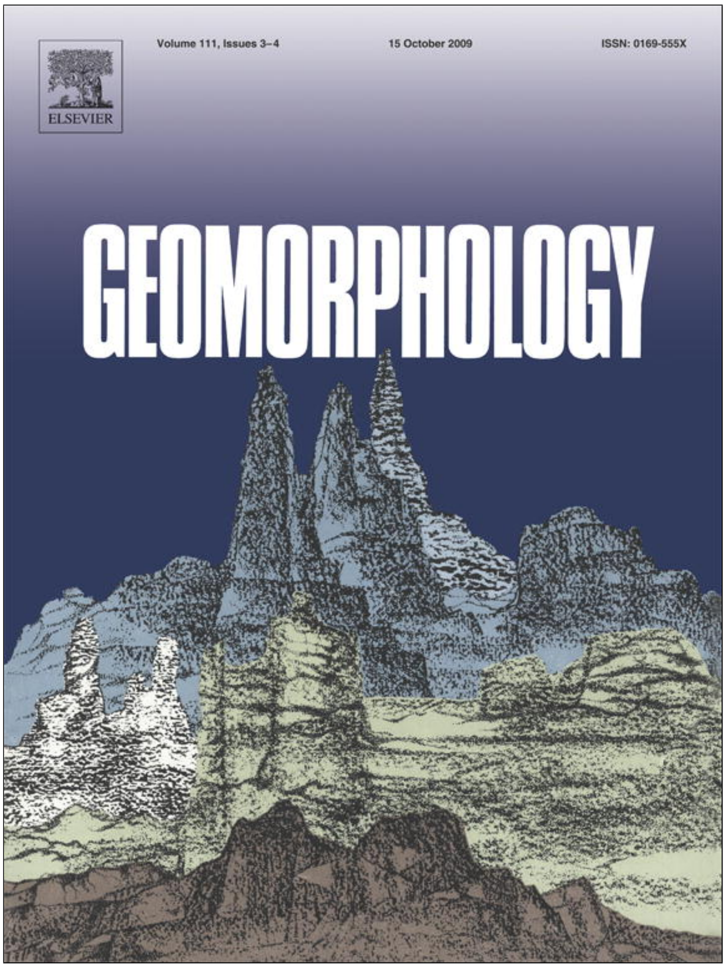
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Influence of rock strength on the valley morphometry of Big Creek, central Idaho, USA

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ABSTRACT

Analysis of valley morphometry and bedrock strength along Big Creek, central Idaho, shows that valley floor width is strongly controlled by bedrock. We performed statistical analysis of Schmidt hammer rock strength as a function of lithology and aspect and of valley morphometry as a function of rock strength. Rock strength is significantly greater on the south side of the valley and in Eocene granodiorites. Rock strength is weakest in Eocene volcanic tuffs. Valley floor width depends negatively on weakest valley-side rock strength, and hillslope gradient on the north side of the valley depends positively on rock strength. Stream gradient does not depend on rock strength. Valley floor width appears to be controlled by bedrock strength on the weaker side of the valley, which was generally the north (south-facing) side. We speculate that a higher degree of weathering via freeze–thaw cycles contributes to lower strength on the north side. The positive dependence of hillslope gradient on rock strength on the north side provides evidence that differential weathering across lithologies determines the gradient that can be maintained as lateral migration of the stream erodes valley walls. These results suggest that in situ rock strength exerts strong influences on some measures of valley morphometry by modulating hillslope mass wasting processes and limiting lateral erosion.

1. Introduction

Since the nineteenth century, rock strength has been recognized as an important influence on valley and canyon morphometry (Gilbert, 1877; Powell, 1895). Despite decades of research, however, few studies have documented relationships between measured rock strength and valley morphometry. Therefore, the specific relationships between rock strength and morphometry remain relatively unknown, and detailed studies are necessary to fill those gaps in knowledge.

Parameters of river valley morphometry (such as valley floor width, hillslope gradient, and streambed gradient) are presumably affected to some degree by the strength of the bedrock through which the river is eroding. One generalization of geomorphology is that rivers cutting through hard rock will have narrow, steep streambeds surrounded by steep hillslopes. However, few studies (e.g., Mackley and Pederson, 2004) have tried to quantify the way rock strength controls valley morphometry.

Although Mackley and Pederson (2004) found a correlation between rock strength and river valley morphometry, their study focused on very large rivers cutting through flat-lying, uniform sedimentary rocks of the Colorado Plateau. A lack of empirical studies of smaller-scale, compositionally diverse river valleys still exists. One of the motivations for this study is the desire to understand valley morphometry as a function of the realistic strength properties that exist in the field, rather than simply on lithology. We hope this study will shed light on some of the processes influencing mountain drainage basin development and evolution, particularly how rock strength properties affect erosion.

Rock strength is a fundamental resisting force in geomorphic processes. However, rock strength encompasses a variety of factors and its influence on geomorphic processes is not always straightforward. Quantifying the relationship between rock strength and valley morphometry allows us to better understand the specific ways in which resisting forces and driving forces interact, which is the very basis of geomorphology.

Measuring the rebound of an unweathered and unfractured rock sample with a Schmidt hammer provides a measure of intact rock strength. Intact rock strength describes the virgin mechanical strength properties of the material but is largely irrelevant because rocks are typically modified by weathering, strain, or discontinuities. Discontinuities in the rock, weathering, surface roughness, microfractures, and joints will affect Schmidt hammer measurements by attenuating energy (Williams and Robinson, 1983; McCarroll, 1991; Sumner and Nel, 2002). Therefore, use of the Schmidt hammer in field conditions measures not only intact rock strength, but also other factors that contribute to overall rock strength (see Appendix A). The Schmidt hammer is convenient because in situ bedrock strength can be

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measured without arduous sample collection and destruction. Use of the Schmidt hammer decreases the detailed knowledge of rock strength parameters, but dramatically increases potential sample size and thus increases statistical validity.

