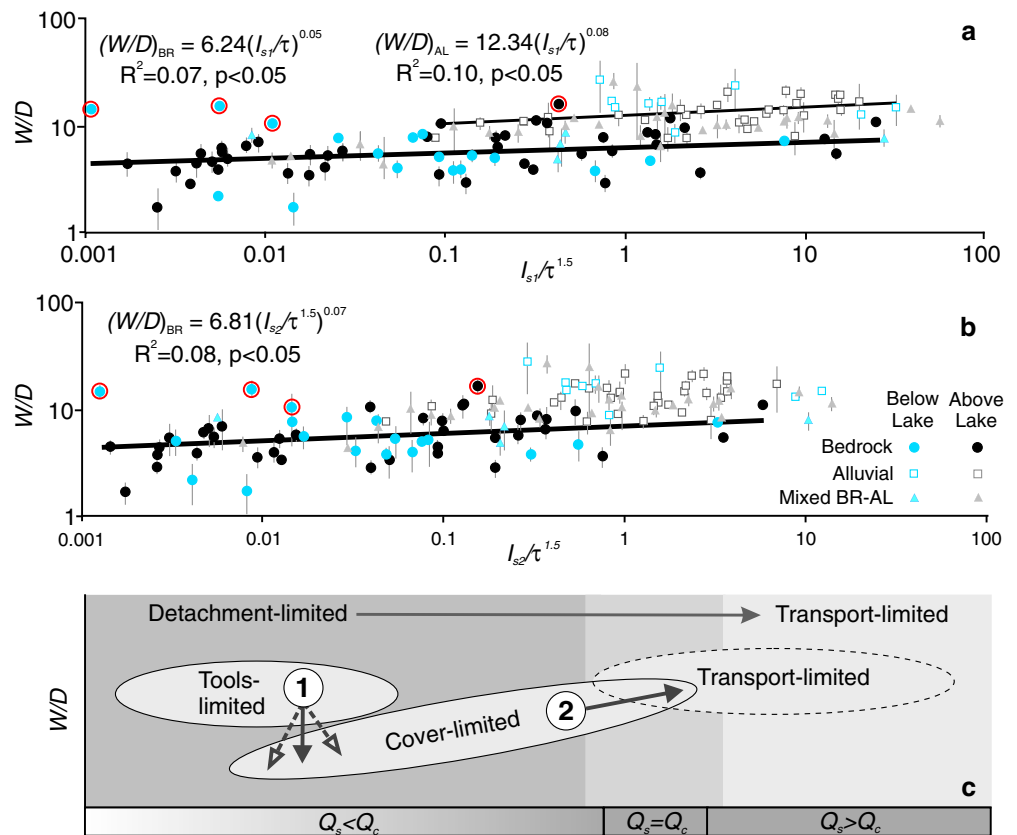
## 5. Discussion

### 5.1. Summary of Controls

Our results show that the cross-section geometry of Scottish Highland streams is primarily a function of drainage area, which can be related to discharge, with bank and bed materials exerting secondary control. Channel slope was not found to directly influence channel width but does play a critical role in governing the bedrock-alluvial channel transition (where  $Q_s = Q_c$ ) due to its influence on sediment transport capacity. Slope therefore influences channel morphology by controlling the distribution of bedrock and alluvial channels and the degree of bed cover. Sediment, affected by the distribution of lakes, appears to have a relatively minor but complex influence on the morphology of bedrock channels within the study catchments. We consider these controls on channel morphology in more detail in the following discussion.

### 5.2. Substrate Control of Channel Geometry

The strong power law scaling observed between bedrock channel width and depth with catchment area accounts for 85% of the variation in channel width and  $>60\%$  of the variation in depth. Bedrock channel width is typically one third to two thirds that of alluvial channels for a given drainage area. Systematic differences in bedrock and alluvial channel geometry support the hypothesis that bedrock channels will be narrower than alluvial channels for a given drainage area (hypothesis 1) and contrasts with

