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1 **Quantifying the competing influences of lithology and throw rate on bedrock**
2 **river incision.**

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12 **ABSTRACT (MAX 250 WORDS)**

13 River incision in upland areas is controlled by prevailing climatic and tectonic
14 regimes, which are increasingly well-described, and the nature of the bedrock
15 lithology that is still poorly constrained. Here, we calculate downstream variations in
16 stream power and bedrock strength for six rivers crossing a normal fault in Western
17 Turkey, to derive new constraints on bedrock erodibility as function of rock type.
18 These rivers are selected as they exhibit knickzones representing a transient
19 response to an increase in throw rate, driven by fault linkage. Field measures of rock
20 mass strength show that the metamorphic units (gneisses and schists) in the
21 catchments are ~ 2 times harder than the sedimentary lithologies. Stream power
22 increases downstream in all rivers, reaching a maxima upstream of the fault within
23 the metamorphic bedrock but declining markedly where softer sedimentary rocks are
24 encountered. We demonstrate a positive correlation between throw rate and stream
25 power in the metamorphic rocks, characteristic of rivers obeying a detachment-
26 limited model of erosion. We estimate bedrock erodibility in the metamorphic rocks
27 as $k_b = 2.2 - 6.3 \times 10^{-14} \text{ ms}^2\text{kg}^{-1}$; in contrast, bedrock erodibility values are 5 - 30
28 times larger in the sedimentary units with $k_b = 1.2 - 15 \times 10^{-13} \text{ ms}^2\text{kg}^{-1}$. However, in
29 the sedimentary units stream power does not scale predictably with fault throw rate,
30 and we evaluate the extent to which the friable nature of the outcropping clastic
31 bedrock alters the long term erosional dynamics of the rivers. This study places new
32 constraints on bedrock erodibilities upstream of active faults and demonstrates that
33 the strength and characteristics of underlying bedrock exert a fundamental influence
34 on river behaviour.

35

36 **KEYWORDS:** stream power, rivers, erosion rate, knickzone, Gediz Graben

37

38 1. INTRODUCTION

39 One of the most significant ways in which tectonic and climatic forcing drives
40 landscape evolution is through the action of rivers at the Earth's surface (Lave and
41 Avouac, 2001; Wobus et al., 2006; Kirby and Whipple, 2012; Whittaker et al., 2012;
42 Ferrier et al., 2013; D'Arcy and Whittaker, 2014). Consequently, in the past two
43 decades landscape evolution models (LEMs) have been widely developed to
44 quantify the impact of external factors, such as active faulting, upon bedrock rivers
45 and the surrounding landscape (Braun and Sambridge, 1997; Tucker et al., 2001;
46 Hancock et al., 2002; Whipple and Tucker, 2002; Willgoose, 2005; Van De Wiel et
47 al., 2007; Perron and Fagherazzi, 2012). Owing to the importance of rivers in driving
48 erosion in these models, it is vital that time-integrated fluvial incision is appropriately
49 parameterised (e.g. Tucker and Whipple, 2002; Lague et al., 2014). It is also
50 necessary to constrain how factors such as channel slope, geometry and discharge
51 may control shear stresses on the bed, and thus modulate bedrock erosion over time
52 and space (Lavé and Avouac 2001; Duvall et al. 2004; Whittaker et al., 2007b,
53 Whittaker et al., 2008; Allan et al., 2012; Whittaker and Boulton, 2012). Stream
54 power erosion laws, which relate the rate of bedrock incision, E , to local channel
55 slope, S , water discharge, Q , (or upstream drainage area, A , as a proxy) have been
56 commonly used for this purpose as they are relatively tractable for both modelling
57 and for applications over longer timescales (e.g. Kirby and Whipple, 2012; Lague et
58 al., 2014).

59

60 In general, the family of stream power erosion laws can be expressed for rivers near
61 the detachment-limited end-member as

$$62 \quad E = KA^m S^n \quad (1)$$

63 where exponents m and n can be determined empirically or theoretically, and K is a
64 coefficient which encapsulates, alongside additional variables, bedrock erodibility
65 (Whipple and Tucker, 1999; Tucker and Whipple, 2002). The value of K (and the
66 units in which it is expressed) depends on the choice of stream power model. While
67 equation 1 (and variants thereof) is routinely employed in numerical modeling or
68 topographic analyses (e.g. Wobus et al., 2006; Schwanghart and Scherler, 2014;

69 Clubb et al., 2019), where field data allow, a unit (or specific) stream power model
70 (c.f. Whittaker et al., 2007a; Attal et al., 2008; Zondervan et al., 2020a) can be used.
71 In this case,

$$72 \quad E = k_b \omega = k_b \frac{\rho g Q S}{W} \quad (2)$$

73 where the unit stream power, ω represents energy dissipation per unit channel area
74 on the bed with units of Wm^{-2} , ρ is the density of water, g is the acceleration due to
75 gravity, Q is the water discharge (m^3s^{-1}), S is local channel slope (m/m) and W the
76 channel width (m). Consequently, it follows that specific bedrock erodibility, k_b , has
77 units of $\text{m s}^2 \text{kg}^{-1}$, representing the inverse of stress (c.f. Yanites et al., 2017). Whilst
78 the influence of factors such as planform width (e.g., Montgomery and Gran, 2001;
79 Finnegan et al., 2005; Whittaker et al., 2007b; Turowski et al., 2009), channel
80 steepness (e.g., Whipple and Tucker, 2002; Whipple, 2004; Whittaker et al., 2007a),
81 and sediment supply effects (e.g., Sklar and Dietrich, 2001; Finnegan et al., 2007;
82 Cowie et al., 2008) on stream power erosion laws have been widely investigated, the
83 role of bedrock lithology and erodibility in modulating landscape response to base
84 level change remains poorly understood (Stock and Montgomery, 1999; Reneau,
85 2000; Van der Beek and Bishop, 2003; Brocard and Van der Beek, 2006; Allen et al.,
86 2012; Bursztyn et al., 2015). Unlike other parameters in a stream power model,
87 bedrock erodibility cannot be directly measured from lithological maps or rock
88 strength tests, nor is there agreement about what measure or measures of rock
89 strength are most suited to capture variations in time-integrated substrate erodibility
90 (c.f. Sklar and Dietrich, 2004; Bursztyn et al., 2015). However, for bedrock rivers
91 whose erosional dynamics can be approximated by a specific stream power model,
92 k_b can be calculated if and where incision rates are known and where the down-
93 system distribution of stream power can be computed from field data (Stock and
94 Montgomery, 1999; Attal et al., 2008, Zondervan et al., 2020a).

95

96 Lithology has long been identified by researchers as being of potential significance in
97 controlling river behaviour (Howard et al., 1994; Goldrick and Bishop, 1995; Cook et
98 al., 2009; Allen et al., 2013; Ferrier et al., 2013; Croissant and Braun, 2014).
99 Recent landscape evolution modelling demonstrates that variations in bedrock
100 erodibility may play an important role in determining the speed and coherence of
101 knickpoint signals driven by base level change (e.g. Forte et al., 2016; Roy et al.,

102 2016; Yanites et al., 2017), and when combined with dipping bedrock, complex
103 interactions between fluvial erosion rate and topographic evolution can developed
104 (Forte et al., 2016, Perne et al., 2016). These effects are particularly enhanced
105 where modelled bedrock erodibilities vary by several orders of magnitude. However,
106 there are few field data sets that show these effects unequivocally (c.f. Whittaker and
107 Boulton, 2012; Kent et al., 2017), and few constraints on what bedrock erodibilities
108 would actually be appropriate to characterise the range of rock types encountered in
109 real-world catchments. For instance, Stock and Montgomery (1999) calibrated
110 Equation 1 using geologically-constrained forward modelling of river profiles in a
111 number of localities world-wide and obtained K values that varied by at least 5 orders
112 of magnitude ($10^{-7} - 10^{-2} \text{ m}^{0.2} \text{ yr}^{-1}$) for $m = 0.4$. Kirby and Whipple (2001) and
113 Whipple et al. (2000) also obtained K values for rivers in Alaska and the Siwalik hills
114 of ca. $10^{-4} \text{ m}^{0.2} \text{ yr}^{-1}$ although these values embed significant differences in discharge
115 and vegetation between the respective landscapes. In contrast, erodibility values of
116 the order of $1.5 - 4 \times 10^{-6} \text{ m}^{0.2} \text{ yr}^{-1}$ have recently been estimated for the Gulf of
117 Corinth (Pechlivanidou et al., 2018). Significantly, none of the above studies linked
118 these estimates to any physical measurements of bedrock strength or local
119 measurements of channel hydraulic geometry.

120 In fact, the limited number of field-related studies that have specifically focussed on
121 the role of the bedrock lithology in landscape evolution have generally been
122 undertaken in post-orogenic or tectonically quiescent regions with low or negligible
123 rates of tectonic uplift. For example, Bursztyn et al. (2015) argued that lithological
124 strength is a first-order control on the fluvial geomorphology of the Colorado Plateau
125 using a combination of *in situ* strength tests and lab tensile testing. Strong positive
126 correlations between rock strength and unit stream power, as well as between rock
127 strength and river slope / width were found in bedrock reaches of the river. Similarly,
128 rock strength as measured by Schmidt Hammer was shown to strongly correlate with
129 river steepness across the Pyrenean (Bernard et al., 2019) and High Atlas
130 (Zondervan et al., 2020b) mountain ranges. Recently, Yanites et al. (2017) argued
131 that lithology played an important role in the post orogenic evolution of the Jura
132 mountains in the Swiss Alps, although mapped rock type and tectonic evolution
133 provided the context for a range of model results rather than as a fully independent
134 data set.

135

136 Consequently, field studies are lacking in tectonically-active areas where lithology
137 and uplift rates differ but are independently well-constrained. In principle, for a
138 bedrock river incising to keep pace with known fault uplift with differing footwall
139 lithologies, Equation 2 suggests stream power should scale with the fault slip rate,
140 but should also be predictably modulated by bedrock erodibility (c.f. Cowie et al.,
141 2008; Attal et al., 2008). That stream power scales to fault slip rates has been
142 demonstrated for one set of active faults in the Central Italian Apennines (e.g.
143 Whittaker et al., 2007a; Whittaker et al., 2007b), using field observations. However,
144 the hypothesis that stream power scales with bedrock erodibility in these settings
145 has not (c.f. Kent et al., 2017). Moreover, if measures of rock mass strength were
146 known independently in such a situation, it would enable variations in physical
147 measures of rock properties to be linked explicitly to estimates of bedrock erodibility,
148 k_b , from Equation 2 (c.f. Burzryn et al., 2015; Zondervan et al., 2020a,b). Finally,
149 where bedrock incision rates do not scale predictably with throw rates, or with
150 measures of bedrock strength, a deduction may be also made about the role of
151 sediment in modulating bedrock incision, as Cowie et al. (2008) demonstrate for
152 catchments draining across active faults bordering the Gulf of Evia, Greece.

