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21ST KECK RESEARCH SYMPOSIUM IN GEOLOGY SHORT CONTRIBUTIONS

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Keck Director
Franklin & Marshall College

Keck Geology Consortium
Franklin & Marshall College
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2007-2008 PROJECTS:

Tectonic and Climatic Forcing of the Swiss Alps
John Garver (Union College), Mark Brandon (Yale University), Alison Anders (University of Illinois), Jeff Rahl (Washington and Lee University), Devin McPhillips (Yale University)
Students: William Barnhart, Kat Compton, Rosalba Queirolo, Lindsay Rathnow, Scott Reynhout, Libby Ritz, Jessica Stanley, Michael Werner, Elizabeth Wong

Geologic Controls on Viticulture in the Walla Walla Valley, Washington
Kevin Pogue (Whitman College) and Chris Oze (Bryn Mawr College)
Students: Ruth Indrick, Karl Lang, Season Martin, Anna Mazzariello, John Nowinski, Anna Weber

The Árnes central volcano, Northwestern Iceland
Brennan Jordan (University of South Dakota), Bob Wiebe (Franklin & Marshall College), Paul Olin (Washington State U.)
Students: Michael Bernstein, Elizabeth Drewes, Kamilla Fellah, Daniel Hadley, Caitlyn Perlman, Lynne Stewart

Origin of big garnets in amphibolites during high-grade metamorphism, Adirondacks, NY
Kurt Hollocher (Union College)
Students: Denny Alden, Erica Emerson, Kathryn Stack

Carbonate Depositional Systems of St. Croix, US Virgin Islands
Dennis Hubbard and Karla Parsons-Hubbard (Oberlin College), Karl Wirth (Macalester College)
Students: Monica Arienzo, Ashley Burkett, Alexander Burpee, Sarah Chamlee, Timmons Erickson Andrew Estep, Dana Fisco, Matthew Klinman, Caitlin Tems, Selina Tirtajana

Sedimentary Environments and Paleoecology of Proterozoic and Cambrian "Avalonian" Strata in the United States
Mark McMenamin (Mount Holyoke College) and Jack Beuthin (U of Pittsburgh, Johnstown)
Students: Evan Anderson, Anna Lavarreda, Ken O'Donnell, Walter Persons, Jessica Williams

Development and Analysis of Millennial-Scale Tree Ring Records from Glacier Bay National Park and Preserve, Alaska (Glacier Bay)
Greg Wiles (The College of Wooster)
Students: Erica Erlanger, Alex Trutko, Adam Plourde

The Biogeochemistry and Environmental History of Bioluminescent Bays, Vieques, Puerto Rico
Tim Ku (Wesleyan University) Suzanne O'Connell (Wesleyan University), Anna Martini (Amherst College)
Students: Erin Algeo, Jennifer Bourdeau, Justin Clark, Margaret Selzer, Ulyanna Sorokopoud, Sarah Tracy

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THE RELATIONSHIP BETWEEN CHANNEL MORPHOLOGY OF BEDROCK RIVERS AND EROSIONAL PROCESSES IN TICINO, SWITZERLAND

ELIZABETH WONG: Yale University
Research Advisor: Mark Brandon

INTRODUCTION

The Lepontine dome in the Alps lies in the Lago Maggiore region in Ticino, the southernmost canton of Switzerland. This area is inside one of the two major wet anomalous zones, extending over an area greater than 600km^2, and receives about four to five times more precipitation than the surrounding regions. Precipitation peaks in May, before falling and reaching a secondary maximum in October (Frei and Schar, 1998). This area of maximum precipitation also corresponds to a structural low, and, together with the high-density river network, suggest that fluvial erosion and bedrock incision are significant processes in determining the landscape evolution of the Lepontime dome.