Other studies have investigated the relationship between rock structure and rock strength. For example, Selby (1980) developed the Rock Mass Strength Index for incorporating a variety of modifying factors (weathering, joint spacing, joint orientation, joint width, joint continuity, and outflow of groundwater) with the intact rock strength. Whipple et al. (2000) argued that jointing is the most important control on the type of erosional process. Kahraman (2001) investigated the relationship between *P*-wave velocity, number of joints, and Schmidt hammer rebound value and found that *P*-wave velocity is strongly negatively correlated to the number of joints in a sample and that Schmidt hammer rebound values are strongly positively correlated to *P*-wave velocity. This suggests indirectly that Schmidt hammer rebound value may be negatively correlated to the spacing of joints in a rock sample. In other words, fewer joints yield higher rock strength and higher Schmidt hammer rebound values. In the present study, the Schmidt hammer is used to characterize rock strength in valley walls by measuring the relative rebound of in situ bedrock.

In this article, we report on a study of rock strength and valley morphometry conducted in the Big Creek drainage, a tributary to the Middle Fork Salmon River in central Idaho, USA. The approach of this study was to measure the relative in situ rebound value of bedrock in the valley walls and compare it to three valley parameters: valley floor

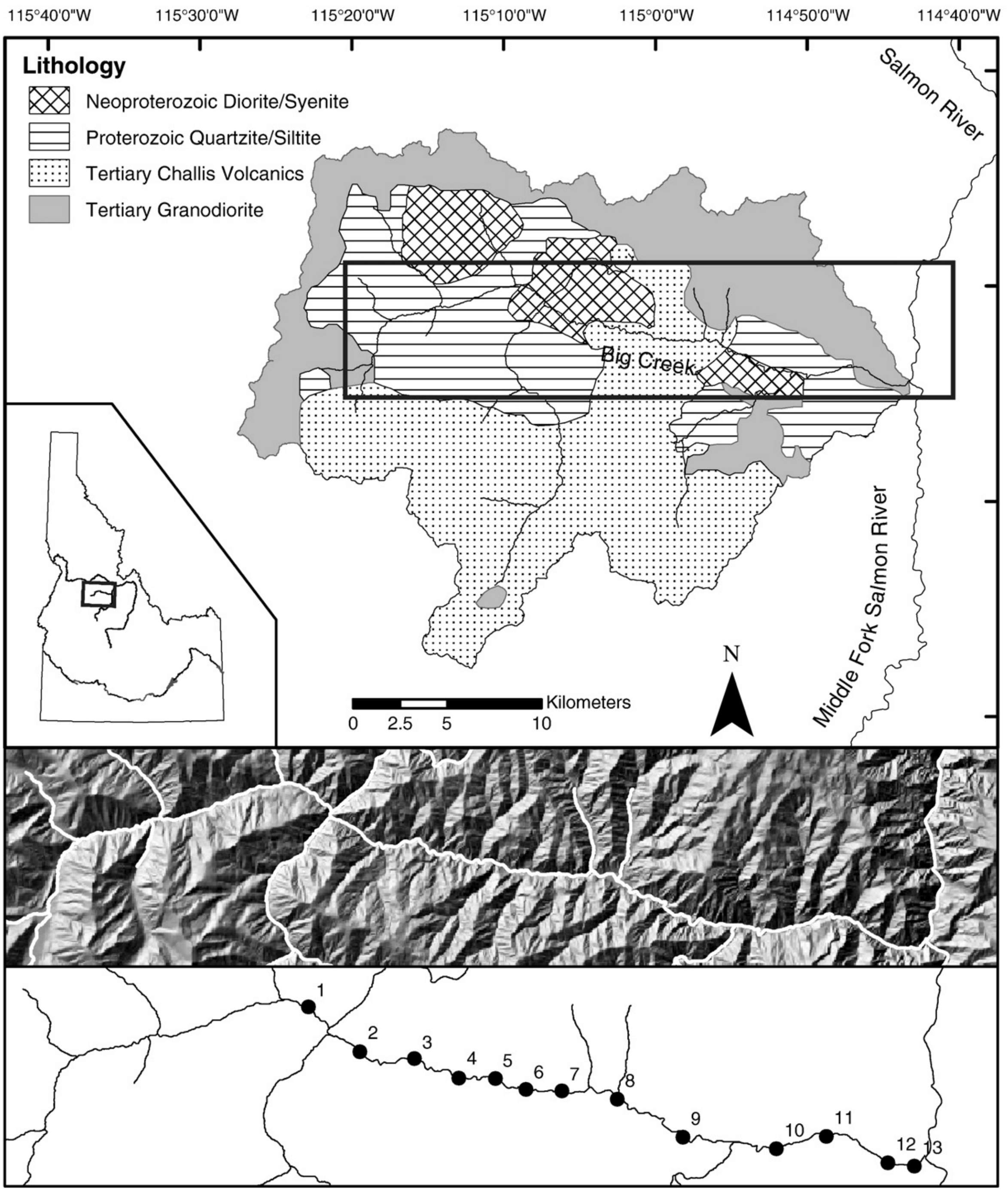

Fig. 1. Map of study area showing lithology, topography, and reach locations. Geology after Lund (2004) and Stewart et al. (Idaho Geological Survey, unpublished maps 2004).

width, channel gradient, and hillslope gradient. The data were analyzed to test two hypotheses: (i) that rock strength differs between valley sides and among lithologic groups, and (ii) that valley morphometry is determined by rock strength. Finally, we offer explanations for the correlations or lack of correlations using principles of rock mechanics and fluvial and hillslope processes.