153 In this contribution we address these challenges. We exploit an exceptional field site
154 on the southern margin of the actively-extending Gediz Graben of Turkey, where
155 tectonic and lithological boundary conditions can be constrained independently to: (i)
156 quantify the relationship between stream power and fault uplift rates where rivers
157 cross active faults; (ii) quantify the relationship between stream power, lithology and
158 rock mass strength; (iii) derive the magnitudes of bedrock erodibility for the
159 metamorphic and clastic sedimentary units outcropping in the region. Finally, we
160 explore the extent to which bedrock lithology can influence the erosional dynamics of
161 incision upstream of an active fault.

162

163 **2. STUDY AREA**

164

165 The Gediz Graben (Figure 1) can be used as a natural laboratory to study the fluvial
166 response of the landscape to active faulting, as the geologic context and tectonic
167 evolution of the area are well-understood and constrained (i.e., Seyitoğlu and Scott,

168 1996; Seyitoğlu et al., 2002; Bozkurt, 2003; Bozkurt and Sözbilir, 2004; Çiftçi and
169 Bozkurt, 2009a; Oner and Dilek, 2011; Kent et al. 2016; 2017). In particular, the
170 northward-draining range-transverse rivers have been previously documented to be
171 responding transiently to a documented change in tectonic boundary conditions at
172 0.8 ± 0.2 Ma (Kent et al., 2016; 2017).

173

174 The Gediz (also known as the Alaşehir) Graben is located in western Anatolia
175 (Figure 1), which has been experiencing ~ N-S extension since early Miocene times,
176 probably as the result of roll-back on the Hellenic subduction zone (Okay and Satir,
177 2000; ten Veen et al., 2009). The tectonic history of the southern margin of Gediz
178 Graben can be divided into two main phases (Bozkurt and Sözbilir, 2004). Initial
179 extension caused the uplift of the Menderes Massif metamorphic core complex,
180 along the now-inactive low-angle north-dipping Gediz detachment fault (Gessner et
181 al., 2001; Seyitoğlu et al., 2002; Ring et al., 2003). In the footwall of this fault is the
182 Menderes Massif metamorphic core complex composed of greenschist to
183 amphibolite-facies schists, augengneisses, paragneisses, small amounts of marble
184 and phyllite, and small syn-tectonic granodiorites (Gessner et al., 2001; Ring et al.,
185 2003). The majority of the Menderes Massif lithologies exhibit features associated
186 with early ductile deformation such as schistosity and low-angle foliations (Hetzl et
187 al., 1995). These are overprinted by later fabrics such as mineral lineations (with top
188 to the NE sense of shear), cataclasites and mylonites (Hetzl et al., 1995). Fault-
189 bound slivers of high-grade gneisses show less internal structure and are thought to
190 have been thrust-emplaced upon low-grade lithologies in an earlier phase of
191 deformation (Hetzl et al., 1995). Intense shearing, mylonites and cataclasites are
192 observed for 20-50 m (Hetzl et al., (1995) below the detachment surface, which is a
193 regionally extensive, corrugated surface dipping $\sim 20^\circ$ northwards (Purvis and
194 Robertson, 2005).

195

196 Following the cessation of slip on the detachment at ca. 2 Ma (Buscher et al., 2013),
197 strain stepped basinwards (i.e., northwards) onto high-angle faults, including the
198 presently active normal fault forming the southern margin of the present topographic
199 graben (Figure 1). The Gediz Graben boundary fault has a record of historical and
200 recent earthquakes. In 17 AD the Lydia earthquake caused extensive damage to the

201 region and to the city of Sardis, capital of the ancient kingdom of Lydia and close to
202 the modern town of Sart (Ambraseys, 2009). More recently, a large magnitude
203 earthquake occurred in 1969 ($M_s = 6.5$), resulting in a surface rupture > 30 km in
204 length with up to 20 cm surface displacement along the eastern section of the high
205 angle boundary fault (Arpat & Bingol, 1969; Eyidoğan & Jackson, 1985).

206

207 In the uplifted footwall of this active fault are friable sedimentary rocks, deposited
208 originally in the hangingwall of the Gediz detachment during the early Miocene to
209 Pleistocene. These sedimentary units unconformably overly the metamorphic
210 basement and are comprised of a complex sequence of variably cemented and
211 laterally discontinuous sandstones and conglomerates deposited mainly in alluvial
212 fan and fluvial environments on the edge of the graben (Purvis and Robertson, 2004;
213 2005; Çiftçi and Bozkurt, 2009b). Furthermore, mudstones, siltstones, marls and
214 limestones are locally present associated with lacustrine conditions in depocentres
215 or overbank deposition (Çiftçi and Bozkurt, 2009b). These predominantly clastic
216 sediments have numerous sedimentary structures and bed thicknesses that vary
217 from decimeters to > 10 metres (Çiftçi and Bozkurt, 2009b). It should also be noted
218 that the sedimentary cover to the detachment is cut by numerous small high-angle
219 normal faults, synthetic and antithetic to the boundary fault, which are observed
220 displacing the older detachment surface (Purvis and Robertson, 2005).

221

222 The Bozdağ Range, the southern bounding range of the Gediz Graben, has a
223 characteristic topographic asymmetry. The range is steeper on the southern side,
224 with the drainage divide in places offset towards the Küçük Menderes Graben. To
225 the south of the drainage divide, the range slopes steeply into the Küçük Menderes
226 Graben, giving the Bozdağ Range the topographic characteristics of a horst uplift
227 (Figure 1B) with inferred uniform uplift across the tectonic block. Although faults in
228 the Küçük Menderes Graben are not well mapped nor the activity upon them
229 constrained, catchment averaged erosion rates derived from cosmogenic ^{10}Be
230 measurements in southward draining rivers are generally comparable to those
231 draining northwards (Heineke et al., 2019) supporting this supposition. Transverse
232 bedrock rivers flowing northwards into the Gediz Graben are generally deeply
233 incised with prominent knickpoints and gorges upstream of the active fault (Figures 2
234 and 3). These tectonic knickpoints are not coincident with lithological boundaries or

235 the detachment surface and have characteristic morphologies of slope-break
236 knickpoints (Kent et al., 2017). The knickpoints mark the upstream extent of a
237 transient wave of river incision, caused by an increase in slip on the graben
238 bounding fault as a result of fault linkage 0.6 – 1 Ma, as explored in detail by Kent et
239 al. (2016, 2017), who integrated structural, tectonic and geomorphic data sets to
240 evaluate the fault slip histories and associated landscape response. As a result of
241 this linkage present day throw rates (the vertical component of the slip rate) are now
242 higher than the long-term geological average, with rates of up to 2 mmyr⁻¹ calculated
243 at the centre of the fault array (Kent et al., 2017).

244

245 Kent et al. (2017) documented the transient fluvial characteristics of, and calculated
246 the knickpoint retreat rate, knickpoint celerity and post-linkage slip rate at the range
247 front for 29 rivers crossing the fault array. Here, six of these rivers are selected to
248 investigate the lithological and tectonic controls on transient river behaviour (Figure
249 1). The rivers represent a variety of boundary conditions in terms of the throw rate on
250 the fault and bedrock lithologies (Figure 2 and 3). The highest throw rates are found
251 in the middle section of the basin bounding fault, ~ 60 km along strike, while the
252 lowest throw rates are found towards the fault array tips (Kent et al., 2017). The
253 rivers cross the fault in different locations and therefore, are affected by a three-fold
254 spread of quantified post-linkage throw rates, ranging from 0.7 - 2 mmyr⁻¹ (Figure 1;
255 Table 1).

256

257 A range of lithologies are also exposed along the six river channels (Figures 1 - 3),
258 including gneisses and schists of the Menderes metamorphic complex outlined
259 above (Çiftçi and Bozkurt, 2009b; Oner and Dilek 2011), and the Neogene to recent
260 conglomerates and sandstones found in the hangingwall of the inactive Gediz
261 detachment (Purvis and Robertson, 2004; 2005; Çiftçi and Bozkurt, 2009b). The
262 proportion of these lithologies varies between the channels: for instance, the
263 Akçapınar River (Figure 3A) incises only through the metamorphic basement rocks
264 upstream of the fault, while all five of the remaining rivers incise through varying
265 amounts of clastic sedimentary units in the footwall of the active fault in addition to
266 the metamorphic basement upstream of the inactive detachment fault (Figure 3C).
267 The Yeniköy represents the opposite end member with around half the length of the
268 river being underlain by the uplifted clastic sedimentary rocks. These two broad

269 groupings of rock types, metamorphic and sedimentary, have been used for the
270 subsequent quantitative analyses given that our qualitative field observations
271 suggested the differences between these two main lithological groups are clearly
272 and consistently observed in the resultant morphology of the six rivers. Finally, it
273 should be noted that the climate of the Bozdağ Range is generally uniform within the
274 study area, with precipitation rates varying from 500 to 1000 mm/yr in the highest
275 parts of the range (Şensoy et al., 2008).

276

277 **3. METHODS**

278

279 **3.1 Field Measurements**

280 We traversed the six selected rivers in the field from their headwaters in the Bozdağ
281 Range to the range-bounding fault. Detailed channel measurements (**supplemental**
282 **data tables S1 to S6**) were taken every 200 – 500 m downstream; study locations
283 were mapped using a hand-held GPS with a spatial precision of ± 5 m. In the field
284 we measured bankfull channel width (W); maximum channel depth (H) and local
285 channel slope (S). Hydraulic geometries were measured using a TruPulse laser
286 rangefinder with a precision of < 5 cm. The channel width and height were measured
287 at bankfull stage (Leopold and Maddock, 1953; Knighton, 1998), estimated from
288 features such as the limits of active abrasion, vegetation boundaries, the highest
289 levels of bleaching on boulders and water-washed surfaces, and the remains of high
290 stage flood debris (e.g. Montgomery and Gran, 2001; Snyder et al., 2003; Whittaker
291 et al., 2007a). The local channel slope measurements were measured over 10 – 30
292 m as appropriate for the location in which they were taken. Repeat measurement
293 variation associated with hitting the target with the laser range-finder was $\pm 0.2^\circ$.

