Bedrock channel adjustment to tectonic forcing: Implications for predicting river incision rates

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ABSTRACT

We present detailed data of channel morphology for a river undergoing a transient response to active normal faulting where excellent constraints exist on spatial and temporal variations in fault slip rates. We show that traditional hydraulic scaling laws break down in this situation, and that channel widths become decoupled from drainage area upstream of the fault. Unit stream powers are ~4 times higher than those predicted by current scaling paradigms and imply that incision rates for rivers responding to active tectonics may be significantly higher than those heretofore modeled. The loss of hydraulic scaling cannot be explained by increasing channel roughness and is an intrinsic response to tectonic forcing. We show that channel aspect ratio is a strongly nonlinear function of local slope and demonstrate that fault-induced adjustment of channel geometries has reset hillslope gradients. The results give new insight into how rivers maintain their course in the face of tectonic uplift and illustrate the first-order control the fluvial system exerts on the locus and magnitude of sediment supply to basins.

Keywords: rivers, tectonics, faulting, channel geometry, geomorphology.

INTRODUCTION

Predicting landscape response to tectonic forcing in mountainous catchments requires a full understanding of fluvial incision processes and rates in bedrock rivers (Whipple and Tucker, 1999). Furthermore, to infer the presence of features such as active faults from channel characteristics where direct structural or geodetic data are unavailable (cf. Kirby et al., 2003) requires prediction of channel adjustment to tectonically generated changes in channel boundary conditions. Stream incision rate, ε, is commonly understood to depend strongly on bed shear stress, τ, which scales with unit stream power, ω (i.e., $\varepsilon \sim \omega \sim [\tau V]^a$, where V is flow velocity and a is a process-dependent constant which is ~1). The variables τ and V depend on both gradient and channel cross-sectional geometry (Howard et al., 1994; Whipple and Tucker, 1999). Consequently, changes in channel width, W, and depth, H, are integral, but largely neglected, components of stream response to tectonic forcing (Finnegan et al., 2005; Wobus et al., 2006). Because channel dimensions are difficult to measure from digital elevation models (DEMs) and require time-consuming field measurement, erosion laws are normally combined with hydraulic scaling relationships that define downstream river morphology: Channel geometry is expressed using power-law functions of drainage area, A (as a proxy for discharge, Q) (Leopold and Maddock, 1953), under the key assumption that river banks adjust to a dominant channel-forming flow:

$$W = K_1 A^b \quad (1)$$

and

$$H = K_2 A^c. \quad (2)$$

In tectonically quiescent areas with uniform lithology, bedrock rivers exhibit scaling relationships with exponents comparable to alluvial rivers, i.e., $b \sim 0.5$; $c \sim 0.35$ (Montgomery and Gran, 2001). Consequently, these values have been widely adopted for landscape modeling, incorporating the implicit assumption that slope, S, is the main variable that responds to tectonic forcing (Whipple and Tucker, 2002).

However, adjustment of width and/or depth is a key mechanism by which rivers respond to changing boundary conditions (e.g., Stark, 2006). Thus, any predictive model that fixes $W \propto A^{0.5}$ (Eq. 1) and allows only variations in S to drive incision is unlikely to capture the true response of fluvial systems to disequilibrium conditions. Field studies have already suggested that channel geometry can be highly variable in tectonically perturbed landscapes (e.g., Harbor, 1998; Lavé and Avouac 2001; Duvall et al., 2004). In these examples, uplift-driven changes in stream gradient have led to adjustments in channel geometry, and, hence, the distribution of shear stress and incision. To successfully model these adjustments, we require case studies in which the natures of both the forcing and the response are temporally well constrained. We address this challenge by presenting new field data that document downstream changes in hydraulic geometry for a river crossing an active normal fault where excellent constraints exist on the history of fault movement.