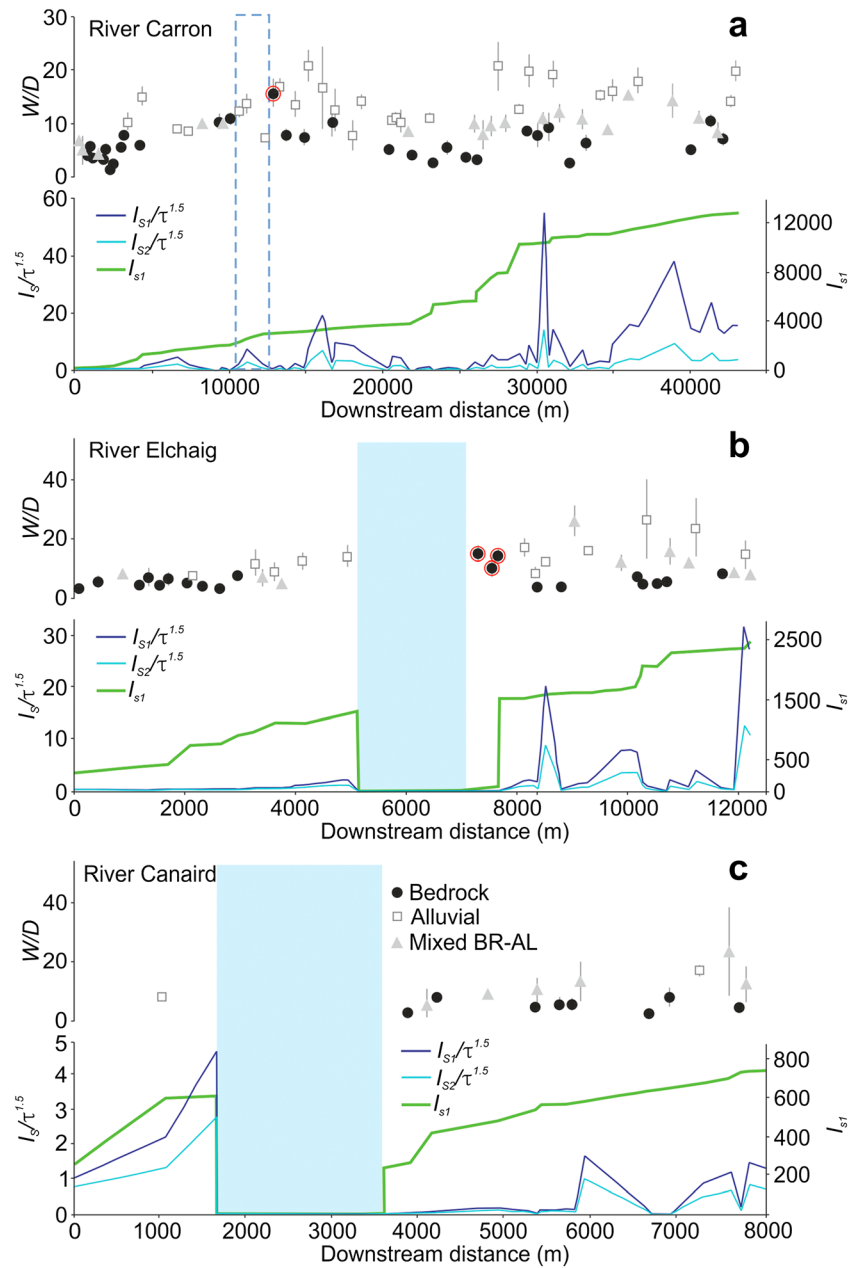


**Figure 7.** Channel  $W/D$  as a function of (a)  $I_{S1}/\tau^{1.5}$  and (b)  $I_{S2}/\tau^{1.5}$ . Power law-scaling relations for bedrock channel data are significant at  $P < 0.05$ . The derivation of the sediment indices ( $I_{S1}$  and  $I_{S2}$ ) and shear stress ( $\tau^{1.5}$ ) are discussed in the text. High  $W/D$  outliers from the bedrock channel regression are marked by the red circles and discussed in the text. (c) Sketch plot of  $W/D$  versus sediment flux ( $I_{S2}/\tau^{1.5}$ ) with interpretation of process domains. The grey arrows indicate hypothetical increases in sediment related to changes in channel form; an increase in  $Q_s/Q_c$  at low values (1) may result in a decrease in  $W/D$ , and at higher  $Q_s/Q_c$ , an increase in  $Q_s/Q_c$  may result in an increase in  $W/D$  (2).

observations from previous comparative studies [Montgomery and Gran, 2001; Wohl and David, 2008]. The bedrock channel  $W$ - $A$  scaling exponent ( $b' = 0.3$ ) is lower than the 0.5 value that is commonly assumed in stream-power-based erosion models [e.g., Howard et al., 1994; Whipple, 2004] but is similar to values reported elsewhere (Table 4) [Lague, 2014]. The lower scaling exponent indicates that bedrock channel width typically shows less variation with catchment area than does alluvial channel width, supporting hypothesis 2.