294

295 Where exposed, the type of bedrock was documented and the rock mass strength
296 measured (Figure 4 and 5). Intact rock strength was determined using a Schmidt
297 hammer, from this the Selby rock mass strength index (Figure 4 [SRMS]) was
298 calculated (Selby, 1980). Twenty-two Schmidt hammer rebound readings were made
299 at each location and after the highest and lowest value were removed as outliers, the
300 mean value was calculated. Schmidt hammer rebound values are a proxy for
301 uniaxial compressive stress (e.g. Bursztyn et al., 2015). We chose this tool as it is
302 portable and convenient for remote field locations, as measurements are easily and

303 rapidly repeatable at a field site and because the values obtained are widely quoted
304 in the literature (c.f. Goudie et al., 2006). Tensile strength or bedrock cohesion
305 measurements typically require laboratory experiments that are very dependent on
306 the typically small number of samples collected, although we return to this choice of
307 bedrock strength characterisation technique in the discussion, below. In contrast, the
308 SRMS represents a semi-quantitative assessment of rock mass strength based not
309 just on rock strength but also the degree of weathering; ground water saturation; and
310 the orientation, size and spacing of joints and bedding (Supplemental methods,
311 Table S15). Values for the SRMS can range from 0 – 100 with soils and
312 unconsolidated rock having values of < 25 (Selby, 1980; Sklar and Dietrich, 2001).
313 Because the SRMS highlights relative differences in intact rock strength and
314 hardness (Sklar and Dietrich, 2001), the method can give an indication of the
315 bedrock resistance to erosion (Goudie, 2016). This is important because intact rock
316 strength alone is sometimes argued to be a poor indicator of erodibility in heavily
317 jointed lithologies (Whipple et al., 2000; Scott and Wohl, 2019). Paired Schmidt
318 hammer / SRMS assessments were undertaken at 170 locations and data are
319 presented both along the length of each river and shown as violin plots for the
320 different lithological groups measured. For the five catchments where the clastic
321 sedimentary units are exposed near the fault, we also estimated the first order grain
322 size of sediments in transport in the river channel near the basin-fault using a
323 Wolman point count technique on five available scaled photographs (supplementary
324 material Figures S3, S4). Grain size data are not available for ca. 180 individual field
325 sites.

326

327 **3.2 Hydraulic Scaling, Discharge and Unit Stream Power**

328 The first step to deriving unit stream power in the six rivers studied is to estimate the
329 bankfull water discharge. As our rivers are not gauged, we used Manning's equation
330 (Manning, 1891) in conjunction with our channel hydraulic geometry data near the
331 faults to help estimate plausible fluid velocities and river discharges, Q , for each of
332 the channels, where

$$333 \quad Q = vC = \frac{1}{n} H^{2/3} S^{1/2} C \quad (3)$$

334 where v is velocity in ms^{-1} , C is the cross-sectional area of flow (m^2), H is the flow
335 depth (m), and n is a roughness coefficient. However estimates of bankfull

336 discharge from point measures of hydraulic geometry vary both up and down over
337 short length scales, while we are seeking a consistent and monotonic relationship
338 between Q and increasing downstream drainage area, A , for each river (c.f.
339 Whittaker et al., 2007b). To obtain the evolution of catchment drainage areas for
340 each river we used data from a 30 m DEM originally presented in Kent et al. (2017).
341 We subsequently calculated the ratio of $A:Q$ at each site from our field hydraulic
342 geometry measurements, and scaled our discharge estimates to the median value of
343 this ratio. We repeated this process for all 6 rivers (supplementary material, Tables
344 S7 to S12). Finally, given that rainfall does not significantly vary across the local
345 study area and all rivers are expected to have a similar discharge-drainage area
346 scaling, we derived a mean scaling of $A:Q$ from all six rivers and used this ratio to
347 obtain regionally consistent discharge values downstream that we use in all of our
348 subsequent calculations (Table S14 and Fig. S1). This means that variations in
349 stream power and eventual bedrock erodibility downstream, or between the
350 catchments, are not driven by local point variations in discharge estimates or by
351 unjustified differences in drainage area to discharge scaling (c.f. Whittaker et al.,
352 2007). A full description of these calculations is included in the supplementary
353 material.

354

355 Unit stream power, ω (eq. 2), represents the downstream rate of energy dissipation
356 on the bed of a stream per unit channel width. In bedrock channels draining the
357 footwalls of normal faults, existing studies have shown that unit stream power can
358 and does scale with the magnitude and distribution of footwall uplift (e.g. Whittaker et
359 al., 2007b; Cowie et al., 2008). We calculated unit stream power explicitly using field
360 measures of channel width, as rivers crossing active faults are known not to obey
361 typical hydraulic scaling that is often assumed in stream power calculations (Leopold
362 and Maddock, 1953; Whittaker et al., 2007a). Our estimates of Q from our field-
363 estimated regional discharge scaling are combined with the hydraulic geometry data
364 collected downstream for each river (section 3.1) to derive the downstream
365 distribution of specific stream power, ω . As field-derived point measures of ω are
366 particularly sensitive to local channel slope, we also averaged unit stream powers
367 down-system over intervals of approximately 2 km to capture longer-wavelength
368 downstream trends in stream power, consistent with previous workers (e.g. Lave &
369 Avouac, 2001; Whittaker et al., 2007; Cowie et al., 2008). The maximum reach-

370 averaged stream power within the metamorphic lithologies was documented for each
371 river. In addition, the maximum reach averaged stream power for a 2 km interval
372 upstream of the active fault was recorded; this comprised clastic sedimentary
373 lithologies for all channels apart from the Akçapınar River, which is only underlain by
374 schists and gneisses. In each river, the knickpoint representing the farthest extent of
375 the wave of incision lies between 6.5 km and 10 km upstream of the fault for all 6
376 catchments (Figure 3), so these data represent stream power averages in the zone
377 upstream of the active fault that has already adjusted to the relative base level
378 change (c.f. Kent et al., 2017).

379

380 **3.3 Comparison of Stream Powers, Throw Rates and Rock Strength Data**

381 We compare maximum reach-averaged stream power data in both the metamorphic
382 rocks and in the sedimentary units upstream of the active faults with our rock mass
383 strength data to evaluate our hypothesis that stream power should correlate with
384 rock hardness or SRMS values. We also test whether unit stream power scales with
385 the documented variation in fault throw rate, U , in both the sedimentary and
386 metamorphic lithologies (c.f. Whittaker et al., 2007b; Cowie et al., 2008). As the
387 rivers are documented to be undergoing a transient response to fault linkage (Kent et
388 al., 2017) and there are no fault scarps in the channel where the rivers cross the
389 basin bounding faults, indicating they are keeping pace with tectonics, we make the
390 reasonable assumption that river incision rates equal fault uplift rates upstream of
391 the basin-bounding faults and downstream of the knickpoints. The range-bounding
392 fault is not always a single plane so a 2 km reach-averaged stream power estimate
393 upstream of the fault zone enables us to average effectively across a number of data
394 points near the faulted basin margin. However, for comparison, we also compare
395 point measures of stream power where we locate the main strand of the fault with
396 our throw rate data (supplementary material, Table S14 and Fig. S3). Given the
397 structural configuration of the Bozdağ Range as a horst block (section 2), uplift
398 across the footwall block is assumed to be approximately uniform in the absence of
399 corroborating data, so we calculate minimum values of bedrock erodibility, k_b , for our
400 lithological classes in each river (in units of $\text{m s}^2 \text{kg}^{-1}$) as:

$$401 \quad k_b = \omega_{\max} / U \quad (3)$$

402 where ω_{\max} is the reach averaged maximum stream power in the 2 km upstream of
403 the fault for the sedimentary lithologies, or the reach averaged in the metamorphic
404 lithologies where they outcrop in the studied rivers.

405

406

407 **4 RESULTS**

408 ***4.1 Relationship of Rock Strength to Lithology***

409 All six of the rivers studied drain the uplifted Bozdağ Range horst block, between
410 bounding faults of the Gediz and Küçük Menderes grabens (Figure 1), and enter the
411 main Gediz River, which runs roughly east-west through the Gediz Graben. Within
412 each river is a tectonically induced knickpoint was identified by Kent et al. (2017) that
413 does not correlate with known lithological boundaries. However, there are
414 differences in the bedrock lithology across and along the studied rivers. For
415 example, the Akçapınar River incises only through the metamorphic basement rocks
416 upstream of the fault (Figure 3A), whereas the Yeniköy flows through the continental
417 clastic lithologies for half its length (Figure 3B).

418

419 Schmidt hammer rebound readings (Figure 4) indicate that the metamorphic rocks
420 are broadly ca. 2 times harder than the sedimentary units. Schmidt hammer
421 readings for the metamorphic gneisses have an average of rebound value of 53 (n =
422 1180) and schists have an average Schmidt hammer value of 44 (n = 1200). In
423 contrast the average rebound value of sediments is 28 (n = 880). This range of
424 values is like those reported by other studies for similar rock types (i.e., Goudie,
425 2016). The SRMS (which also considers the orientation and characteristics of joints)
426 of the metamorphic rocks is 1.3 times that of the sediments (Figure 4), with the
427 gneiss having an average SRMS of 58 (n = 59), the schists an average of 56 (n =
428 60), while the average SRMS for the sediments is 44 (n = 44).

429

430 There is also variability in the strength of the bedrock as determined from the
431 Schmidt hammer and SRMS assessments along the length of the rivers, although
432 the SRMS values show less variability and a smaller range in values than the
433 Schmidt hammer rebound readings (Figure 5). Additionally, the metamorphic rocks
434 have more variable rebound strengths than the sedimentary units. The greatest

435 variability in rebound strength is evident along the Akçapınar (Figure 5A), Sartçay
436 and the Bozdağ (Figure 5B and C) in the west of the range, whereas the values from
437 the metamorphic rocks of the Kabazlı, Yeniköy and Badınca rivers (Figure 5D – E) in
438 the east are more similar. This observation potentially reflects the dominance of
439 schist in the bedrock of the western rivers, whereas gneiss is more prevalent in the
440 east. However, the overall rebound strength of the rocks is consistent with the
441 regional mean. By contrast, rebound strength and SRMS values for sedimentary
442 rocks exhibit remarkable downstream consistency with limited variability within
443 individual rivers. Despite the observed downstream variation, it is evident that the
444 metamorphic bedrock has greater rock strength than the sedimentary strata. We
445 therefore proceed with these two lithological classes in the analysis below.