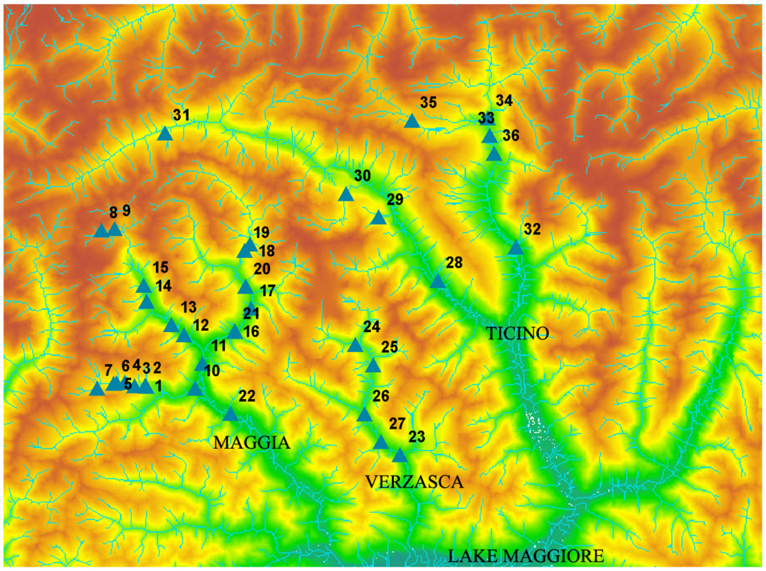

Figure 1. Topographic map of the Ticino canton, Switzerland, showing study sites numbered 1 to 36 along three major rivers (Maggia, Verzasca and Ticino) and their tributaries. All three rivers drain into Lake Maggiore.

Most models that attempt to explain the dynamics of bedrock incision and sediment transport have used unit stream power and mean bed shear stress to characterize river incision (see Whipple, 2002). In particular, the stream-power model relates bedrock erosion rate, E, as a function of river slope, S, and discharge, Q: $E = kQ^mS^n$, where k, m and n are controls on discharge and slope. Hence, to understand precipitation controls on landscape of the region, this study examines several variables that define channel morphology – slope, discharge, mean shear stress, bedload transport capacity and clast size.

STUDY AREA

The Lago Maggiore region was glaciated in the Quatanery, and hence the basin morphology is characterized by broad U-shaped valleys filled with alluvial deposits and occasional glacier moraines. The entire region is underlain by highly metamorphosed and very resistant bedrock. Three major rivers flow into Lake Maggiore: the Maggia, Verzasca and Ticino rivers, whose headwaters can be found up north in the south rim of the Alpine ridge. These rivers are mixed alluvial-bedrock rivers. In all cases, a thin alluvial cover was present. In some of the study sites, the river was flowing through a spatially extensive and temporal alluvial fill, whose bed and banks were made up of sediments (Whipple, 2002).

Measurements were taken at 36 alluvial reaches scattered throughout all three rivers and several of their tributaries (Fig. 1). The channel morphologies of the sites fell into either the step-pool or the plane-bed category with a submergence ratio between 3 – 10 (Fig. 2) (Montgomery and Buffington, 1997; Whipple, 2002).

Figure 2. A characteristic example of a study reach. Upstream view of site 20 on the River Maggia. Site shows a step-pool channel morphology, implying a submergence ratio (H/D_{50}) between 1 to 4 (Whipple, 2003).

METHODS

Measurements

Local slope and width measurements were taken at each site, if possible. When taking local slope measurements, we were careful to stand as close as possible to the water surface for the best approximation of energy slope. Clast sizes in the course surface layer were measured using the Wolman pebble count method, where a minimum of 100 fluvially-deposited clasts were chosen through a 'random walk' process. The size of the clast was taken to be the length of the intermediate b-axis, determined by slipping the clast into holes on a metal template ('gravelometer'), with each succesive hole 0.5Φ larger. Mean annual discharge, Q ($m^3/m/s$) was estimated using the product of local upstream drainage area, A, and local precipitation for a 90-m-resolution DEM.

Calculations

The mobility of the bed surface materials can be evaluated through selective entrainment equations. Selective entrainment is dependent on the progressive entrainment of clasts from a mixed deposit from fine to coarse as shear stress increases. Komar (1987) proposes an expression derived from the classic Shields' criterion equation which accounts for deposits of mixed sizes, to estimate the critical shear stress (τ_c) needed to entrain a particular clast size D relative to the median clast size of bed material, D_{50}:

Equation (1): $\tau_c = 0.045 (\rho_s - \rho)\, g\, D_{50} 0.5\, D0.4$

where ρ_s is the density of the clasts (2650 kg/m^3), ρ is the density of water (1000 g/m^3) and g is gravitational acceleration. Substituting D with D_{50} and D_{84}, we get estimates for critical shear stress for that specific clast size, τ_c (D_{50}) and τ_c (D_{84}) respectively (Table 1).

At the point of entrainment, bed shear stress τ must be equal or greater than the critical shear stress. Therefore, using the defining equation of bed shear stress, critical shear stress can be substituted in place of bed shear stress to calculate minimum flow depth, d_c in meters:

Equation (2): $\tau_c = \tau = \rho\, g\, d\, S$
(minimum condition needed for entrainment).