2. Study area

Big Creek is located in Valley and Idaho Counties in central Idaho (Fig. 1). It is an east-flowing tributary to the Middle Fork Salmon River, which eventually drains to the Salmon River, and is part of the larger Columbia River drainage basin. The Big Creek drainage basin covers 1539 km^2, and the main trunk of the stream is 67 km long. Mean elevation of the basin is 2101 m. Basin relief is 1876 m, with a maximum elevation of 2906 m and a minimum elevation of 1030 m at its outlet (Fig. 1). The drainage basin lies entirely within the Payette National Forest and the 9300-km^2 Frank Church–River of No Return Wilderness Area. This setting is ideal for assessing relationships of intact rock strength and canyon morphometry because of the minimal anthropogenic alteration, the narrow canyon, and the varied rock types along its length.

The Miocene to Recent uplift history of central Idaho is not well established. Significant uplift has occurred to create the high elevation and high relief that characterize the region. Topographic maps and DEMs show that the Salmon River Mountains, especially north of Big Creek, have many high, apparently concordant, plateaus that may define an ancient low-relief topographic surface. Deeply incised river canyons have been cut into this high surface, creating a “bimodal” topography.

Meyer and Leidecker (1999) also noted the bimodal topography described above along the Middle Fork Salmon River and, using an estimated incision rate of 0.12–0.16 m/ky, estimated that the ~300-m-deep inner gorge (i.e., the most recently incised portion of the canyon characterized by steep and narrow walls below a sharp break in slope) was formed since 2.63–1.85 Ma.

Two underlying geologic conditions are important to the morphometry of the modern Big Creek drainage. First is the Eocene Cow Creek fault, a major NE-trending, NW-dipping normal fault with weak Eocene volcanic tuff in the headwall and stronger Eocene diorite in the footwall. This fault is a major structural control on the longitudinal gradient of Big Creek. The stream gradient upstream of this point is adjusted to the local base level of the fault. Second, a combination of Miocene–Recent surface uplift and Late Miocene–Pliocene lowering of base level has caused incision and created a knickpoint that has migrated up Big Creek from the Middle Fork of the Salmon River. The stream gradient below the Cow Creek fault is controlled by this knickpoint and has not yet equilibrated to the base level drop (Fig. 2).

The underlying geology of the Big Creek drainage basin is diverse, both temporally and compositionally (Fig. 1). The rocks of the Big Creek drainage basin are divided into four groups: (i) Mesoproterozoic

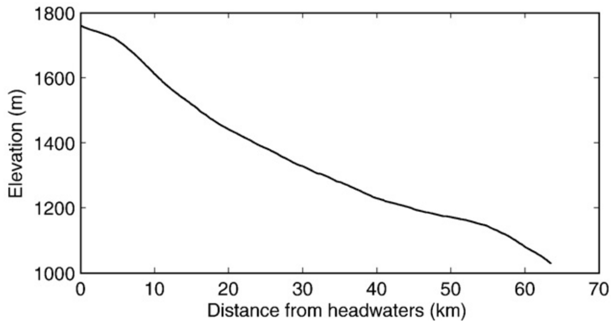

Fig. 2. Longitudinal profile of Big Creek.

quartzites and siltites, (ii) Neoproterozoic diorite/syenite mafic intrusions, (iii) Eocene intrusive granodiorites, and (iv) Eocene volcanic tuffs and porphyry dikes of the Challis Volcanic Group (Lund, 2004). The map units described by Stewart et al. (Idaho Geological Survey, unpublished maps, 1995–2004) are adopted in this study. Field observations indicate that the Eocene volcanic tuffs and porphyry of the Challis Volcanic Group are by far the most variable lithologic group. Mafic intrusive rock weathers easily along mineral grain boundaries, and granodiorite commonly weathers to grus. The Mesoproterozoic quartzite/siltite lithologic group tends to be less resistant to weathering and is generally more consistent in the way it weathers, alters, and fractures than the other lithologic types. The local contrasts in rock type and strength are key to the rock strength controls we explore in this study.

3. Methods

Three types of data were collected in this study. First, rock strength data for both sides of the valley were obtained with the Schmidt hammer. Second, valley floor widths were measured in the field with a laser rangefinder. Third, the stream gradient and hillslope gradient of each reach were calculated using digital elevation models (DEMs) and topographic maps. Additional data extracted from DEMs included relief, main trunk channel length, mean basin elevation, basin area, and local hypsometric data. Methods and ancillary measurements are detailed fully in Lifton (2005).

3.1. Reach delineation

Thirteen reaches were chosen for detailed study on the basis of valley floor width, as measured on U.S. Geological Survey (USGS) 7.5′ topographic maps. Reaches are numbered sequentially, upstream to downstream (Fig. 1). Valley segments of uniform width bounded by marked changes in width at both ends were identified. This follows the general criteria outlined by Grant and Swanson (1995). A range of widths was chosen to represent the variety of morphometry in Big Creek valley. Reaches that encompass major tributary junctions were not used because of the confounding effects of a point source sediment supply. Reach #8, at the confluence of Cabin Creek (Fig. 1), is an exception to this rule. It was retained as a study reach because Big Creek valley is wide in this reach, and Cabin Creek is a minor tributary that does not contribute enough sediment to alter the morphometry of the valley floor or stream channel. Most reaches are bounded by bedrock outcroppings. In some cases, reach boundaries (i.e., distinct changes in valley floor width) correspond to lithologic boundaries.

3.2. Rock strength

The Schmidt hammer, a portable tool that rapidly measures the energy rebounded by a rock when impacted, was used to measure the in situ rebound values of bedrock outcrops exposed in the valley walls. Rebound measurements were taken on both the north and south valley walls for discrete analyses. The number of measurements collected varied by reach, based on reach lengths and amount of exposed bedrock, but ranged from 50 to 179 measurements per valley side (Table 1). Measurements were spaced by at least 10 cm. The Schmidt hammer was always oriented normal to the surface being measured. The lowest elevation outcrops were measured, ranging from river level to 100 m above the valley floor. Because a purely objective grid method of taking measurements was impractical in the field where bedrock exposure was sporadic, measurements were obtained to represent the variability of rock characteristics in each reach. For example, outcrops with a uniform distribution of joints or fractures and weathering were sampled uniformly with measurements taken on weathered surfaces, fresh surfaces, between fractures, and directly on or near fractures. This approach provides a measure of

Table 1
Summary of Schmidt hammer rebound values and valley morphometry measurements by reach.