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STUDY AREA AND METHODS

The central Italian Apennines contains a network of active normal faults that were initiated in mid-Pliocene times (Lavecchia et al., 1994; Fig. DR1A in the GSA Data Repository[1]). Fault throw rates over the last 3 m.y. are well constrained through structural mapping, biostratigraphy, tephrachronology, paleoseismic studies, and displacement measurements across postglacial fault scarps (Roberts and Michetti, 2004, and references therein). Mesozoic limestones are uplifted in the footwalls of these faults, while the hanging walls are underlain by Miocene flysch and partly covered by Pleistocene fill (Fig. DR1B, see footnote 1). We focus on the northern part of the network, in particular the Fiamignano fault (Fig. 1).

The ~25-km-long Fiamignano fault strikes NW-SE and downthrows to the SW; total throw and throw rate at the center are ~1800 m and ~1 mm/yr, respectively, and both decrease systematically toward the fault tips (Roberts and Michetti, 2004). The fault increased its slip rate from 0.3 to 1 mm/yr at ca. 1 Ma as a result of interaction with adjacent faults in the array (Cowie and Roberts, 2001). To document the fluvial response to this acceleration, field measurements were made of high flow channel width, W_b, depth, H, valley width, W_v, and local channel slope, S, for the Rio Torto, a 65 km^2 perennial river catchment that crosses the fault near its center (Fig. 1). Selby rock strength (Selby, 1980) and grain-size data were collected to assess whether variations in lithological resistance or surface roughness also play a role in controlling fluvial incision (Table DR1; see footnote 1). These results for the Rio Torto are compared with the Fosso Tascino, a 45 km^2 catchment that crosses the Leonessa fault, which lies 20 km N of Fiamignano (Fig. DR1B), strikes NW-SE, and has had a constant slip rate of 0.35mm/yr for 3 m.y. (Roberts and Michetti, 2004).

RESULTS

Channels crossing the Fiamignano fault do not display typical concave-up profiles (Fig. 2). While the upper parts of the Rio Torto and its tributary, the Vallone Stretta, are characterized by wide valleys and meandering, partly alluviated channels, the lower parts of the river form a deep gorge that incises directly through bedrock, with little sediment cover on the bed. The gorge contains a prominent convex reach directly upstream of the fault (Fig. 2) that has a vertical height of ~400 m. Downstream of the fault, the river shallows and alluviates. The prominent slope break on the Rio Torto at 6 km does not coincide with any discernible change in lithology or rock mass strength (Fig. 2), so this alone cannot explain the convexity in long profile above the fault. Convex river profiles are predicted to develop in response to changes in uplift rate for both detachment-limited and sediment flux–influenced erosion systems (but importantly, not for purely transport limited channels; Whipple and Tucker, 2002). Therefore, the Rio Torto is a strong candidate to be interpreted as undergoing a transient response to the increase in fault slip rate that occurred ca. 1 Ma.

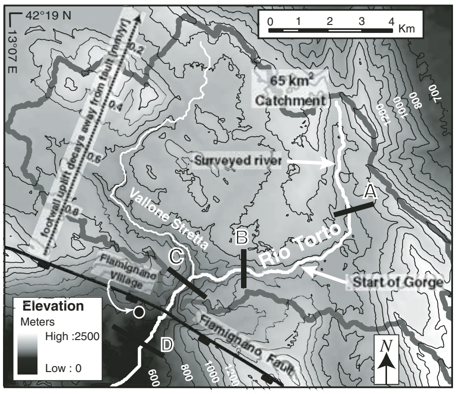

Figure 1. Fiamignano fault, Rio Torto, and Vallone Stretta tributary; contours are every 100 m; lines A–D refer to valley profiles in Figure 4B.

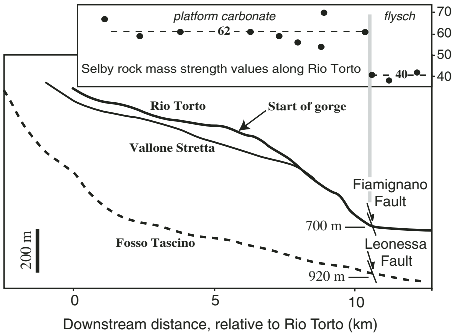

Figure 2. Channel long profiles for Rio Torto, Vallone Stretta (line), and Fosso Tascino (dashed); inset shows Selby rock strength for Rio Torto.