Estimates of d_c were employed in estimating cross-section average velocity U given by the empirical Manning (1981) equation:

Equation (3): $U = R^{2/3} S^{1/2} / n$,

where R is the hydraulic radius, S is the local slope, and Manning's n is an empirical roughness constant. d approximates R in natural channels. Critical discharge per unit bed width qc can be then calculated:

Equation (4): $q_c = \tau_c\, U / W_{bf}$

where W_{bf} is the channel width at bankfull condition. q_c estimates can be found in Table 1.

Ferguson (1994) formulates a different expression for critical discharge per unit bed width, using a modified version of the Shields' criterion and the Darcy-Weisbach empirical flow law:

Equation (5): $q_c = a\, D_{50} 1.5\, (D / D_{50})\, (1-x)^{(c+1.5)} / S^{\,c+1}$

Equation (6): $a = m\, (8g)\, 0.5\, ((\rho_s / \rho - 1)\, \tau_{c\ 50}^{*})\ c+1.5$

where x is the hiding factor (x=0.90; Parker, 1990), m and c are constants found in the flow law (c=0.37, m=1.14; Thompson and Campbell, 1979), and $\tau_c{}^*{}_{50}$ is the critical dimensionless shear stress ($\tau_c{}^*{}_{50}$ = 0.045). Ferguson's (1994) qc estimates can also found in Table 1. qc estimates derived from Komar (1987) and Ferguson (1994), though formulated with different theoretical basis, are similar (r^2 = 0.9918), proving the validity of both methods.

Both Komar's (1987) and Ferguson's (1994) equations were formulated on the basis of selective entrainment. However, equal mobility, unlike selective entrainment, occurs when clasts of all sizes in a mixed deposit are entrained at one particular mean shear stress. There is no size selection due to the hiding and shielding effects (the hiding factor x = 1). Several studies have found that equal mobility might be applicable at high excess shear stresses and transport rates above twice the critical value of motion (Ashworth and Ferguson, 1989; Parker, 1990). Since the Lago-Maggiore region is known to be affected by severe precipitation events (see Frei and Schar, 1998), the occurrence of equal mobility in river channels especially during these extreme precipitation events is not an implausible concept.

Mueller and Pitlick (2005) employs the concept of equal mobility in formulating an expression for bed load transport capacity Q_s, where the amount of sediment transported is a function of excess shear stress ($\tau^*_{bf} - \tau^*_r$):

Equation (7): $Q_s = 11.2 (\tau^*_{bf} - \tau^*_r)$ 4.5 [(s – 1) g $D_{50}{}^3$] 0.5 $W_b f / \tau^*_b f^3$

where τ^*_{bf} is the dimensionless bed shear stress, τ^*_r is the dimensionless reference shear stress, s is the specific gravity of sediment (s=2.65).

To obtain τ^*_{bf} and $\tau^*_{r,}$ I used these empirical relations in Mueller et al. (2005) that correspond to a submergence ratio of 5 – 10:

Equation (8): $\tau^*_{bf} = 3.5S + 0.0027$
Equation (9): $\tau^*_r = 3.23S + 0.013$

These empirical relations were developed from 45 gravel-bed rivers in the western United States and Canada in a wide range of slopes and sizes. However, four of our study sites (5, 29, 30 and 31) have slopes that are greater than 0.1 and therefore lie outside Mueller et al.'s (2005) range of slopes. Results from these sites should be treated with caution.

RESULTS

If we compare the mean annual discharge value Q with the critical discharge estimate Q_c, we find that $Q >> Q_c$ such that the ratio of $Q / Q_c > 2$. Therefore, if precipitation was constant year-round, we expect not only equal mobility to occur, but also that the bed load would be constantly moving. Field observations, however, do not support the model of a continually-moving bed load at normal flow conditions. This result would support the record of severe precipitation events that produce extreme discharge values, so that Q values are not characteristic of normal flow conditions.

In addition, values of Q_s, the bed load transport capacity, are negligible or even zero because $\tau^*_r > \tau^*_{bf}$. For some of the sites, the bed shear stress at bankfull condition is not greater than the reference shear stress. Hence, the results suggest that for most of these study sites, no sediment is transported even at bankfull condition because of excess shear stress values are negligible.