Reach	Lithology	Schmidt hammer rebound						Valley floor width (m)	Hillslope gradient (degrees)		Stream gradient (%)
		North side			South side						
		Mean	S.D.	n	Mean	S.D.	n		North	South	
1	Neoproterozoic diorite/syenite	38.04	16.68	55	46.57	12.19	84	89.83	33.85	26.36	1.2
2	Neoproterozoic diorite/syenite	50.38	10.44	50	51.07	9.82	106	21.15	38.17	34.21	1.1
3	Neoproterozoic diorite/syenite	42.07	10.98	60	41.22	13.89	78	95.86	36.17	29.87	0.9
4	Neoproterozoic diorite/syenite	39.58	13.95	55	41.76	8.97	55	63.56	34.94	30.41	0.9
5	Eocene Challis volcanics	52.74	9.91	89	55.75	10.56	83	25.84	31.95	33.64	0.9
6	Eocene Challis volcanics	34.61	13.05	84	43.90	9.62	179	208.53	33.33	32.69	0.7
7	Eocene Challis volcanics	47.64	11.31	135	44.06	8.53	84	22.30	38.7	32.43	1.1
8	Eocene Challis volcanics	29.77	12.20	110	48.52	8.84	171	412.45	26.79	33.90	0.5
9	Neoproterozoic diorite/syenite	44.66	13.06	50	42.73	11.50	129	103.24	31.04	38.91	0.5
10	Mesoproterozoic quartzite/siltite	43.32	13.71	69	46.11	11.82	177	109.08	33.86	29.14	0.7
11	Eocene granodiorite	53.58	10.21	120	N/A	N/A		18.06	45.66	37.65	1.4
12	Eocene granodiorite	49.59	13.71	116	N/A	N/A		22.24	40.18	40.04	1.3
13	Eocene granodiorite	48.24	12.22	85	53.11	10.02	85	20.12	40.79	34.46	2.0

the variability of rebound and, thus, an overall rebound value that incorporates not only intact rock strength but also irregularities such as weathering and fracture density (see Appendix A).

While hardness decreases with length of exposure and extent of weathering at the surface, we don't think this has caused significant differential weathering at Big Creek. Big Creek is a relatively rapidly eroding valley and chemical weathering rates are probably quite low because of the semi-arid climate. Based on our observations, we assume that any weathering that has occurred at the base of the valley walls where measurements were taken has been uniform. Most of our measurements were taken on outcrops very near the active channel which have been relatively recently exposed.

In addition, we don't find any evidence for systematic differential erosion in Big Creek valley that might expose fresher bedrock and thus bias our strength measurements. The valley cross section is symmetric in most locations and there is no known tilting pushing the channel either north or south. Well-preserved fluvial terraces above the active channel are the same elevation on both sides of the valley, suggesting that no north-south tilting has occurred. As for non-tectonically driven shifts in the channel, we assume that the channel will meander and apply uniform erosive force to the valley walls within the time scale of valley widening.

3.3. Valley morphometry

3.3.1. Valley floor width

Valley floor width is defined as the width of the flat valley bottom between valley walls. This includes the stream channel, floodplain, and any fluvial fill terraces up to 4 m above the floodplain. In Big Creek canyon, the valley floor is marked by a distinct break in slope between the valley walls and valley floor. The width of the valley floor in Big Creek varies from just the width of the channel to a broad alluvial floodplain.

The valley floor width was measured with an infrared laser rangefinder. Six to ten transects were measured across the valley floor perpendicular to the valley axis at each reach. The transect distances for each reach were averaged together to obtain an average valley floor width for each reach.

3.3.2. Stream gradient

Stream gradients for each reach were calculated from USGS 7.5′ topographic maps with contour intervals of 40 and 80 feet. Gradients were also calculated from 10-m DEMs, but errors inherent in those DEMs rendered the gradient calculations less useful.

3.3.3. Hillslope gradient

We determined the hillslope gradients of the valley walls bounding the valley. Only valley walls (i.e., facets) whose strike parallels the valley axis were used. Valley walls of tributaries and gullies were not included in the measurement. The slopes of the valley-parallel facets were measured from stream level to ~250–300 m above the valley floor. This elevation corresponds roughly to the tops of most facets and is a break in slope representing the initiation of recent rapid incision (e.g., Meyer and Leidecker, 1999). Slopes higher than this threshold elevation are too distant from the Schmidt hammer measurements to be considered valid.

Hillslope gradient was measured separately for the north and south side of each reach. Measurements were made from 7.5′ USGS SDTS DEMs with a resolution of 10 m in ArcMap 9.0 software (Environmental Systems Research Institute, Inc., 2004). Slope was calculated over the area of each hillslope facet.

3.4. Data analysis

Two sets of statistical analyses were used to test the hypotheses that (i) rock strength differs between valley sides and among lithologic groups, and (ii) valley morphometry is related to rock strength. The first hypothesis was tested using a two-factor, fixed-effects Analysis of Variance (ANOVA) in which the factors were valley side (north and south) and lithologic group (diorite/syenite, Challis volcanics, quartzite/siltite, granodiorite, as described above). We tested for main effects and interaction between valley side and lithologic group. Residuals departed only slightly from normality and had constant variance across all valley side/lithologic groups. Thus no data transformation was necessary. We tested for differences in mean Schmidt hammer rebound (i) between valley sides, (ii) among the four lithologic groups, and iii) among each valley side/lithology combination. To assess differences between pairs of factor levels (e.g., between northside granodiorite and southside granodiorite), we used multiple comparison tests with Bonferonni's correction (see, e.g., Zar, 1974; Ramsey and Schafer, 2002). The number of Schmidt hammer measurements varied across the sampled locations (Table 1), but this imbalance does not affect the outcome of the hypotheses we tested with the ANOVA (Hocking, 1996).