446

447 **4. 2 Downstream Evolution in Stream Power**

448 The downstream evolution of unit stream power in each channel system gives a
449 measure of how effectively each channel is keeping pace with uplift on the fault. For
450 example, in the Akçapınar River (Figure 6a, Table 1) the stream power grows
451 progressively from minimum values of $< 100 \text{ Wm}^2$ in the headwaters to reach-
452 averaged values of $\text{ca } 1500 \text{ W/m}^2$, within the knickzone, 2 km from the active fault. At
453 the fault, stream powers reduce to $\text{ca. } 500 \text{ W/m}^2$ although the metamorphic bedrock
454 is weaker and more fractured in this location (Figure 5a). By contrast, on the Sartçay
455 River the stream power at the fault is only 105 W/m^2 in the sedimentary rocks near
456 the fault (Figure 6b, Table 1). However, maximum and reach averaged stream
457 powers of 2500 and 1600 W/m^2 respectively occurs 5 km upstream from the active
458 fault in metamorphic rocks (gneisses), while values decrease after this peak towards
459 the fault, corresponding to the river channel eroding through a sliver of schist and
460 weaker sedimentary lithologies (Fig. 5b). Similar observations of peak stream power
461 in the metamorphic rocks upstream of the fault, but within the knickzone can be
462 made on the Bozdağ, Kabazlı, Yeniköy, Badınca Rivers (Figs. 6c-f).

463

464 As the sedimentary units have lower hardness and SRMS values than the
465 metamorphic rocks in the Bozdağ Range we deduce rock type has a significant
466 effect on the stream powers produced. While peak and maximum reach averaged
467 stream power differ between the six rivers, which experience different fault uplift
468 rates, a common trend in all the rivers is observed. Stream power significantly

469 increases downstream of the knickpoint in the hard metamorphic rocks, and then
470 declines rapidly at or towards the fault where soft sedimentary rocks are
471 encountered (Figure 6b-f). We also note that the rivers have low ($<100 \text{ W/m}^2$) stream
472 powers upstream of the knickpoint within the channel, even though the bedrock is
473 hard. Moreover, we directly observe that the rivers are keeping pace with fault uplift
474 at the basin margin due to the absence of fault scarps in the channels.

475
476 These data strongly suggest that bedrock lithology influences the downstream
477 development of stream power. In particular, where the Akçapınar River is compared
478 to the other rivers there is a measurable difference between the stream power at the
479 fault (Table 1) because the Akçapınar incises through schist and gneiss along its
480 whole length while the other rivers incise through weak clastic sediments in varying
481 proportions of catchment area and channel length. Below we investigate this
482 observation further.

483

484 ***4.3 Influence of Rock Strength on Stream Power***

485 From field observations and rock strength measurements using a Schmidt hammer
486 and the SRMS, it is obvious that the clastic sedimentary rocks within the catchment
487 are on average weaker and more easily eroded than the metamorphic basement
488 (Figs 4, 5). In the rivers where there are sedimentary lithologies upstream of the
489 active fault there is also a significant reduction in the stream power in this zone
490 (Figure 6). Typically for Schmidt hammer rebound values of ca. 20 and SRMS values
491 of ca. 50, stream power is $< 330 \text{ W/m}^2$. However, when stream power is compared
492 to Schmidt hammer readings of the metamorphic bedrock, especially immediately
493 upstream of the lithological boundary where maximum stream power is calculated,
494 there is a broad trend of increasing stream power with increasing Schimdt hammer
495 values (circles, Figure 7a). A similar but weaker relationship can be observed for an
496 increase in stream power with higher SRMS values.

497

498 If we assume that the maximum reach-averaged stream powers developed in both
499 the metamorphic and sedimentary rocks are sufficient to balance uplift on the active
500 fault (Equation 3), we can use this data (and associated uncertainty) to make a first
501 order estimate of k_b as a function of rock strength (Equation 2, Figure 7b). This
502 calculation suggests that $k_b = 2.2 - 6.3 \times 10^{-14} \text{ m s}^2 \text{ kg}^{-1}$ in the metamorphic rocks and

503 $k_b = 1.2 - 15 \times 10^{-13} \text{ m s}^2 \text{ kg}^{-1}$ for the sedimentary units – a difference of up to 30
504 times in the value of implied bedrock erodibility is obtained for a less than 3 fold
505 variation in Schmidt hammer rebound value. This result suggests that relatively
506 small differences in rock mass strength measurements can translate into very large
507 differences in long-term bedrock erodibility. However, we also note that for any given
508 rock strength there can be up to a > 3 fold difference in the calculated stream power
509 at the locality of measurement (Figure 7a). For example, on the Bozdağ River the
510 highest stream power developed in a ca. 2 km metamorphic reach is 2895 Wm^{-2} with
511 an average Schmidt hammer reading of 45 and average SRMS of 65. Yet, for the
512 Badınca the Schmidt hammer and SRMS values are similar to the Bozdağ at 46 and
513 66, respectively, but the maximum stream power developed is 962 Wm^{-2} (Table1).
514 Consequently, rock strength is not the only parameter that influences the stream
515 power of the river. Given that the Bozdağ is experiencing throw rates of 2 mmyr^{-1}
516 and the Badınca throw rates of 0.7 mmyr^{-1} it is expected that the variation in throw
517 rates along strike of the active fault is also a significant control on the stream power.

518

519 **4.4 The Influence of Throw Rate on Stream Power**

520 Assuming river incision is keeping pace with tectonic uplift in the vicinity of the fault,
521 we compare stream powers within the knickzone upstream of the fault to evaluate if
522 along-strike variations in throw rate can account for the differences in stream power
523 observed, given that rock strength variation alone does not fully explain the
524 variations in unit stream power observed. As the clastic sediments within the
525 catchment are weaker and more erodible than the metamorphic bedrock, we first
526 compare the effect of tectonic throw rate on the maximum reach average stream
527 power in the knickzone (Figure 8a) as in all cases this value lies within the harder
528 metamorphic lithology. The Bozdağ Range is an uplifting horst block likely with
529 broadly uniform uplift (Kent et al., 2016), so the uplift rate experienced by the river in
530 the area of peak stream power should be representative of the magnitude of the
531 throw rate at the fault. If the tectonic uplift occurring on the southern margin of the
532 Gediz Graben is the only factor determining the stream power of the rivers where
533 metamorphic lithologies dominate (and hence their incision can be described a
534 simple 'detachment-limited' erosion model), it would be expected that the stream
535 power would be proportional to the estimated uplift rate. The present-day throw rates
536 determined by Kent et al. (2017) show a ~ 3 fold difference in the throw rates

537 between the rivers, so in theory it should be possible to observe a similar scaling in
538 the stream power.

539

540 We do observe a positive correlation between throw rate and the maximum stream
541 power developed in the metamorphic rocks in the knickzone upstream of the fault
542 (Figure 8a). We note there is a factor of three difference in maximum reach
543 averaged unit stream power between the river eroding the slowest slipping part of
544 the fault block (the Badınca; 972 W/m^2) and the fastest slipping (Bozdağ River; 2985
545 W/m^2), for a throw rate of 0.7 and 2.1 mmyr^{-1} , respectively. However, the Akçapınar,
546 Kabazlı and Sartçay rivers have somewhat lower stream power than a simple linear
547 scaling might suggest between these end-members. Some of this variation may
548 reflect bedrock erodibility differences, captured imperfectly in the SRMS and Schmidt
549 hammer rebound measurements shown in Figure 7a and/or a non-linear relationship
550 between stream power and throw rate upstream of the fault. The Yeniköy (white
551 symbol, Figure 8a) is an outlier to this trend. The peak reach-averaged stream
552 power developed in the metamorphic rocks along the Yeniköy is approximately a
553 factor of 2 smaller than for other rivers subjected to similar uplift rates (Figure 8a;
554 Table 1). We note there is a significant amount of clastic sediment observed in this
555 river and we return to this observation in the discussion.

556

557 The same type of analysis as above can be undertaken to examine the impact of
558 sedimentary bedrock on the stream power of the rivers near the fault (Figure 8b).
559 Where sedimentary rocks are present in the 2 km upstream of the fault (all rivers bar
560 the Akçapınar), the stream power is markedly lower than in the metamorphic
561 lithologies ($< 330 \text{ Wm}^{-2}$). But as figure 7a shows, the values do not appear to scale
562 with the (relatively limited) variation in Schmidt hammer rebound or SRMS values.
563 However, the reach-averaged stream power of rivers incising through sedimentary
564 bedrock in the 2 km upstream of the fault also does not increase with throw rate,
565 contrary to the way that the peak reach-averaged stream power does in the
566 metamorphic rocks (Figure 8b). In four of the rivers, the measured stream power
567 within the sediments are lower than the stream power of the Badınca River, even
568 though this is the river crossing the part of the fault with the lowest slip rate; only in
569 the Sartçay River (slipping at 1.85 mm/yr) is higher stream power than the Badınca
570 recorded. Point measurements of stream power in the sedimentary units at the fault

571 lead to identical conclusions, with no scaling between stream power and fault throw
572 rate (supplementary material, Fig. S3).

573

574 Consequently, while fault throw rates exert a first order control over the stream
575 power within the metamorphic lithology present upstream in all rivers, which is
576 consistent to first order with a detachment limited model of erosion. The variations in
577 unit stream power do not simply explain the ability of the rivers to cut across the
578 sedimentary rocks in the footwall of the Gediz Graben where throw rates vary by up
579 to a factor of 3, we return to this observation in the discussion below (c.f. Cowie et
580 al., 2008).

581

582

583 **5 ANALYSIS AND DISCUSSION**

584

585 ***5. 1 Bedrock erodibility and lithological controls on fluvial incision***

586 Our results demonstrate that both bedrock strength and fault throw rate influence the
587 magnitude of unit stream power developed upstream of active fault segments in the
588 Gediz Graben of Turkey. It is well established that bedrock lithology should be an
589 important a factor in determining the rate and style of landscape response to a
590 change in relative base level (Stock and Montgomery, 1999; Reneau, 2000; Bishop
591 et al., 2005; Brocard and van der Beek, 2006; Anthony and Granger, 2007; Cook et
592 al., 2009, Kent et al., 2017, Yanites et al., 2017, Zondervan et al., 2020a). However,
593 the precise nature of the control exerted on river response to tectonic perturbation
594 remains an outstanding issue (Crosby and Whipple, 2006; Anthony and Granger,
595 2007; Haviv et al., 2010; Whittaker and Boulton, 2012). This study provides further
596 insight into the nature of this control and provides new constraints on the links
597 between theoretical estimates of bedrock erodibility and physical measurements of
598 rock mass strength.