If there is no excess shear stress even at bankfull condition, this implies that the river is unable to incise. The bankfull discharge therefore does not appear to control channel morphology. To support this, consider that τ_c (D_{84}) has been regarded as the shear stress are the principal factor in determining the channel roughness and therefore can be regarded as the formative shear stress in studies of gravel-bed rivers (e.g. Jansen, 2006; Howard, 1980; Howard, 1987). If so, then one would expect that a relation between minimum flow depth of entrainment, d_c (D_{84}) and the bankfull width, W_{bf}. This because natural channels have a constant width-to-depth ratio (Finnegan et al., 2005). However, there seems to be no systematic pattern between the two variables (Fig. 3).

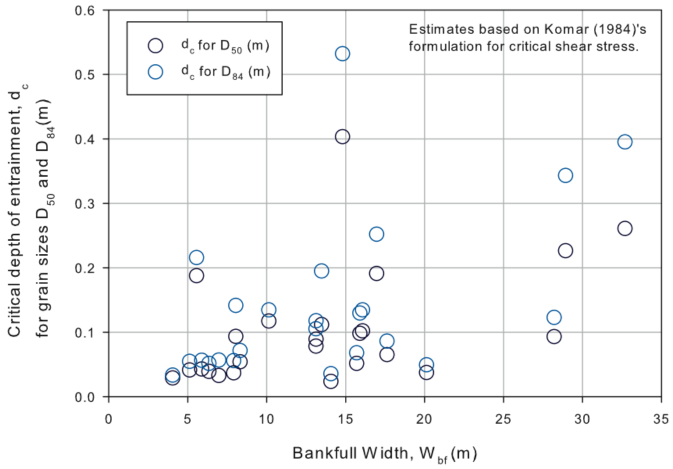

Figure 3. Weak relationship between bankfull width, Wbf(m) and critical depth of entrainment, d_c for grain sizes D50 and D84 (m). The critical shear stress that entrained bed surface material in a particular channle is not reflective of its channel morphology

DISCUSSION

Equation (7) proposed by Mueller and Pitlick (2005) does not seem predict annual bed load transport capacity well. Excess shear stress values, $\tau^*_{bf} - \tau^*_{r}$, which controls Q_s, is negligible and implies that rivers in this region are not actively-incising. The result is also not supported by the ratio of Q / Q_c which, for most sites, is greater than 2, minimum condition needed for equal mobility and hence active sediment transport the entire year.

The reason for the disparity in the two observations is that equation (7) assumes that slope and therefore channel morphology remains constant. Considering the “flashiness” of the Ticino region, it is likely that when an extreme precipitation event occurs, the channel morphology would be altered to allow for the increase in discharge and stream power. The thin alluvial cover that makes up the bed and banks would be eroded, flow depth and velocity would increase, and active fluvial incision into the bedrock would occur. This explanation also suggests that most of the sediment transport occurs during these precipitation events, which explains negligible values of Q_s (which is calculated from S during normal, low-flow conditions – not steep enough to transport bed load). For most of the year, bed shear stress is unable to carry sediment.

To quantify this phenomenon, the next step would be to find out how often the threshold for incision is exceeded in these mixed alluvial-bedrock rivers, and link that to the flow record to determine the geomorphically-effective (channel-shaping) discharge and associated grain size. With knowledge of the effective discharge and bed shear stress, the erosion rate and sediment transport can be constrained.