We tested the second hypothesis with simple linear regression analysis, using reaches as the sampling units. For each reach, the values of the independent and dependent variables were taken to be reach means. We performed linear regression of (i) valley width against Schmidt hammer rebound on the weakest side of the valley, (ii) northside hillslope gradient against Schmidt hammer rebound on the north side of the valley, (iii) southside hillslope gradient against Schmidt hammer rebound on the south side of the valley, and (iv) stream gradient against Schmidt hammer rebound on the weakest side of the valley. Valley width and stream gradient were approximately log-

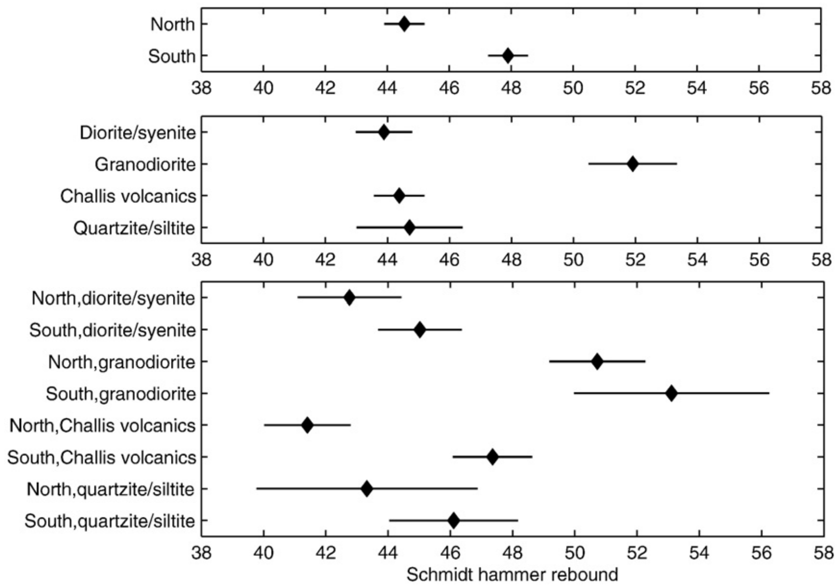

Fig. 3. Results of multiple comparisons from two-factor ANOVA. Markers are group means of Schmidt hammer rebound; bars indicate simultaneous Bonferroni confidence intervals around means for testing null hypotheses that group means are equal. Nonoverlap of bars indicates significant difference in means at $\alpha = 0.05$.

normally distributed, so we used the natural logarithm of these variables in the analysis (i.e., we tested for an exponential relationship between these variables, respectively, and Schmidt hammer rebound). Hillslope gradients were normally distributed, so the analyses were performed on the raw data. All hypothesis tests were performed at the $\alpha = 0.05$ significance level.

4. Results

4.1. Dependence of rock strength on valley side and lithology

The ANOVA showed that Schmidt hammer rebound varied significantly between north and south sides ($F_{1,2301} = 25.57$, $P < 0.001$) and among the lithologic groups ($F_{3,2301} = 30.04$, $P = 0.008$). Rebound was significantly higher on the south side of the valley and significantly higher for granodiorite than for any of the other lithologic groups (Fig. 3). Mean rebound did not differ significantly among the other three lithologic groups. The interaction of valley side and lithologic group was also significant ($F_{3,2301}$, $P = 0.0142$). Mean rebound was higher on the south side of the valley within each of the four lithologic groups, but this difference was significant only for the Challis volcanics (Fig. 3). At the reach level, which we consider more relevant than that of the lithologic map unit level, mean rebound was higher on the south side for all but reaches 3 and 7 (Table 1). The greatest differences between north- and southside rebound values occurred in reaches 6 and 8, both within the Challis volcanics.

4.2. Relationship between valley morphometry and rock strength

The regression analysis showed a significant, decreasing exponential relationship between valley floor width and weakest-side Schmidt hammer rebound ($F_{1,9} = 31.0$, $P < 0.001$ $r^2 = 0.795$; Fig. 4). Hillslope gradient on the north side of the valley was significantly and positively related to Schmidt hammer rebound ($F_{1,11} = 10.6$, $P = 0.008$, $r^2 = 0.490$; Fig. 5), but hillslope gradient showed no significant relationship to Schmidt hammer rebound on the south side ($F_{1,9} = 0.445$, $P = 0.522$, $r^2 = 0.047$). Stream gradient was not significantly related to weak-side Schmidt hammer rebound ($F_{1,9} = 3.43$, $P = 0.097$, $r^2 = 0.276$).

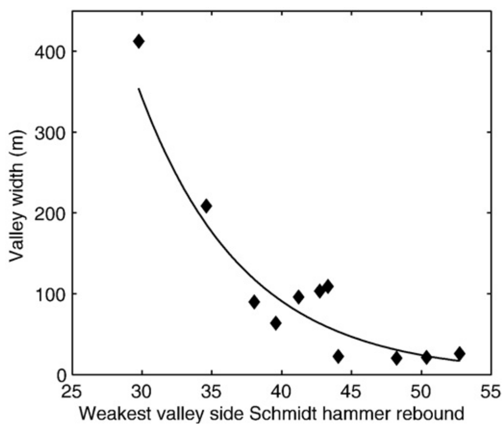

Fig. 4. Valley floor width as a function of Schmidt hammer rebound on the weakest side of the valley. Least-squares exponential function has equation $y = 18413e^{-01328x}$, and fit is significant ($P < 0.001$, $r^2 = 0.795$).