599

600 The magnitude and range of bedrock erodibilities in published studies varies widely,
601 both as a function of study location and the erosion law chosen (Stock and
602 Montgomery, 1999; Zondervan et al., 2020a). In this study, assuming a unit stream
603 power erosion law in which channel width and discharge are derived directly from

604 field data, we find that there is a significant difference in k_b between the two broad
605 rock types, where $k_b = 2.2 - 6.3 \times 10^{-14} \text{ m s}^2 \text{ kg}^{-1}$ in the metamorphic rocks and $k_b =$
606 $1.2 - 15 \times 10^{-13} \text{ m s}^2 \text{ kg}^{-1}$ for the sedimentary units. These values are comparable to,
607 but greater than the bedrock erodibilities of $1.8 \pm 0.3 \times 10^{-14}$ and $6 \pm 2 \times 10^{-15} \text{ m s}^2$
608 kg^{-1} recently derived by Zondervan et al. (2020a) for conglomerates and limestones,
609 respectively, upstream of the East Eliki fault on the southern margin of the Gulf of
610 Corinth, Greece, in a study where a similar methodology has been used. While it is
611 notable that order of magnitude differences in bedrock erodibility are obtained
612 between the metamorphic rocks and the sedimentary units for relatively limited
613 differences in measured rock strength, it is equally important to observe that these
614 bedrock erodibilities are up to 10 orders of magnitude less than values that have
615 been used in numerical modelling studies where units of erodibility can be compared
616 directly (Roy et al., 2015; Yanites et al., 2017). Because landscape response times
617 to tectonics, expressed for instance in terms of knickpoint retreat rates, scale directly
618 with bedrock erodibility (e.g. Whittaker and Boulton, 2012), these differences imply
619 landscape response times that are geomorphically long (i.e. multi-million year
620 timescales for catchments of the scale studied here). Consequently, we argue that
621 our results underline the importance of deriving field constraints on bedrock
622 resistance to erosion much more widely.

623

624 The impact of lithological variation on bedrock erodibility between the sedimentary
625 and metamorphic rock types investigated here has to be contextualised with respect
626 to the 'many orders of magnitude difference' in rock resistance to fluvial incision that
627 some authors have obtained, using a more generalised form of the stream power
628 erosion law (Eq. 1). For example, Stock and Montgomery (1999) studied the effect of
629 lithology on the K parameter in this simple form of the unit stream power model
630 where erodibility values are expressed in units of $\text{m}^{(1-2m)} \text{ yr}^{-1}$. They noted that K
631 varied over five orders of magnitude between mudstones and volcanoclastic rocks
632 from Japan and California (10^{-2} to $10^{-5} \text{ m}^{0.2} \text{ yr}^{-1}$) and granitoids and metasediments in
633 Australia (10^{-6} to $10^{-7} \text{ m}^{0.2} \text{ yr}^{-1}$), although the proportion of the variation owing to
634 climate was unresolved because precipitation-driven discharge scaling with drainage
635 area was not considered. Brocard and Van der Beek (2006) also determined that for
636 bedrock rivers in French Alpine rivers K should fall between $1.8 - 4.7 \times 10^{-5} \text{ m}^{0.4} \text{ yr}^{-1}$
637 for marl and between $1.1 - 3.7 \times 10^{-5} \text{ m}^{0.4} \text{ yr}^{-1}$ for limestone, whereas van der Beek

638 and Bishop (2003) determined that for a river crossing crystalline basement rocks in
639 SE Australia that $K = 7 \times 10^{-7} \text{ m}^{0.4} \text{ yr}^{-1}$. K parameters determined for crystalline rocks
640 on the Pacific Islands of Makira and Guadalcanal also fall in the range 1×10^{-5} to 5
641 $\times 10^{-8} \text{ m}^{0.1} \text{ yr}^{-1}$ (Boulton, 2020). Similarly, the knickpoint retreat rate data of Whittaker
642 and Boulton (2012) imply K of 10^{-6} to 10^{-7} yr^{-1} assuming $m = 0.5$. Comparison with
643 our results expressed in a drainage-area dependent form equivalent to Eq. 1
644 suggests K ca. 10^{-7} yr^{-1} in the metamorphic rocks and values $> 10^{-6} \text{ yr}^{-1}$ in the
645 sedimentary units of the Bozdağ Range. These studies therefore indicate that K
646 does vary markedly worldwide, and even if discharge data are known or derived,
647 allowing for climate differences to be accounted for, our results support Stock and
648 Montgomery's (1999) initial conclusion that the timescale of landscape evolution
649 must vary greatly with lithology. A key area for future work is to link modelling
650 studies with realistic long-term values of bedrock erodibility, ideally derived from
651 locations where the timescale of erosion rates can be derived independently.

652

653 **5.2 Rock mass strength, bedrock erodibility**

654 Bedrock erodibility is not a physical parameter that can be measured directly in the
655 same way as compressive or tensile strength of bedrock, or the density of fractures
656 (e.g. Bursztyn et al., 2015; Zondervan et al., 2020a,b). It is also true that mapped
657 lithological units (or lithological groups as used here) typically can encompass
658 significant variations in local lithofacies and bedrock resistance to fluvial erosion at a
659 map scale (c.f. Yanites et al., 2017). Consequently, our analysis is intended to
660 describe differences in rock resistance to incision between broad rock types at a
661 coarse resolution. In this study, a regional variation in Schmidt hammer rebound
662 values of a factor of ca. 2 between clastic sedimentary rocks and metamorphic
663 schists and gneisses produces greater than an order of magnitude difference in k_b
664 (Figure 7b). At a finer resolution, small variations in SRMS index appear to be linked
665 to marked variations in unit stream power developed upstream of the faults (Figures
666 6 and 7a). The length scale over which such variations in rock strength matter for
667 governing landscape evolution over 10^5 to 10^6 years remains unresolved. Variations
668 in fracture density have been clearly linked to the local or reach scale morphology of
669 rivers (see Scott and Wohl, 2019 for an extensive review), yet at large scales (10^1 -
670 10^3 km) numerical analyses suggest that the spectral power in river long profiles is
671 not controlled by lithology (Roberts, 2019). Here we suggest that a broad grouping of

672 two main lithotypes on the 1 to 10 km scale is sufficient to illustrate the important role
673 of lithology in driving landscape response to tectonics for a fault block where the
674 length scale of the footwall rivers is of comparable magnitude to the spatial variations
675 in rock strength.

676

677 Nonetheless, a marked variation in bedrock erodibility is calculated in some
678 instances, even where Schmidt hammer rebound values are similar (Figure 7).
679 These observations raise the question of what are the most appropriate physical
680 measures of rock strength that correlate with estimates of bedrock erodibility derived
681 from field examples of landscape incision over timescales of 10^4 to 10^6 years. The
682 Schmidt hammer is a quick, portable and convenient tool to estimate effective
683 compressive strength in a field setting (c.f. Zondervan et al., 2020a), but theoretical
684 and observational studies suggest that tensile rock strength might more effectively
685 represent the resistance of bedrock to the impact of clasts in a river as previously
686 noted (Sklar and Dietrich, 2001; Bursztyn et al., 2015). However, lab measurements
687 on individual rock samples do not typically capture the full range of factors such as
688 joint density and orientation that influence rock strength and bedrock erodibility at
689 scales greater than the sample size (e.g. greater than a few centimetres), while it is
690 hardly practical to quantify joint densities at high resolution across an entire
691 catchment that may have a drainage area of tens or hundreds of square kilometres.
692 The SRMS index does include a range of jointing parameters, but only in a semi-
693 quantitative sense, and it was not originally designed to measure bedrock resistance
694 to fluvial incision. In this study there appears to be little correlation of Selby jointing
695 sub-parameterisation with down-system distributions of stream power
696 (Supplementary Data Tables S1 to S6). At the same time, relatively small variations
697 in measurements of physical rock strength here translate into order of magnitude
698 differences in long-term bedrock erodibility, suggesting that landscapes upstream of
699 active normal faults are highly sensitive to rock strength differences. We also note
700 that in this study the SRMS index in general shows less variability in rock strength
701 between lithological units than our Schmidt hammer rebound values alone, so
702 although this rock strength parameterisation is widely used (c.f. Whittaker et al.,
703 2007b; Zondervan et al., 2020), we would suggest that it is not a particularly
704 sensitive measure of long-term bedrock erodibility captured by k_b . Consequently, an
705 outstanding area of further work is to probe further what physical and *in situ*

706 measures of rock strength correlate effectively with estimates of bedrock erodibility
707 derived from constraints on long-term landscape evolution over a range of spatial
708 scales, and to further investigate if composite measures of rock strength such as the
709 SRMS can be adapted to better capture differences in long-term substrate resistance
710 to fluvial erosion.

711

712 **5.3 Stream Powers, Throw Rates and Sediment Transport Capacity**

713

714 In this study, for the clastic sedimentary units upstream of the fault, stream power is
715 markedly lower than in the metamorphic rocks for all six rivers (Figure 7a). This
716 observation is consistent with qualitative observations of the friable nature of these
717 lithologies compared to the metamorphic rocks, and with quantitative observations of
718 lower Schmidt hammer rebound and SRMS values. However, our results also show
719 that there is not a clear correlation between estimated fault throw rate and stream
720 power at the fault in the sedimentary lithologies (Fig. 8b), which is what a simple unit
721 stream power model of fluvial incision would predict for rivers incising uniform
722 lithology. Therefore, our results raise the question of whether a simple detachment-
723 limited erosion law is too basic to characterise the time-integrated incisional
724 dynamics of these rivers.

725

726 One consideration could be that the Schmidt hammer rebound values and SRMS
727 measures are not very good at picking up relative variations in bedrock strength in
728 soft lithologies, particularly in cases where the hammer does not measure a marked
729 rebound. However, an alternative possibility is that the nature of the bedrock itself
730 also influences the long-term erosional dynamics of the channels. A version of this
731 effect has previously been noted by Cowie et al. (2008), in a comparative study of
732 rivers crossing active normal faults each with broadly similar lithologies near the Gulf
733 of Evia, Greece and in the Central Apennines. In this research, it was found that
734 increasing bed shear stresses, calculated from hydraulic geometry measurements,
735 did not scale with increased fault throw rates, despite clear evidence of the individual
736 rivers keeping pace with the rate of faulting. Instead, the authors argued that
737 increasing sediment flux down-system, derived from outcropping conglomerates,
738 provided a 'tools' effect that markedly boosted the ability of the river to incise across
739 the fault.