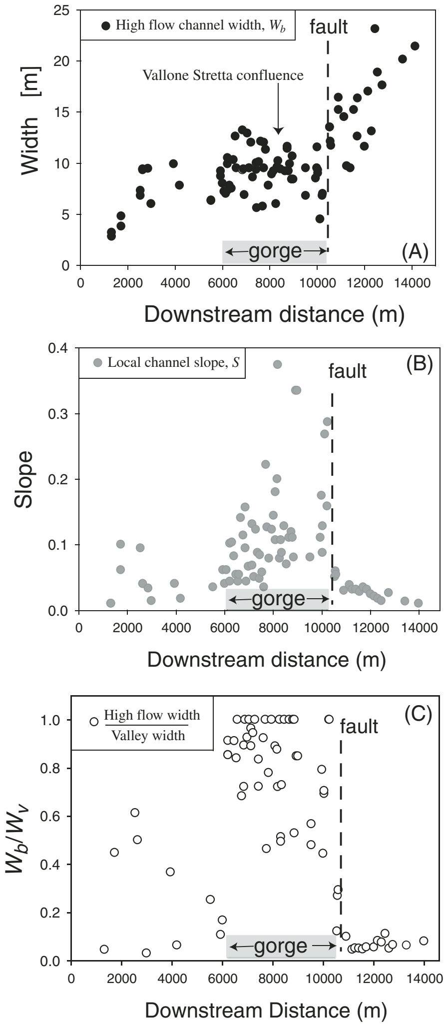

Figure 3. A: High flow channel width, W_b. B: Local channel slope, S. C: Ratio of W_b to valley width, W_v, against downstream distance in the Rio Torto.

The river shows systematic variations in hydraulic geometry downstream (Fig. 3). High flow width (Fig. 3A) increases to ~9 m in the first 3 km downstream, but it then remains essentially constant within the gorge, despite a considerable increase in A at ~8.5 km downstream (Fig. 1). Local channel slopes (Fig. 3B) are generally <0.05 (~3°) both upstream of the break in slope and downstream of the fault. In contrast, S in the gorge is >0.05, and some reaches exceed 0.3 (~17°); furthermore, the minimum slopes increase downstream in the gorge, indicating that the entire channel has steepened in the zone of maximum uplift near the fault. W_b/W_v (Fig. 3C) is close to 1 throughout the convex reach, indicating that long-term erosion in the proximal footwall is concentrated in a narrow zone, typically <10 m wide. In contrast, this ratio is low in the headwaters and downstream of the fault.

These significant variations are seen clearly in the channel aspect ratio, W_b/H (Fig. 4A). W_b/H is a strongly nonlinear function of slope, in contrast to assumptions and predictions of recent models (Finnegan et al., 2005; Wobus et al., 2006). Steep slopes, >0.1, are associated with $W_b/H < 6$, which is characteristic of narrow, deep channels in the gorge. Low slopes, <0.05 are associated with wider, shallower channels, in the headwaters and beyond the fault. For the low-gradient channels, W_b changes significantly on the reach scale, whereas W_b/H is locked into a narrower range where slope is high. The relationship between aspect ratio and slope can be empirically fitted with a power law, giving $W_b/H = 2.6S^{-0.34}$. Most of the variation in aspect ratio relates to ~×5 vari-

[1]GSA Data Repository item 2007030, study location and geology, methods, and shields stress, is available online at www.geosociety.org/pubs/ft2007.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