Although bedrock and alluvial substrates were observed to influence channel morphology, lithology has a minor influence on the morphology of the bedrock channels studied here; only those channels cut in strongly vertically jointed rocks were found to be generally narrower and deeper. The minor influence of lithology may be due to the relatively small range of rock strength exhibited by metamorphic rocks sampled in the Highland streams. A compilation of  $W$ - $A$  scaling exponents from published studies indicates that  $b'$  values vary with substrate type (Table 4). The mean value of  $b'$  for bedrock channels cut in resistant metamorphic and igneous rocks is  $0.30 \pm 0.04$ . But, for weaker sedimentary rocks, a mean value of  $b'$  of  $0.46 \pm 0.07$  is found, very similar to the exponent for channels in unconsolidated alluvial sediment (mean  $0.46 \pm 0.09$ ).

Differences in bedrock and alluvial channel geometry are likely to be most pronounced in landscapes composed of very resistant rocks, and the influence of lithology on bedrock channel geometry may be apparent only where channels in rocks with large differences in resistance are compared. When modeling channels and landscapes sculpted in weak sedimentary rocks, use of a single  $b' \approx 0.4$ – $0.6$  may adequately account for downstream variation in channel geometry regardless of reach type, although different values



**Figure 8.** Downstream variations in  $W/D$ ,  $I_{s1}/\tau^{1.5}$ ,  $I_{s2}/\tau^{1.5}$ , and  $I_{s1}$  for the (a) River Carron, (b) River Elchaig, and (c) River Canaird. The locations of lakes on the Rivers Elchaig and Canaird are shown by blue shading. The area marked by dashed lines in Figure 8a marks the location of a broad alluvial plain associated with a deep alluvial channel interpreted as an infilled lake. Outlier reaches (red circles) from the regressions of bedrock channel  $W/D$  with  $I_{s1}/\tau^{1.5}$  and  $I_{s2}/\tau^{1.5}$  (Figure 7) are discussed in the text. These outliers represent bedrock reaches immediately downstream of a lake (Elchaig) or infilled lake (Carron), and they exhibit the greatest  $W/D$  ratios of all bedrock reaches along both rivers. Note the different horizontal and vertical scales.

of  $k_w$  or  $k_w'$  may need to be applied (Table 4). However, in regions characterized by resistant metamorphic and igneous rocks such as those characteristic of cratons and postorogenic belts, or terrains with marked contrasts in rock resistance [e.g., Allen *et al.*, 2013], the use of a single scaling exponent may fail to account for variability in channel geometry at subcatchment scales.

### 5.3. Sediment, Slope, and Channel Morphology

Bedrock channel  $W/D$  is commonly assumed or predicted to be constant in purely detachment-limited erosion models [e.g., Finnegan *et al.*, 2005; Wobus *et al.*, 2006a]. However,  $W/D$  is thought to scale with