740

741 In our case all but one of the rivers are incising across Neogene poorly-consolidated
742 clastic sedimentary units near the faults, rocks which were originally of fluvial or
743 alluvial fan origin. Based on our results in Figure 8b, we explore the additional
744 hypothesis that in the downstream reaches of five of the studied rivers, the friable
745 clastic bedrock substrate provides an abundant supply of sediment and in effect, the
746 river may be limited in its ability to incise by its capacity to transport the poorly
747 cemented clastic material that effectively forms the substrate. We do not know the
748 long term sediment supply relative to the discharge history of the river, but if this is
749 the case, then we might expect the rivers to be configured such that their sediment
750 transport capacity, Q_c , is greater where fault throw rates are larger. One way to
751 check if this true for the sedimentary bedrock units is to use the channel hydraulic
752 geometry data to determine a dimensional Meyer-Peter-Muller bedload transport
753 capacity in kg s^{-1} which can be calculated (Whittaker, 2007) as:

$$754 \quad Q_c = \left[\rho_s \left(\frac{\rho_s - \rho_w}{\rho_w} g D^3 \right)^{\frac{1}{2}} C (\tau^* - \tau_c^*)^{\frac{3}{2}} \right] * W \quad (4)$$

755 where ρ_s and ρ_w are the densities of sediment and water, respectively; D is the
756 median grain-size; τ^* is the dimensionless shear stress (estimated from $\frac{HS}{D\rho_x}$ where ρ_x
757 represents the excess sediment density of 1.65); τ_c^* is the critical Shields stress for
758 entrainment of gravel, for which we use a constant of 0.06 (e.g. Mueller and Pitlick,
759 2005) and $C = 3.97$ based on the re-analysis by Wong and Parker (2006). To do
760 this, we require estimates of grain size in the rivers near the fault. We do not have
761 data on sediment calibre for the majority of our field sites, but we have some photo
762 estimates for 5 sites in the downstream reaches of the rivers near the fault (Methods,
763 Supplementary Material Fig. S3 and Fig. S4). We take this data to be representative
764 of grain size, to first order, for the 2 km reaches upstream of the range bounding
765 fault.

766 This additional analysis (Figure 9) indicates that instantaneous sediment transport
767 capacity, averaged for a 2 km reach upstream of the fault in the clastic sedimentary
768 units, is strongly dependent on fault throw rate. Transport capacities vary from ca.
769 100 kg/s for the river crossing the slowest slipping section of the fault to > 300 kg/s
770 for the river crossing the fastest slipping section of the fault (which has a throw rate
771 three times greater). The Yeniköy River, which had relatively low stream powers in

772 both the sedimentary and metamorphic rocks fits well with the trend of the data.
773 These data are consistent with the hypothesis that the rivers incising into poorly
774 consolidated clastic bedrock are configured to transport more sediment where fault
775 slip rates are greater. However the relationship between sediment transport capacity
776 and uplift rate shown in figure 9 is not linear and we do not have any information on
777 the frequency and the magnitude of sediment transport events or the exact locus of
778 sediment sourcing, although it was likely wetter in the region at the last glacial
779 maximum (Kent et al., 2017). In addition, we do not know the ratio of the long term
780 sediment flux to the transport capacity, which potentially may have varied over
781 glacial-interglacial timescales and will have determined the time-averaged export of
782 sediment from the catchments.

783 Nevertheless, a sediment transport rate of e.g. 200 kg/s would enable a river to
784 discharge ca. 7000 m³ of sediment if active at bankfull capacity for one day a year,
785 given $\rho_s = 2400 - 2600 \text{ kg/m}^3$, which would represent 'erosion' rates of > 1.4 mm/yr if
786 the sediment was uniformly derived from a plausible clastic bedrock catchment area
787 of e.g. 5 km². The rivers in this study are not gauged, but based on this analysis, we
788 interpret our results to indicate that even where rivers are documented to incise
789 bedrock upstream of fast-slipping faults, the presence of clastic sedimentary rocks
790 can influence fluvial responses to tectonics, not just in terms of absolute rock
791 strength relative to 'harder' lithologies, but also by supplying channels with self-
792 derived clastic detritus that can influence long-term erosional dynamics. Resolving
793 the nature of this control more generally requires a system in which sediment
794 discharge histories can be determined independently and where multiple bedload
795 grain size data points are available.

796

797 **CONCLUSIONS**

798 The Gediz Graben in Western Turkey is a Neogene extensional graben that
799 experienced an increase in fault slip rates ~ 0.8 Ma, causing a wave of river incision
800 to propagate through rivers draining transversely across the faulted basin margin.
801 These rivers incise two main lithological groups – metamorphic gneisses and schists
802 and more friable Neogene clastic sedimentary formations. The channels are
803 responding transiently to this tectonic perturbation by changing their geometry and
804 planform shape up and downstream of knickpoints in the rivers.

805

806 For six selected rivers representing different bedrock lithologies and fault slip rates,
807 unit stream powers were calculated and bedrock strength measured using field data.
808 Similar downstream trends are seen in each case. In the headwaters the stream
809 powers remain low (under 150 W/m^2) until the knickpoint. Downstream of the
810 knickpoint the stream powers rise significantly to peak values found within the
811 metamorphic bedrock. However, where the channel incises the Neogene clastic
812 sediments downstream towards the active fault, stream powers decrease markedly.
813 Schmidt hammer rebound values and Selby rock mass strength (SRMS)
814 measurements demonstrate that the metamorphic rocks are around 2x harder than
815 the sediments while the stream powers in the sedimentary lithologies are
816 significantly lower than in the metamorphic basement. Based on our observation
817 that the rivers are keeping pace with the basin bounding fault and independent
818 estimates of fault throw rate, we use this data to calculate the bedrock erodibility k_b
819 of the two main lithological groups identified. Overall, k_b shows >30 fold variation
820 between sedimentary and metamorphic rocks corresponding to a <3 fold variation in
821 Schmidt hammer rebound strength, suggesting that small differences in rock
822 strength translate into significant differences in long-term bedrock erodibility and
823 landscape response time. We also note that our field-derived values for k_b are
824 smaller than values that have been used in modelling studies or derived from other
825 field measures, highlighting the need for further research into this poorly constrained
826 parameter.

827

828 Although our data show that lithology plays a key role in modulating stream powers
829 upstream of the basin bounding fault, we do find that the rivers exhibit peak stream
830 powers in the stronger and more resistant metamorphic rocks and these values
831 scale with throw rate, suggesting that where the rivers erode metamorphic lithologies
832 the detachment-limited models of erosion can predict bedrock incision rate. We also
833 demonstrate that in the reaches where the rivers incise the poorly consolidated
834 clastic bedrock, unit stream power shows limited correlation to the throw rate.
835 However, when the bedload sediment transport capacity of these rivers is compared
836 to the throw rate there is a trend of increasing transport capacity with increasing
837 throw rate on the fault, and we suggest that the supply of easily erodible clastic
838 material directly from the bedrock has a profound effect on the behaviour of our

839 studied rivers. Consequently, we argue that bedrock type influences landscape
840 response to faulting, not just in terms of the erodibility of the substrate, but also in
841 terms of the dynamics of the long-term incision process itself.

842

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848

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1117

1118 FIGURE CAPTIONS

1119

1120 Figure 1. A) Map of the Aegean region showing major plate boundaries and the
1121 location of the area of the Western Anatolian Extensional Province (WAEP) shown in
1122 B; B) Shaded topographic map of the central part of the Eastern Anatolian
1123 Extensional Province overlain with major active normal fault systems and associated
1124 grabens (BMG – Büyük Menderes Graben; KMG – Küçük Menderes Graben; GZG –
1125 Gediz Graben; DG – Demirci Graben; GG- Gördes Graben). Topographic profiles
1126 shown to the right cross the Bözdağ block, a horst located between the GZG and
1127 KMG; the location of the profiles are indicated on the map. C) SRTM 30m DEM
1128 overlain with simplified lithology, major faults, catchments of studied rivers and slip
1129 rates at the range front where those rivers discharge (map modified and data from
1130 Kent et al., 2017).

1131

1132 Figure 2. Photographs illustrating key features of the fluvial geomorphology. A)
1133 Incised metamorphic bedrock reach on the Akçapınar River; B) wide fluvial plain
1134 near the basin bounding fault on the Yeniköy River; C) View of uplifted and eroding
1135 Cenozoic sedimentary units typical of the incised landscape downstream of the
1136 knickpoints; D) View of the upper reaches of the Akçapınar river showing ~ 300 m of
1137 incision into a low relief pre-incision landscape.

1138

1139 Figure 3. River long profiles showing main lithological groups found in the footwall of
1140 the active fault, the location of the range front fault and the tectonic knickpoint (star)
1141 for the A) Akçapınar, B) Sartçay, C) Bözdağ D) Kabazlı, E) Yeniköy and F) Badınca
1142 Rivers

1143

1144 Figure 4. Violin plots showing the variation in Schmidt hammer and SRMS data for
1145 the three main lithological groups. Note: the violin plot shows the probability density
1146 of data at each value, and incorporates a box and whisker plot; where the box
1147 indicates the upper/lower quartiles, the vertical lines the data range and the
1148 horizontal line the data median. Solid circles indicate outliers in the data. The
1149 median for the Schmidt hammer data for the sediments overlaps with the lower
1150 quartile.

1151

1152 Figure 5. Downstream variation in SRMS and Schimdt hammer rebound strength for
1153 each river, with the bedrock type, fault and knickpoint location shown in the bar
1154 above for the A) Akçapınar, B) Sartçay, C) Bözdağ D) Kabazlı, E) Yeniköy and F)
1155 Badınca Rivers.

1156

1157 Figure 6. Downstream variation in unit stream power measured at each field site
1158 (small squares) and reach-averaged stream power (circles) for the A) Akçapınar, B)
1159 Sartçay, C) Bözdağ D) Kabazlı, E) Yeniköy and F) Badınca Rivers. As in previous
1160 graphs the location of the fault is shown, as well as the knickpoint, knickzone and the
1161 extent of metamorphic and sedimentary bedrock.

1162

1163 Figure 7. A) Graph of unit stream power against Schmidt hammer rebound strength
1164 and SRMS for different lithologies and river reaches. Note the overall trend of
1165 increasing stream power with higher strength indices. B) Calculated bedrock
1166 erodibility k_b , for the sedimentary lithologies (white circles) and metamorphic rocks
1167 (black circles); note that the two populations of data fall in distinct groups differing by
1168 an order of magnitude.