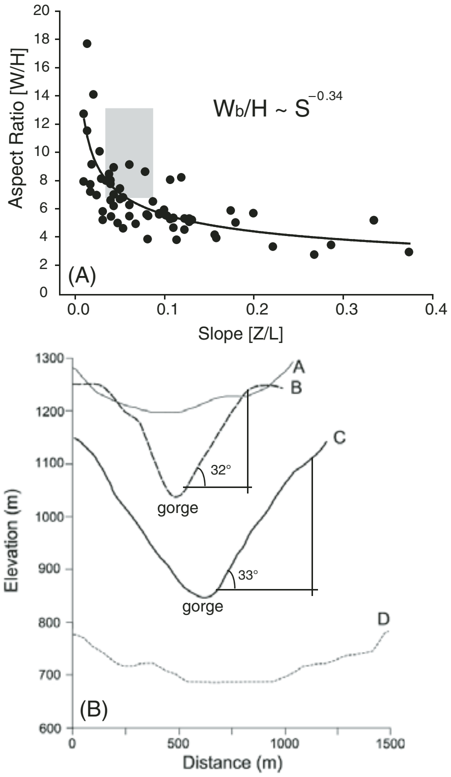

Figure 4. A: Aspect ratio, W_b/H against S for Rio Torto. Gray box shows typical values of W_b/H for Fosso Tascino, Leonessa. B: Profiles of valley form at A–D (Fig. 1C). Z—elevation; L—downstream distance.

ation in W_b, rather than H, which suggests that much of the channel response lies in narrowing rather than deepening.

IMPLICATIONS

The strong dependency of channel planform and slope on position relative to the fault has important implications for the coupling between fluvial incision and hillslope processes (Fig. 4B). Comparison of a valley cross section through A, in the headwaters (Fig. 1), with B, <500 m downstream of the break in channel slope, reveals marked differences over distances <2 km: While maximum elevations are similar at A and B, the hillslopes at B are much steeper. The gradient (32°) and planar form of the valley sides suggest that slopes have reached the threshold angle for stability. Comparison of cross-sections A and B also suggests excavation of the material contained within the valley walls. Downstream of the major confluence, C shows similar steep (~33°) hillslope angles, whereas D shows that the valley widens again immediately after the river emerges onto the hanging wall of the fault (Fig. 4B). These cross sections imply that the hillslopes upstream of the fault have responded on a similar time scale to the channel adjustment process itself, and that landslide debris enters directly into the channel where the

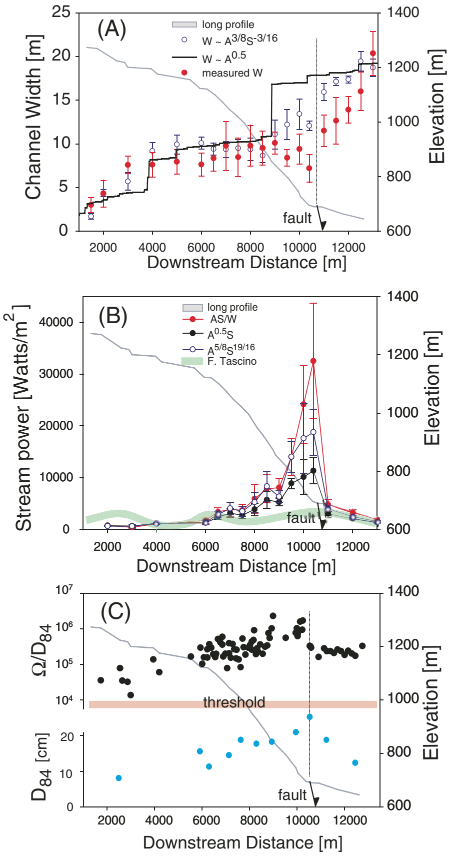

Figure 5. A: Channel widths predicted by (1) $W_b \sim A^{0.5}$ (line), (2) $W_b \sim A^{3/8}S^{-3/16}$ (blue), and (3) measured values (red). B: Unit stream power predicted from hydraulic scaling, $A^{0.5}S$ (black), Finnegan width model, $A^{5/8}S^{19/16}$ (blue), and measured channel widths (red). Green swath shows unit stream power for Fosso Tascino, scaled to fault position. Data are binned over 500 m intervals downstream; error bars show 2σ variation for each interval. A values were obtained from 20 m digital elevation model. To calculate unit stream power, we use Q = 100 m³/s at the fault, consistent with estimates derived from applying Manning's equation to channel cross-sections and assume $Q \sim A$. Using a different value of Q affects magnitude of ω, but not differences between predicted and measured values. C: D_{84} (blue circles) and Ω/D_{84} (black) against L. Red box shows empirical threshold for onset of good scaling.