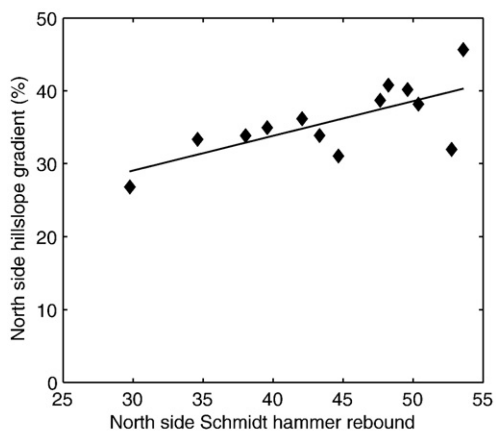

Fig. 5. Northside hillslope gradient as a function of Schmidt hammer rebound. Least-squares line has equation $y = 0.477x + 14.7$, and fit is significant ($P = 0.008$, $r^2 = 0.490$).

5. Discussion

5.1. Bedrock strength

5.1.1. Aspect variability

Statistical analysis of the Schmidt hammer rebound values demonstrates that a significant difference exists between the rebound values of the north and south sides of Big Creek. The north side of the valley (south-facing) has lower Schmidt hammer rebound values than does the south side of the valley (north-facing). Two processes might explain this phenomenon: (i) during the summer, diurnal heating and cooling of rocks on south-facing slopes by solar radiation creates strong thermal gradients and enough stress to cause fracturing (McFadden et al., 2005); and (ii) during the winter, the south-facing slopes receive more solar radiation and therefore experience more freeze–thaw cycles than do the shaded north-facing slopes. Thermal cracking is plausible, but it has only been tested in an arid desert environment (McFadden et al., 2005) so it is unknown if it could apply to a location in central Idaho, which is cooler, wetter, and at a higher latitude than the arid desert. North-facing slopes typically have more vegetative cover than south-facing slopes in the Big Creek valley, thus providing protective shade.

Burnett et al. (2002) observed the opposite of this rock strength phenomenon in the Colorado Plateau region. Their Schmidt hammer data showed that bedrock rebound values on the south side of the valley (north-facing slopes) were lower than those on the north side of the valley. They suggested that increased soil moisture on north-facing slopes (caused by greater vegetative cover) increased the rate of weathering through clay hydration and expansion. Biologically enhanced water retention and increased weathering of bedrock is certainly plausible at Big Creek, where aspect strongly influences vegetation distribution; but the conclusions of McFadden et al. (2005) suggested that south-facing slopes should have lower rock strength because of greater exposure to solar radiation. At Big Creek, where mean minimum temperatures are below freezing seven months out of the year (Taylor Ranch weather station; Western Regional Climate Center, 2005), the high frequency of freeze–thaw cycles is the most likely explanation for the decreased rock strength on the north side of the valley.

5.1.2. Lithology

Statistical analysis of Schmidt hammer rebound data demonstrates that the granodiorite lithologic group has significantly higher rebound values than do the other three lithologic groups, but the Schmidt hammer rebound values of those other lithologic groups (Mesoproterozoic quartzites and siltites, Neoproterozoic mafic intrusions, and Eocene volcanic tuffs and porphyry dikes of the Challis Volcanic Group) are not statistically distinct from one another. These results demonstrate that lithology alone may not be diagnostic of rock strength parameters or of valley morphometry. Degree of weathering and the presence of fractures and joints must be taken into account when determining rock strength.

5.2. Valley floor width

The clearest statistical relationship between rock strength and valley morphometry links Schmidt hammer rebound and valley floor width. The weaker average Schmidt hammer rebound value (south vs. north side) is strongly negatively correlated to the width of the valley floor. We found no relationship between rebound value on the strongest side of the valley and valley floor width. Thus, the weak side of the valley appears to control the width of the valley floor of Big Creek.

Both vertical and lateral erosion occur in a channel, and in both cases rock is removed (by abrasion and/or plucking) at the base of the adjacent hillslopes. Removal of rock oversteepens these lower slopes and reduces their stability. Rock masses with no joints or weathering typically fail along shear planes when the force of gravity acting on a rock mass overcomes the strength of that rock mass. The maximum height of a vertical cliff of unjointed rock is approximately equal to the uniaxial compressive strength divided by the unit weight of rock (Terzaghi, 1962). At high compressive strengths, this vertical height limit is ~1500 m (Selby, 1980). Joints, fractures, or faults occur in almost all rock masses and act as planes of weakness or failure, so the expected maximum cliff height is rarely achieved. Oversteepened slopes are susceptible to erosion, and hillslope mass wasting processes will move material from the oversteepened slope into the channel through shear failures, topples, falls, or slides. Assuming that the river has enough transport capacity, it will remove the material and leave the valley floor wider.

Our data show a strong dependence of valley floor width on rock strength on the weaker side of the valley. This relationship suggests that bedrock with high strength is resistant to lateral fluvial erosion and can hold an oversteepened slope, preventing further widening, when lateral fluvial erosion does occur. Conversely, bedrock with low strength is less resistant to lateral fluvial erosion and easily fails when oversteepened, thus facilitating valley floor widening. Furthermore, widening of the valley floor reduces stream power, which initiates lateral migration of the channel and may be a feedback for continued widening.

5.3. Stream gradient

Reach-scale stream gradient is not significantly dependent on rock strength in Big Creek. However, Mackley and Pederson (2004) found a strong correlation between these two parameters on the Colorado River in Grand Canyon and Glen Canyon. J.L. Pederson (personal communication, Utah State University, 2005) has also observed a relationship between bedrock strength and stream gradient on other rivers within the Colorado Plateau. As a generalization, one would expect higher bedrock strength to correspond to steeper stream gradients in equilibrated streams (Gilbert, 1877; Powell, 1895). The lack of correlation at Big Creek is probably because of a combination of the small scale of spatial observations, an abundance of late Pleistocene sediment cover (Pierce and Scott, 1982), and a longitudinal profile that is not in equilibrium.