1169

1170

1171 Figure 8. A) Graph of maximum reach-averaged stream power in the metamorphic
1172 basement rocks against fault throw rates where the rivers enter the Gediz Graben
1173 (taken from Kent et al. (2017)). Note that the Yeniköy (in white) is an outlier to the
1174 clear trend of increasing stream power with higher throw rates. B) Reach-averaged
1175 stream power in the sedimentary units within 2 km of the fault of the fault. The
1176 Akçapınar river is not shown as it only incises gneisses and schists.

1177

1178 Figure 9. Meyer Peter Muller (MPM) bedload transport capacity against fault throw
1179 rate, averaged for the sedimentary units 2 km upstream of the fault. Greater fault
1180 throw rate is associated with higher maximum transport capacities. Error bars
1181 together represent one standard deviation in values across all measuring sites within
1182 the reach for which the transport capacity is calculated.

1183

TABLE 1

| | Akçapınar | Sart | Bozdag | Kabazli | Yanikoy | Badica |
|--|-----------|----------|----------|----------|----------|----------|
| Throw rate at fault ¹ (mm/yr) | 1.41 | 1.84 | 2 | 1.74 | 1.3 | 0.72 |
| Catchment drainage area, A, at fault (km ²) | 46.7 | 73.2 | 70.8 | 26.5 | 14.5 | 28.8 |
| Max reached-averaged stream power, ω , W/m ² | 1372 | 1657 | 2895 | 1435 | 653 | 962 |
| Max streampower \pm error ($\frac{1}{2}\sigma$) | 216 | 325 | 763 | 587 | 243 | 34 |
| Averaged stream power \sim 2km upstream of fault W/m ² | 788 | 327 | 130 | 154 | 28 | 190 |
| Averaged streampower \pm error ($\frac{1}{2}\sigma$) | 309 | 114 | 67 | 43 | 2 | 34 |
| Schmidt hammer rebound average 2km from fault | 38.9 | 20.0 | 27.8 | 23.3 | 19.7 | 21.7 |
| Schmidt \pm error (σ) | 11.3 | 0.0 | 13.0 | 2.8 | 0.6 | 3.0 |
| Selby RMS 2km from fault | 59.1 | 52.0 | 60.5 | 44.7 | 51.3 | 52.0 |
| Selby \pm error (σ) | 6.6 | 0.0 | 7.3 | 2.1 | 6.9 | 0.0 |
| Schmidt hammer rebound average for max stream power reach | 49.9 | 60.9 | 45.6 | 35.7 | 54.0 | 45.9 |
| Schmidt \pm error (σ) | 12.5 | 3.3 | 7.1 | 13.8 | 6.2 | 7.1 |
| Selby RMS average for max stream power reach | 67.0 | 63.1 | 65.2 | 56.0 | 66.5 | 66.5 |
| Selby RMS \pm error (σ) | 8.7 | 3.2 | 4.2 | 10.8 | 14.8 | 6.8 |
| Calculated k_b metamorphics (m s ² kg ⁻¹) | 3.26E-14 | 3.52E-14 | 2.19E-14 | 3.84E-14 | 6.31E-14 | 2.37E-14 |
| Min k_b metamorphics (m s ² kg ⁻¹) | 2.82E-14 | 2.94E-14 | 1.73E-14 | 2.73E-14 | 4.60E-14 | 2.29E-14 |
| Max k_b metamorphics (m s ² kg ⁻¹) | 3.87E-14 | 4.38E-14 | 2.97E-14 | 6.51E-14 | 1.01E-13 | 2.46E-14 |
| Calculated k_b sedimentary lithologies (m s ² kg ⁻¹) | | 1.78E-13 | 4.88E-13 | 3.59E-13 | 1.49E-12 | 1.20E-13 |
| Min k_b sedimentary lithologies (m s ² kg ⁻¹) | | 1.32E-13 | 3.22E-13 | 2.81E-13 | 1.38E-12 | 1.02E-13 |
| Max k_b sedimentary lithologies (m s ² kg ⁻¹) | | 2.73E-13 | 1.01E-12 | 4.97E-13 | 1.61E-12 | 1.46E-13 |
| MPM transport capacity average \sim 2 km from fault (kgs ⁻¹) | | 168 | 333 | 172.6 | 120 | 111.7 |
| MPM \pm error ($\frac{1}{2}\sigma$) | | 70.8 | 83 | 16.5 | 12 | 17 |