**Table 4.** Published Empirical Channel Width-Area Scaling Relations<sup>a</sup>

Geology	Bedrock $W-A$ Exponent ( $b'$ )	Alluvial $W-A$ Exponent ( $b'$ )	River/Region	Study
Metasediments	0.30	0.41	Scottish Highlands (NW)	This study
Metasediments	-	0.53	Scottish Highlands (E)	S. Addy (Hierarchical controls on channel morphology in montane catchments in the Cairngorms, north-east Scotland, University of Aberdeen, unpublished PhD Thesis, 2010)
Granite/limestone	0.37	0.45	Yuba River, California	Montgomery and Gran [2001]
Quartzite	0.26	-	Scottish Highlands (W)	Jansen et al. [2010]
Metasediments (nonquartzite)	0.28	-	Scottish Highlands (W)	Jansen et al. [2010]
Sedimentary and igneous ( $R = 40-50$ )	$0.35 \pm 0.04^b$	-	Northern California, USA	Snyder et al. [2003]
Metamorphic and sedimentary ( $R = \sim 40$ )	$0.28 \pm 0.02^c$	-	Central Nepal	Craddock et al. [2007]
Metamorphic and igneous	$0.33 \pm 0.17$	-	San Gabriel Mountains, California	Dibiase and Whipple [2011]
Mean <sup>h</sup>	$0.27^d$	-		
Standard error (SE) mean	$0.22^e$			
Standard deviation	$0.30 \pm 0.04$			
Sedimentary	0.02			
Sedimentary	0.05			
Sedimentary	0.53	0.32	Knowles Creek, California ( $>1 \text{ km}^2$ )	Montgomery and Gran [2001]
Poorly consolidated Sandstone and shale	0.51	0.55	Sullivan and Larson Creeks, California	Montgomery and Gran [2001]
Sandstone and mudstone	0.47	0.49	Satsop Creek, California	Montgomery and Gran [2001]
Sandstone and mudstone	$0.43^f$	-	California, USA	Duvall et al. [2004]
Mean <sup>h</sup>	$0.33^g$			
Sandstone and mudstone	0.42	-	Clearwater River, Washington State, USA	Tomkin et al. [2003]
Sandstone and mudstone	0.54	-	Peikang River, Taiwan	Yanites et al. [2010]
Mean <sup>h</sup>	$0.46 \pm 0.07$	$0.46 \pm 0.09$		
SE mean	0.03	0.04		
Standard deviation	0.08	0.09		

<sup>a</sup>Data grouped according to the general bedrock geology.

<sup>b</sup>Average for four catchments in a high uplift zone, bedrock, and alluvial channels not discriminated.

<sup>c</sup>Average for two catchments in a low uplift zone, bedrock, and alluvial channels not discriminated.

<sup>d</sup>Low-relief zone.

<sup>e</sup>High-relief zone.

<sup>f</sup>High-uplift zone.

<sup>g</sup>Low-uplift zone.

<sup>h</sup>Mean values in italics. Two sample  $t$  tests indicate that the mean  $b'$  value for bedrock channels in metamorphic and igneous terrains is significantly different to that of both bedrock channels in sedimentary bedrock terrains and alluvial channels ( $P < 0.005$ ), whereas the  $b'$  values for bedrock channels in sedimentary rocks are not significantly different to that of alluvial channels ( $P = 0.95$ ).



catchment area and/or slope when erosion is dependent on sediment flux [e.g., Turowski *et al.*, 2007; Wobus *et al.*, 2008]. Similarly,  $W$ - $S$  scaling has been suggested to deviate from a purely detachment-limited relation, where  $W \sim S^{-3/16}$  [Finnegan *et al.*, 2005; Wobus *et al.*, 2006a], in the presence of sediment cover [Turowski *et al.*, 2007; Yanites and Tucker, 2010]. A tendency for sediment cover, i.e.,  $Q_s/Q_c$ , to increase downstream would also accentuate  $W$ - $S$  scaling, as higher cover increases lateral bank erosion relative to vertical incision [Turowski *et al.*, 2007; Wobus *et al.*, 2008; Yanites and Tucker, 2010]. However, observations of scaling in natural channels may be complicated by downstream variations in the distribution of sediment sources and sinks.

Positive scaling of  $W/D$  with  $A$  in the Highland channels appears consistent with the predictions of sediment control on the channel cross section (Figure 4a), but  $W/D$  is also scaled with  $A$  in alluvial channels (>99% bed cover). Furthermore, although  $W/D$  does tend to increase with bed cover in bedrock channels, particularly where bed cover exceeds ~50% (Figure 6), the wide variation in channel slope and the effects of lakes on sediment supply mean that bed cover does not systematically increase downstream. It is possible that the scaling of  $W/D$  with  $A$  arises from the tendency for roughness to decrease relative to channel size, resulting in a downstream increase in flow velocity and a reduction in  $D$  relative to  $W$ , as observed in alluvial channels [Knighton, 1998].

Both  $W_R$  and  $W/D$  tend to have roughly constant values with sparse bed cover (Figures 6e and 6f), which is consistent with predictions for detachment-limited channels [Finnegan *et al.*, 2005; Wobus *et al.*, 2006a], although there are several notable high values where bed cover is minimal. The absence of  $W_R$ - $S$  scaling contrasts with previous findings of width-slope dependence in detachment-limited channels ( $W \sim S^{-3/16}$ ) [Finnegan *et al.*, 2005] or in channels with higher sediment flux, where  $W$  is thought to scale more strongly with  $S$  [cf., Yanites and Tucker, 2010]. In the study channels, slope does not appear to directly control channel width (cf., hypothesis 3), but the lack of  $W_R$ - $S$  scaling may reflect downstream variations in sediment flux.