5.4. Hillslope gradient

Data from this study show that hillslope gradient is moderately dependent on bedrock rebound values. While steep slopes and exposed bedrock indicate that Big Creek is an actively incising river, many slopes at Big Creek are partially covered with talus and colluvium, likely dating to late Pleistocene or Holocene time. Talus tends to form straight slopes that cover bedrock and can mask the bedrock gradient of the hillslope. Given enough time to transport the accumulated hillslope debris, the basin would return to interglacial conditions, more bedrock would be exposed, and the true hillslope gradient would be apparent.

The positive moderate dependence of hillslope gradient on rock strength on the north side of the valley provides evidence that differential weathering across lithologies determines the gradient that can be maintained as lateral migration of the stream erodes valley walls. A given lithology can be strong or weak depending on the degree of weathering and the presence of joints. Stronger bedrock on the south side of the valley is able to maintain oversteepened slopes, while mass wasting occurs more readily on the weaker north side. Thus, because of the stronger mass wasting response, we are more likely to see hillslope gradient reflecting rock strength. The north side, which is generally the weaker and more weathered side of the valley, was the only side where this moderate dependence was found.

6. Conclusions

This study analyzed bedrock strength, as measured by the Schmidt hammer, and compared those Schmidt hammer measurements to three parameters of valley morphometry: valley floor width, stream gradient, and hillslope gradient. The conclusions of this study are (i) a statistically significant difference was found between north- and southside rebound values, with the north side being lower; (ii) the rebound values of the lithologic groups are statistically indistinguishable from one another, except for the granodiorite group, which has significantly higher rebound values; (iii) valley floor width is strongly dependent on the bedrock strength of the weaker side of the valley; (iv) a moderate correlation exists between bedrock strength and hillslope gradient; and (v) no correlation exists between bedrock strength and stream gradient.

Analysis of the longitudinal profile reveals two major controls on the large-scale gradient: (i) a NE-trending normal fault and (ii) rapid incision of the Middle Fork Salmon River and the resulting knickpoint. Stream and hillslope gradient may be partially obscured by an abundance of late Pleistocene sediment. Our model for valley floor width formation suggests that lateral erosion of the river (through abrasion and/or plucking) oversteepens the lower hillslopes, which respond by mass wasting at a rate determined by rock strength (lower strength yields more mass wasting and lower hillslope gradients). When the resulting rock debris is removed, the valley floor is wider.

Rivers are complex systems in which many factors work to control the morphometry. This study has isolated one such factor: bedrock strength as measured by Schmidt hammer rebound, which has a strong control on valley floor width. In mountain drainage basins with diverse lithology, Schmidt hammer rebound is a useful parameter for describing variations in valley floor width. The results of this study may be useful for morphometric classification schemes and could potentially provide a link between the fields of geology, ecology, and geomorphology. Our methods could be applied in other small mountain drainage basins with different lithology, climate, or tectonic history to determine if the same relationships exist between rock strength and valley morphometry.

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Appendix A. Sampling methodology calibration tests

We performed a set of analyses to test the ability of our Schmidt hammer rebound sampling method to reflect the bulk strength properties of the four general classes of bedrock measured in situ in this study. The assumption underlying our use of the Schmidt hammer to measure bedrock strength is that the bulk strength of the bedrock unit depends on a combination of inherent strength of the lithologic type and the degree of fracturing, with higher fracture density resulting in lower bulk strength. We hypothesized that within a bedrock unit, Schmidt hammer rebound measured at a given point would be lower at points along block joints and higher at points in the center of the blocks. Further, we hypothesized that regardless of where the measurement is taken, rebound would decrease with increased joint density (decreased spacing between joints). Finally, we hypothesized that we could measure the bulk strength through an “integrated” sampling method that distributes the location of rebound measurements randomly across the bedrock unit. In theory, the randomness should result in weighted averaging of rebound measurements across joints and block centers, with the weights determined inherently by the probability of an individual measurement point falling on a joint, in the center of a block, or somewhere in between. This weighted average should then represent a measure of the bulk strength, which incorporates joint density, weakening along the joints themselves, and strength of the blocks defined by these joints.

We performed the tests in the field study area on representative exposures of each of the four lithologic types encountered in the in situ study: diorite, granodiorite, quartzite, and welded tuff (referred to simply as “tuff”). We defined three levels of measurement location: “joint” (measurement taken on a joint), “center” (measurement taken in the center of a fracture block), and “integrated” (a set of measurements distributed randomly across the bedrock unit). We then defined five levels of fracture density, as measured (inversely) by mean spacing between joints: 0.0625 m, 0.125 m, 0.25 m, 0.5 m, and >1 m. Mean joint spacing obviously did not equal these idealized values, so we considered joint spacing as a discrete variable with five fixed levels given by these nominal values and sampled in locations where mean joint spacing was approximately equal to the nominal levels. For each rock type we measured Schmidt hammer rebound at 15 points in each location × spacing combination involving joints and centers and spacing levels of 0.0625 m, 0.0125 m, 0.25 m, and 0.5 m (we considered joint spacing >1 m to represent intact rock and so did not consider any locations at this spacing to be either “center” or “joint”). Under the assumption that the integrated sampling method would require more points in order to accomplish the random averaging, we recorded rebound at 30 randomly distributed locations at each of the five joint spacing levels on each rock type. Because of logistical constraints, somewhat fewer measurements were made on granodiorite, but this slight loss of balance in the sampling design neither affects the hypotheses tested nor the distribution of the test statistics used (Hocking, 1996).