gorge has developed. Any propagation of the steep river reach upstream will lead to enhanced erosion and hillslope rejuvenation upstream of B. This illustrates the first-order control that the river system has on transmitting tectonic signals to the landscape, ultimately determining sediment supply to hanging-wall basins.

These data (Fig. 3) imply that existing empirical scaling relations (Eqs. 1 and 2) lose their predictive power under transient conditions. We demonstrate this by comparing (Fig. 5A) measured widths along the Rio Torto with predicted widths using $W_b \sim A^{0.5}$ and also Finnegan et al.'s (2005) modified form, $W_b \sim A^{0.38}S^{-0.19}$, which permits channels to narrow in regions of high slope but assumes topographic steady state and constant W_b/H. While it is possible to achieve a reasonable fit between measured widths and predicted values in the upper section of the gorge, *both* predictions significantly overestimate W_b in the final two kilometers upstream of the fault. Indeed, where uplift rates are highest, the real channel is ~3 times narrower than the $W_b \sim A^{0.5}$ paradigm predicts. The Finnegan model performs better, but the discrepancy between measured and predicted values is significant.

Since geometry and discharge determine the erosivity of any river, decoupling W_b and Q (Fig. 5A) has a significant impact on predictions of peak incisive power in the Rio Torto, and, hence, whether it is able to maintain its course in the face of continuing fault uplift. Unit stream power ($\omega = \rho gQS/W_b$) is typically used as a proxy for variations in channel incision rate in tectonically perturbed areas (Harbor, 1998; Finnegan et al., 2005). However, if W_b for perturbed systems cannot be expressed as a simple power-law function of A (Fig. 5A), then it follows that measured stream powers may differ significantly from those derived from existing empirical models (Fig. 5B). Although all curves show that ω increases toward the fault, driven by increasing Q and S, there are significant differences in the size of the response. Predicted and measured ω values are similar (600 < ω < 1300 W/m²) over the first 6 km of the river, but $W \sim A^{0.5}$ predicts an increase in ω by a factor <7 toward the fault (Fig. 5B). In contrast, ω (using measured widths) increases by more than 25 times, giving values >35,000 W/m² near the fault, which means that the river is up to four times more erosive than existing width scaling relationships predict. Finnegan et al.'s (2005) calibration gives stream powers closer to measured values, but still significantly underestimates ω in the lower part of the gorge (Fig. 5B). This analysis addresses only unit stream power variations, but any incision law that makes assumptions about hydraulic scaling is vulnerable to identical problems.

The key issue is to understand the circumstances in which scaling is lost for rivers affected by active tectonics. We therefore compare our results with data from another river that crosses a similar active fault, north of Fiamignano (Fosso Tascino; Fig. DR1B [see footnote 1]). This is a comparably sized catchment, it crosses identical lithologies, has no difference in climate, and it intersects the Leonessa fault, which has had a constant slip rate since its initiation at 3 Ma (Roberts and Michetti, 2004). The river long profile has a concave-up shape (Fig. 2A), and measured channel geometry scales according to Equations 1 and 2 with $b = 0.51 \pm 0.03$, $c = 0.47 \pm 0.04$, and