The weak positive scaling of  $W/D$  with  $I_s/\tau^{1.5}$  (Figure 7) is compatible with the prediction of hypothesis 4 that increasing  $Q_s/Q_c$  gives rise to wider, shallower bedrock channels; however, the scaling observed may also reflect covariance between sediment flux, slope, and channel cross-section morphology. Our results suggest that sediment trapping in lakes has a minimal impact on channel geometry, with no obvious distinction between channels above and below lakes in terms of  $W/D$  or  $I_s/\tau^{1.5}$  (Figures 7a, 7b, and 8). However, we note some outliers associated with high  $W/D$  values at low  $I_s/\tau^{1.5}$  (Figure 7) and minimal bed cover (Figures 6e and 6f), which may testify to a local sediment control on channel geometry. These outliers indicate wide, shallow bedrock channels immediately below lake outlets and below a wide alluvial plain acting as a long-term sediment store on the River Carron (Figure 8). These reaches are poorly incised, and their sparse bed cover ( $\leq 5\%$ ) is composed of angular boulders derived from thin local till deposits or weathered blocks plucked from the channel bed (Figure 2b). The wide, shallow channels are interpreted to occur at low  $Q_s/Q_c$  values because erosion has been limited by a lack of available sediment; the channels have not been able to erode sufficiently since deglaciation to modify their boundaries. The morphological effects of restricted sediment availability are confined to reaches immediately downstream from lake outlets (Figure 8). Below lakes, bedrock channels occurring downstream from tributary inputs (Rivers Elchaig and Canaird; Figures 8b and 8c), or alluvial reaches incised into glacial valley-floor deposits (River Carron; Figure 8a), have lower  $W/D$  ratios consistent with areas above the lakes. Renewal of sediment supply below lakes is sufficient to reestablish the hydrological control on river geometry despite the lack of supply from the catchment headwaters.

The results allow us to speculate on the nature of the relationship between sediment and bedrock channel morphology by considering potential channel response to an increase in sediment supply (Figure 7c). Bedrock channel response to an increase in  $Q_s$  may vary depending on the initial  $Q_s/Q_c$ , and the complex relationship may help explain the complex morphological responses observed in channels affected by changes in discharge and sediment supply [cf., Snyder and Kammer, 2008]. In channels with high  $Q_s/Q_c$ , increasing  $Q_s$  may result in an increase in  $W/D$  as increased bed cover promotes lateral erosion, with a concomitant decrease in shear stress (i.e.,  $Q_c$ ) [cf., Finnegan *et al.*, 2007; Johnson and Whipple, 2010]. Increasing  $Q_s/Q_c$  may thus result in alluviation of the channel and a switch from detachment-limited to transport-limited processes (Figure 7c). In bedrock channels with low  $Q_s/Q_c$ , restricted sediment supply, resulting in a lack of "tools" to drive erosion [cf., Sklar and Dietrich, 1998], may be associated with

extremely wide, shallow channels. A slight increase in  $Q_s$  may result in considerable narrowing and deepening of these channels through incision, with a concomitant increase in shear stress (Figure 7c). The reduction in  $W/D$  may mark a transition from a “tool-limited” to a “cover-limited” domain, in which channel morphology develops in relation to downstream variations in discharge (area), slope, and sediment supply (which may be considered as a function of catchment area or stream length). As low  $W/D$  configurations are associated with higher shear stress and greater potential for vertical erosion, these results may indicate that in Highland streams, small amounts of sediment are necessary for optimal incision. If so, this result differs from predictions derived from cover-dependent models, which assume that  $W/D$  approaches a constant value at low  $Q_s/Q_c$ , and therefore, erosion is most efficient in purely detachment-limited channels [e.g., Turowski *et al.*, 2007; Yanites and Tucker, 2010].