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Study Site	D_{50} (m)	D_{84} (m)	Slope, S (m)	W_{bf}(m)	A (km^2)	Q (m^3/s)	τ_c (D_{50}) [a] (N/m^2)	τ_c(D_{84}) [a] (N/m^2)	q_c (D_{50}) [b] ($m^3/m/s$)	q_c (D_{50}) [c] ($m^3/m/s$)	Q_c (D_{50}) (m_3/s)	Q/ Qc (D50)	τ^*_{bf} (N/m^2)	τ^*_c(N/m^2)	Q_s(kg/yr)
1	0.064	0.128	0.088	8.32	26.1	1.39	46.6	61.5	0.045	0.035	0.379	3.7	0.297	0.310	0
2	0.032	0.127	0.073	6.97	22.9	1.22	23.3	40.4	0.018	0.016	0.125	9.7	0.247	0.257	0
3	0.045	0.091	0.081	5.12	21.1	1.12	33.0	43.5	0.028	0.023	0.143	7.8	0.276	0.287	0
4	0.045	0.128	0.091	7.91	18.8	0.99	33.0	50.0	0.024	0.020	0.194	5.1	0.308	0.322	0
5	0.064	0.128	0.122	6.34	15.7	0.83	46.6	61.5	0.031	0.023	0.197	4.2	0.406	0.429	0
6	0.045	0.091	0.078	5.89	14.1	0.74	33.0	43.5	0.029	0.025	0.172	4.3	0.266	0.277	0
7	0.064	0.128	0.047	16.07	6.2	0.32	46.6	61.5	0.095	0.084	1.531	0.2	0.163	0.166	0
8	0.045	0.091	0.023	-- [e]	1.4	0.07	33.0	43.5	--	0.129	--	--	0.089	0.085	--
10	0.045	0.128	0.036	8.05	101.0	5.59	33.0	50.0	0.072	0.071	0.578	9.7	0.129	0.129	0.0001
11	0.032	0.064	0.026	28.21	300.2	15.48	23.3	30.7	0.061	0.068	1.717	9.0	0.095	0.092	2.09
12	0.045	0.091	0.008	14.81	102.1	5.34	33.0	43.5	0.388	0.530	5.749	0.9	0.040	0.032	1199
13	0.032	0.091	0.010	28.93	93.9	4.89	23.3	35.3	0.170	0.230	4.933	1.0	0.047	0.039	612
14	0.064	0.091	0.025	5.56	62.0	3.17	46.6	53.5	0.187	0.195	1.042	3.0	0.095	0.091	1.28
15	0.032	0.045	0.082	4.04	1.0	0.01	23.3	26.8	0.015	0.014	0.062	0.2	0.279	0.291	0
16	0.032	0.064	0.064	20.13	159.7	8.05	23.3	30.7	0.021	0.019	0.423	19.0	0.218	0.225	0
17	0.045	0.091	0.0417 [d]	--	152.6	7.67	33.0	43.5	--	0.058	--	--	0.148	0.149	0
18	0.045	0.091	0.018	16.97	58.1	2.86	33.0	43.5	0.166	0.190	2.811	1.0	0.070	0.064	48.0
19	0.032	0.064	0.027	13.13	59.5	2.91	23.3	30.7	0.058	0.064	0.756	3.8	0.099	0.096	0.571
20	0.064	0.257	0.043	13.48	113.3	5.65	46.6	81.3	0.106	0.096	1.425	4.0	0.150	0.152	0
21	0.032	0.064	0.064	20.13	159.6	8.05	23.3	30.7	0.021	0.019	0.423	19.0	0.218	0.225	0
22	0.032	0.091	0.006	--	436.3	23.04	23.3	35.3	--	0.465	--	--	0.033	0.025	--
23	0.032	0.045	0.020	--	125.6	6.75	23.3	26.8	--	0.094	--	--	0.078	0.073	--
24	0.045	0.064	0.029	10.13	20.8	1.13	33.0	37.9	0.094	0.098	0.950	1.2	0.106	0.103	0.253
25	0.032	0.064	0.024	15.90	55.8	3.02	23.3	30.7	0.065	0.073	1.026	2.9	0.091	0.087	2.13
26	0.045	0.128	0.013	32.70	103.5	5.52	33.0	50.0	0.239	0.292	7.822	0.7	0.055	0.048	489
27	0.064	0.128	0.030	--	112.8	6.04	46.6	61.5	--	0.152	--	--	0.111	0.109	--
28	0.023	0.045	0.033	15.70	341.3	15.98	16.5	21.7	0.026	0.029	0.403	39.7	0.118	0.117	9.34E-03
29	0.045	0.128	0.183	8.41	5.3	0.25	33.0	50.0	0.011	0.008	0.092	2.7	0.603	0.642	0
30	0.032	0.091	0.101	14.07	18.6	0.90	23.3	35.3	0.012	0.010	0.172	5.2	0.340	0.357	0
31	0.064	0.128	0.206	16.92	3.4	0.16	46.6	61.5	0.017	0.011	0.286	0.6	0.679	0.724	0
32	0.032	0.064	0.036	17.62	245.3	10.55	23.3	30.7	0.040	0.042	0.708	14.9	0.131	0.130	6.73E-05
34	0.045	0.128	0.037	--	100.5	4.19	33.0	50.0	--	0.068	--	--	0.134	0.134	--
35	0.011	0.023	0.007	--	22.6	0.96	8.2	10.9	--	0.093	--	--	0.034	0.026	--
36	0.064	0.133	0.061	13.12	163.3	6.85	46.6	62.4	0.070	0.059	0.921	7.4	0.209	0.215	0

Table 1: Site Data and Estimations

[a] Estimated using Equation from Komar (1987).

[b] Estimated using Equation from Komar (1987) and Gauckler-Manning's equation.

[c] Estimated using Equation from Ferguson (1994).

[d] Measurement taken off the DEM.

[e] All entries marked as "--" indicate that no measurements were taken, or cannot be estimated from available data.