We performed two sets of hypothesis tests using ANOVA. The first involved all measurements except those recorded on the intact (spacing >1 m) rock units. We tested the null hypotheses that mean rebound did not depend on rock type, measurement location, joint spacing, or any of the three pair-wise interactions of these variables. We also tested the null hypothesis that the mean of the integrated measurements did not differ from the grand mean taken across all three measurement locations. Because we made the same number of integrated measurements as we made on joints and in block centers combined, this grand mean value was not skewed by sample sizes. Rejection of the null hypothesis that mean rebound does not depend on measurement location—and failure to reject the null hypothesis that mean integrated rebound does not differ from the grand mean—together indicate that the integrated sampling method does indeed average the effect of fracturing in the intended way and therefore provides a measure of bulk strength. As this turned out to be the case, we then performed a second ANOVA on only those measurements

Table A1
Analysis of variance table for tests involving effect of measurement location.

Source	*df*	Sum of squares	Mean squares	*F*	*P*
Rock type	3	58,695	19,565	215.5	0
Joint spacing	3	32,183	10,728	118.2	0
Location	2	6416	3208	35.3	0
Rock type × joint spacing	9	8129	903	9.9	0
Rock type × location	6	843	140	1.5	0.159
Spacing × location	6	625	104	1.1	0.333
Error	890	80,806	90.8		
Total	919	199,036			

Reported sums of squares are Type III (adjusted) sums of squares and hence do not add to total sum of squares.

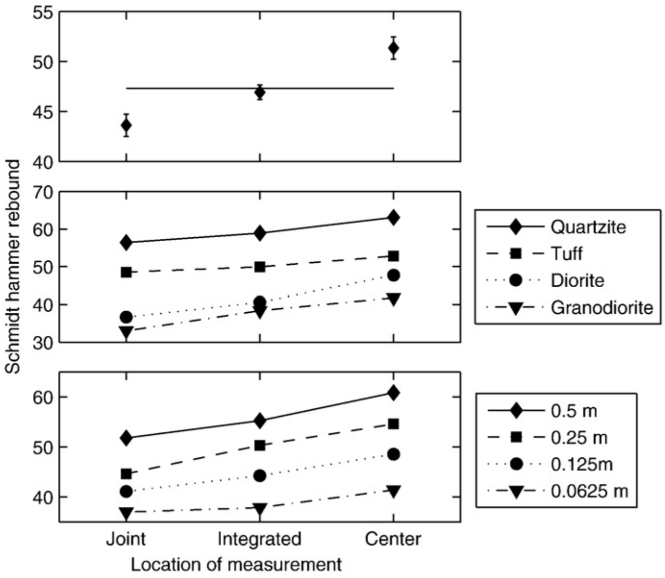

Fig. A1.

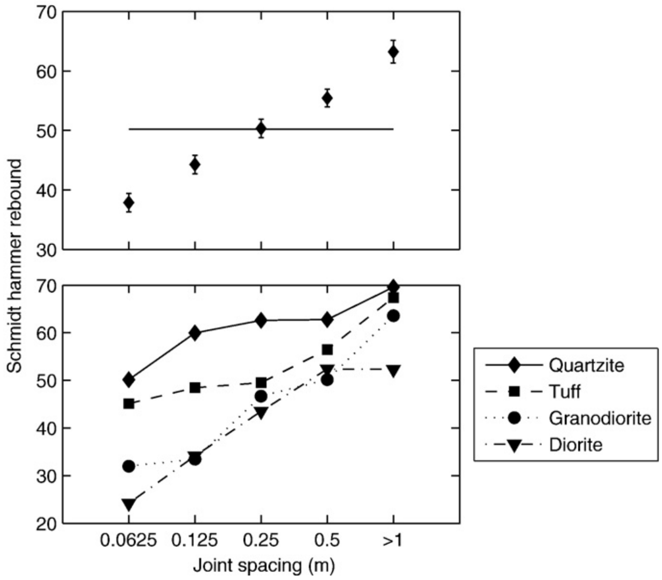

Fig. A2.

collected with the integrated sampling method, which include those recorded on the intact rock units (>1 m joint spacing). We used this ANOVA to test the null hypotheses that rebound did not depend on rock type, joint spacing or the interaction of these two effects. All hypothesis tests were performed at the 5% significance level.

The first ANOVA showed that rebound varied significantly across rock type, joint spacing and measurement location (Table A1). Dependence on joint spacing differed among rock types. As hypothesized, mean rebound was lowest when measured on joints, highest when measured in block centers, and intermediate in value when measured with the integrated method (Fig. A1). When averaged across all rock types and joint spacing values, mean rebound at each location level differed significantly from mean rebound at each of the other two location levels. Furthermore, mean rebound measured with the integrated method did not differ significantly from the grand mean. The pattern of increasing rebound across the three levels of measurement location did not differ among rock types or joint spacing levels (Table A1; Fig. A1). These results confirm our hypothesis that the integrated method averages the effects of fracturing across the rock unit. Furthermore, because neither interaction effect involving location was significant, the integrated method gives results that are consistent across rock types and joint spacing values.

The second ANOVA again showed that rebound varied significantly across rock type and joint spacing and that the rock type × joint spacing interaction was significant (Table A1). As expected, rebound was an increasing function of joint spacing (Fig. A2), and the roughly linear increase in mean rebound with the nominal joint spacing levels we chose suggests that when averaged across all rock types, rebound is roughly a logarithmic function of joint spacing. However, when the interaction of rock type with joint spacing is considered, it is apparent that the rate of increase in rebound with joint spacing is greater for rock types with lower inherent strength than for those with higher strength.

Based on these results, we conclude that the integrated method of sampling we used in the in situ study effectively reflects the bulk strength of bedrock units and provides consistent measurement of bulk strength across different rock types and fracture densities.

Table A2
Analysis of variance table for values measured with the integrated sampling method.

Source	*df*	Sum of squares	Mean squares	*F*	*P*
Rock type	3	34,514	11.505	132.4	0
Joint spacing	4	39,319	9830	113.13	0
Rock type × joint spacing	12	6938	578	6.65	0
Error	575	49,962	86.9		
Total	594	134,643			

Reported sums of squares are Type III (adjusted) sums of squares and hence do not add to total sum of squares.

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