W_b/H ~10 (gray box, Fig. 4A), consistent with published equilibrium values. When measured widths are used to calculate unit stream powers along the river, we find similar values to the Rio Torto in the upper catchment but no stream power spike as the fault is approached (Fig. 5B), suggesting approximately uniform energy dissipation downstream. The clear differences in long profiles, hydraulic geometry, and stream power distribution in these examples suggest that it is simplistic to interpret the loss of hydraulic scaling in the Rio Torto in terms of the presence of an active fault alone: Slip rates between the two faults differ by a factor ≤3, but there is no scarp preserved where the fault cuts either channel, indicating that incision balances uplift at that point in both cases. However, the two faults do vary in terms of their temporal uplift history; both initiated ca. 3 Ma, but only the Fiamignano fault underwent a slip-rate increase. Consequently, we argue that the breakdown in scaling reflects the finite time scale for channel adjustment to the *change* in uplift rate along the Fiamignano fault that occurred ca. 1 Ma, and that the convex reach represents a transient response to tectonics.

To test this interpretation, we also consider the possibility that coarse debris delivered from the steep hillslopes along the gorge could cause the breakdown in hydraulic geometry. Wohl (2004) proposed that hydraulic scaling develops only if the river is able to move the channel substrate. Her empirical analysis showed that if $\Omega/D_{84} > 10^4$ kg/m^3, (Ω = total stream power, and D84 represents the grain-size for which 84% of the substrate is finer) good scaling develops, whereas if $\Omega/D_{84} < 10^4$ kg/m^3, channels tend to scale poorly because the substrate is too coarse to be easily moved by typical flows. Figure 5C shows that D_{84} along the Rio Torto does double toward the fault; however, this is outpaced by the increase in Ω, so that all Ω/D_{84} values are significantly above 10^4 kg/m^3. Moreover, Ω/D_{84} peaks in the zone where the hydraulic scaling breaks down, and this trend is mirrored by Shields stresses that are above the critical threshold for sediment entrainment (Fig. DR2; see footnote 1). Consequently, a downstream increase in roughness cannot explain the loss of hydraulic scaling.

We hypothesize that slip on the fault locally steepens the channel, increasing flow velocity and bed shear stress. If shear stress exceeds a critical erosion threshold, downcutting is initiated in the thalweg where V is highest, resulting in a narrower channel with lower aspect ratio. This process allows the channel to maximize stream power and hence incision rates immediately upstream of the active fault. We can account for this effect by adjusting Finnegan's width equation to include the observed power-law dependence of W_b/H on S (Fig. 4A), giving $W_b \sim A^{0.38}S^{-0.44}$. This shows that S can be as important as A for determining widths in non-steady-state channels.

Finally, we note that the peak in unit stream power along the Rio Torto extends <3 km upstream (Fig. 5B), whereas footwall uplift associated with normal faulting declines to zero over distances of ~10 km in this area, (Lavecchia et al., 1994); this indicates that the channel has not reached topographic steady state. The 400-m-high oversteepened reach represents the imbalance of relative tectonic uplift rate (0.6–1 mm/yr upstream; Fig. 1C) minus the fluvial incision rate, which must match uplift at the fault but declines more rapidly upstream (Fig. 5B). Hence, limits to the applicability of hydraulic geometry are intimately associated with the channel's transient response to tectonic forcing. These perturbations can persist for time periods longer than 1 m.y.

CONCLUSIONS

Our results constitute a unique field example of dynamic channel adjustment to tectonic forcing and demonstrate that equilibrium assumptions of hydraulic geometry, constant aspect ratio, and topographic steady state must be used with extreme caution when evaluating fluvial responses to tectonics. We show that channel narrowing is an intrinsic way by which the fluvial system maximizes its erosional response to tectonics and that aspect ratio is a strongly nonlinear function of channel gradient, so that slope is as important as discharge for determining width. Unit stream powers calculated from field data are up to four times higher than those calculated using traditional scaling relationships, and they explain why antecedent drainages are more common than many landscape models predict (Cowie et al., 2006). The breakdown in hydraulic scaling is best explained as a transient response to a change in fault uplift rate that occurred ca. 1 Ma.

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REFERENCES CITED

Cowie, P.A., and Roberts, G.P., 2001, Constraining slip rates and spacings for active normal faults: Journal of Structural Geology, v. 23, p. 1901–1915, doi: 10.1016/S0191–8141(01)00036–0.