#### 5.4. Postglacial Channel Evolution

Strong scaling of width and depth with  $A$  is observed in Scottish channels despite the prevalence of inherited glacial valley-floor slopes and preexisting inner gorges. Yanites *et al.* [2010] postulated that channel width is adjusted to new conditions after perturbation in “the time needed to vertically erode through one channel depth” [p. 1204]. In the Highland streams, the strong bedrock channel  $W$ - $A$  scaling suggests that postglacial erosion has been sufficient to allow most of the study reaches to adjust with respect to the Holocene discharge regime, i.e., to incise to one channel depth. Given the  $D$ - $A$  relationship for bedrock channels (Table 4), which yields depths ranging from 1.8 to 4.6 m for channels at drainage areas of 10–500 km<sup>2</sup>, this condition would require minimum postglacial erosion rates for these channels of between 0.18 and 0.46 m/kyr. Such rates are an order of magnitude higher than regional uplift rates [e.g., Jansen *et al.*, 2011] but are comparable with incision rates quantified in previous studies of Scottish Highland streams (0.4–2.4 m/kyr) [Jansen *et al.*, 2011; K. Whitbread, unpublished PhD Thesis, 2012]. The poorly incised bedrock reaches below lakes are indicative of very low erosion rates (<0.1 m/kyr), thereby highlighting the local importance of sediment flux in these streams.

The evolution of the geometry of Highland streams demonstrates that fluvial erosion is an important component of landscape change following deglaciation even in areas unaffected by tectonic uplift or relative base-level fall. The incision and narrowing of bedrock channels at knickzones inherited from glacial erosion affect the form and base level of surrounding hillslopes, making it an important component of postglacial landscape evolution [cf., Ballantyne, 2002]. In postorogenic terrains, however, the elevated rates of erosion that occur in bedrock streams are not matched by background uplift rates. Without renewal of relief through tectonic uplift or base-level lowering, channel slope will tend to decrease as incision progresses. As bedrock channels are narrowed and deepened, bed lowering would tend to reduce the stream transport capacity and increase bed cover (in the absence of other external changes). Our results suggest that this increased bed cover may promote channel widening, but it may also lead to a decrease in the vertical erosion rate [e.g., Sklar and Dietrich, 2001]. The feedback between the channel cross section, slope, sediment, and erosion rates in tectonically passive terrains may thus result in systems in which so-called “transient dynamics” could persist for many millions of years [Baldwin *et al.*, 2003; Egholm *et al.*, 2013].

## 6. Conclusions

In the postglacial channels of the Scottish Highlands, the distribution of bedrock and alluvial reaches is constrained by the inheritance of irregular, glacially eroded valley floors. The high resistance of metamorphic and intrusive igneous rocks in this terrain gives rise to contrasting bedrock and alluvial channel dynamics, allowing the evolution of strongly detachment-limited and transport-limited channels to be directly compared. Bedrock channels develop narrower, deeper cross sections than alluvial channels, and their form is less strongly affected by variations in discharge (scaling as  $W \sim A^{0.3}$ , compared to  $W \sim A^{0.4}$  for alluvial channels). The strong width scaling with catchment area in bedrock streams indicates that postglacial fluvial erosion in the 12 to 15 kyr since deglaciation has been sufficient to (re)configure the channel morphology. This adjustment has occurred in those channels incised in the valley floor after deglaciation and in reaches confined by pre-Holocene rock gorges.

Relative sediment flux may also influence the morphology of bedrock channels. The  $W/D$  of bedrock reaches depends on discharge (or area), slope, and sediment flux; increasing  $W/D$  in bedrock channels is associated

with a transition from detachment-limited to transport-limited systems. Sediment trapping in lakes has a minimal impact on channel geometry, except immediately downstream of lake outlets, where lack of available sediment has resulted in limited erosion, leading to locally steep, wide, and shallow bedrock channels.

The distinct behavior of detachment-limited and transport-limited channels means that in resistant bedrock terrains like the Highlands, width-discharge (drainage area) scaling models describing the spatial variability in channel geometry are only likely to be applicable at catchment scale. At the reach scale, channel geometry models must account for spatial variations in fluvial processes and reflect the complex interplay of sediment and substrate in controlling channel morphology. In geological terrains characterized by less resistant sedimentary or strongly fractured rocks, transport-limited conditions are likely to prevail throughout the channel systems. In these terrains, the standard “hydraulic”  $W$ - $Q$  or  $W$ - $A$  scaling models may adequately account for spatial variations in channel morphology at both catchment and reach scales.

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