Cowie, P.A., Attal, M., Tucker, G.E., Whittaker, A.C., Naylor, M., Ganas, A., and Roberts, G.P., 2006, Investigating the surface process response to fault interaction and linkage using a numerical modeling approach: Basin Research, v. 18, p. 231–266.

Duvall, A., Kirby, E., and Burbank, D.W., 2004, Tectonic and lithologic controls on channel profiles and processes in coastal California: Journal of Geophysical Research, v. 109, p. F3, doi: 10.1029/2003JF000086.

Finnegan, N.J., Roe, G., Montgomery, D.R., and Hallet, B., 2005, Controls on the channel width of rivers: Implications for modeling fluvial incision of bedrock: Geology, v. 33, p. 229–232, doi: 10.1130/G21171.1.

Harbor, D., 1998, Dynamic equilibrium between an active uplift and the Sevier River, Utah: The Journal of Geology, v. 106, p. 181–193.

Howard, A.D., Dietrich, W.E., and Seidl, M.A., 1994, Modelling fluvial erosion on regional to continental scales: Journal of Geophysical Research, v. 99, p. 13,971–13,986, doi: 10.1029/94JB00744.

Kirby, E., Whipple, K.X., Tang, W., and Chen, Z., 2003, Distribution of active rock uplift along the eastern margin of the Tibetan Plateau: Inferences from bedrock channel longitudinal profiles: Journal of Geophysical Research, v. 108, no. B4, 2217, doi: 10.129/2001JB000861.

Lavé, J., and Avouac, J.P., 2001, Fluvial incision and tectonic uplift across the Himalayas of central Nepal: Journal of Geophysical Research, v. 106, p. 26,561–26,591, doi: 10.1029/2001JB000359.

Lavecchia, G., Brozzetti, F., Barchi, M., Menichetti, M., and Keller, J.V.A., 1994, Seismotectonic zoning in east-central Italy deduced from an analysis of the Neogene to present deformations and related stress fields: Geological Society of America Bulletin, v. 106, p. 1107–1120, doi: 10.1130/0016–7606(1994)106.

Leopold, L.B., and Maddock, T., 1953, The hydraulic geometry of stream channels and some physiographic implications: U.S. Geological Survey Professional Paper 252, 57 p.

Montgomery, D.R., and Gran, K.B., 2001, Downstream variations in the width of bedrock channels: Water Resources Research, v. 37, p. 1841–1846, doi: 10.1029/2000WR900393.

Roberts, G.P., and Miccheti, A.M., 2004, Spatial and temporal variations in growth rates along active normal fault systems: An example from Lazio-Abruzzo, central Italy: Journal of Structural Geology, v. 26, p. 339–376.

Selby, M.J., 1980, A rock mass strength classification for geomorphic purposes, with tests from Antarctica and New Zealand: Zeitschrift für Geomorphologie, v. 24, p. 31–51.

Stark, C.P., 2006, A self-regulating model of bedrock river channel geometry: Geophysical Research Letters, v. 33, p. L04402, doi: 10.1029/2005GL023193.

Whipple, K.X, and Tucker, G.E., 1999, Dynamics of the stream power incision model: Implications for the height limits of mountain ranges, landscape response timescales and research needs: Journal of Geophysical Research, v. 104, p. 17,661–17,674, doi: 10.1029/1999JB900120.

Whipple, K.X, and Tucker, G.E., 2002, Implications of sediment-flux dependent river incision models for landscape evolution: Journal of Geophysical Research, v. 107, no. B2, 20 p., doi: 10.1029/2000JB000044.

Wobus, C.W., Tucker, G.E., and Anderson, R.S., 2006, Self-formed bedrock channels: Geophysical Research Letters, v. 33, L18408, 6 p., doi: 10.1029/2006GL027182.

Wohl, E., 2004, Limits of downstream hydraulic geometry: Geology, v. 32, p. 897–900, doi: 10.1130/G20738.1.

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