¹estimated uncertainty on throw rate at fault is ± 0.2 mm/y

Figure 1

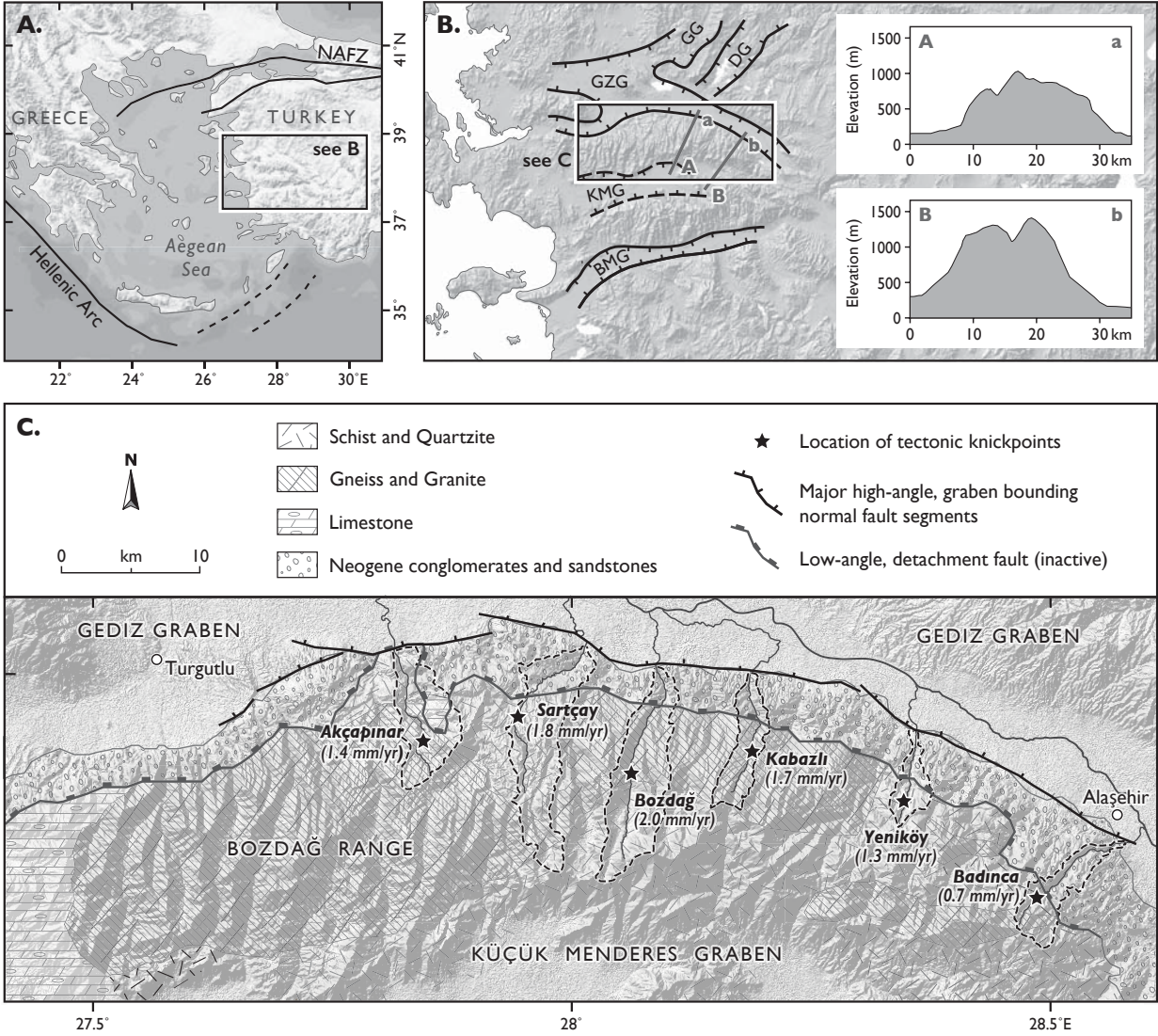


Figure 2

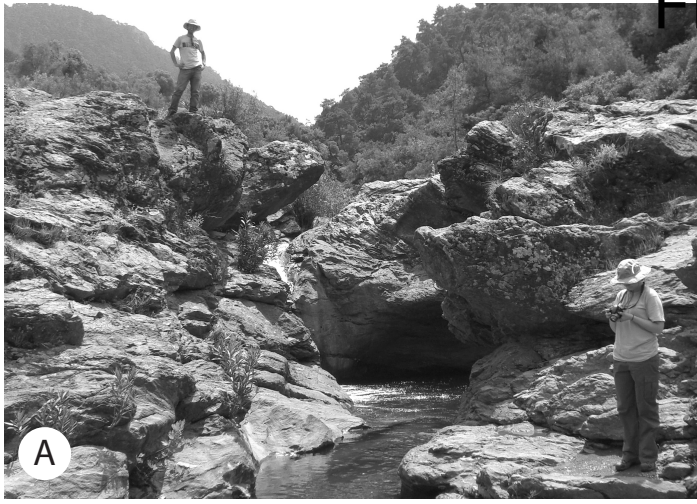


Figure 3

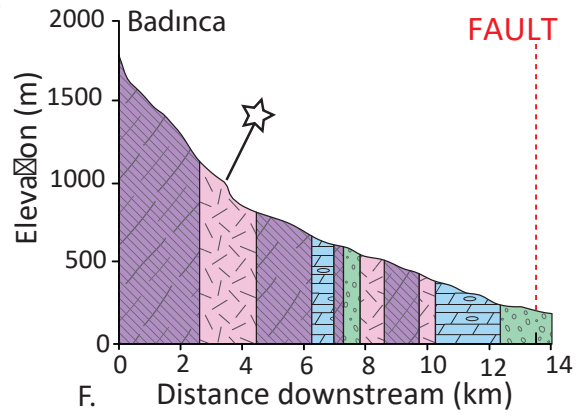
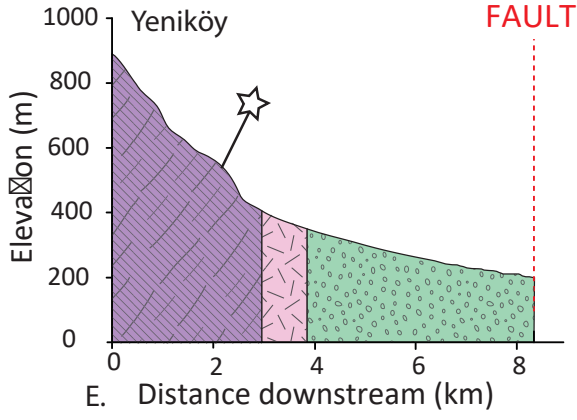
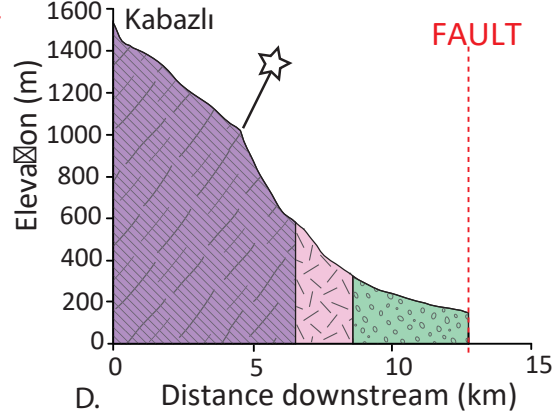
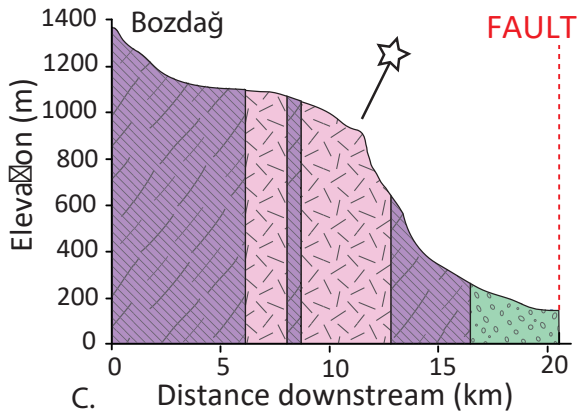
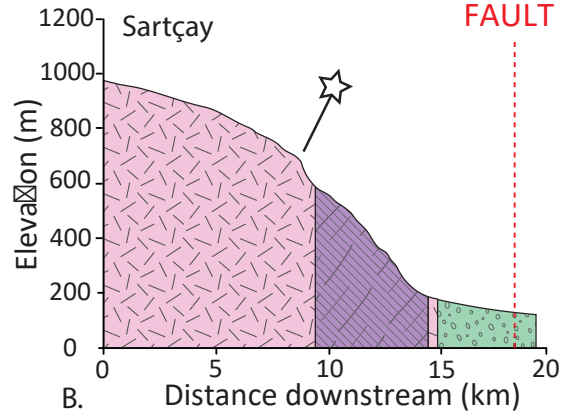
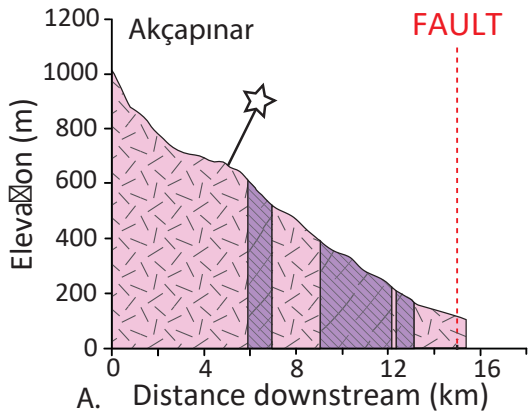


Figure 4

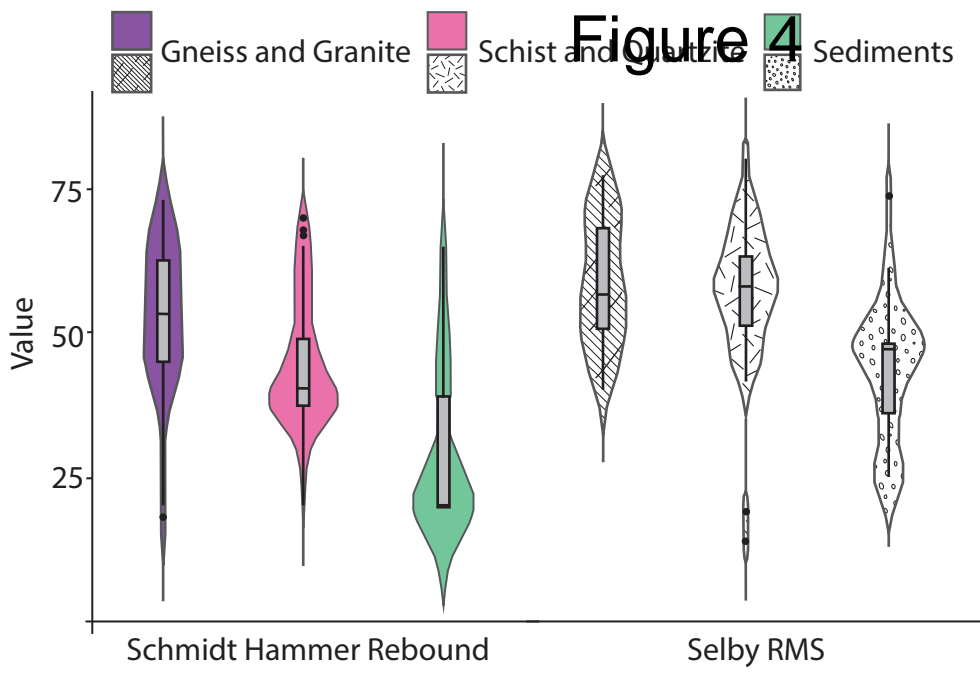


Figure 5

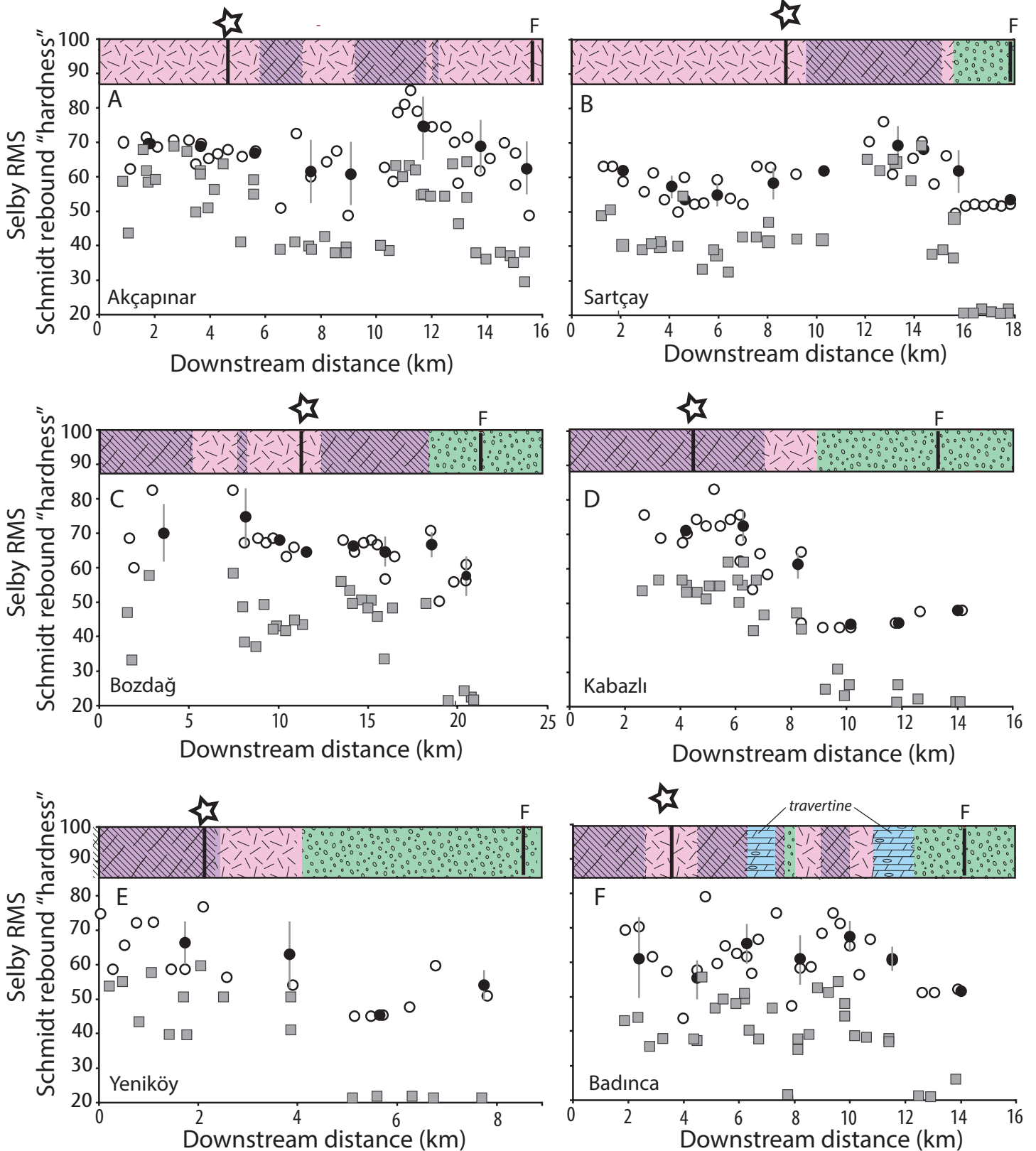
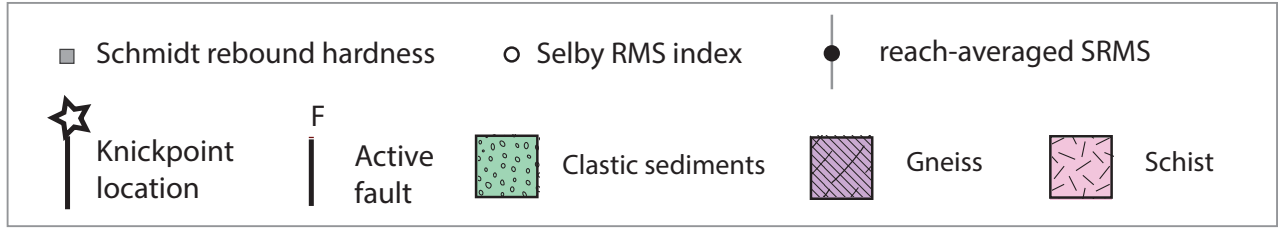


Figure 6

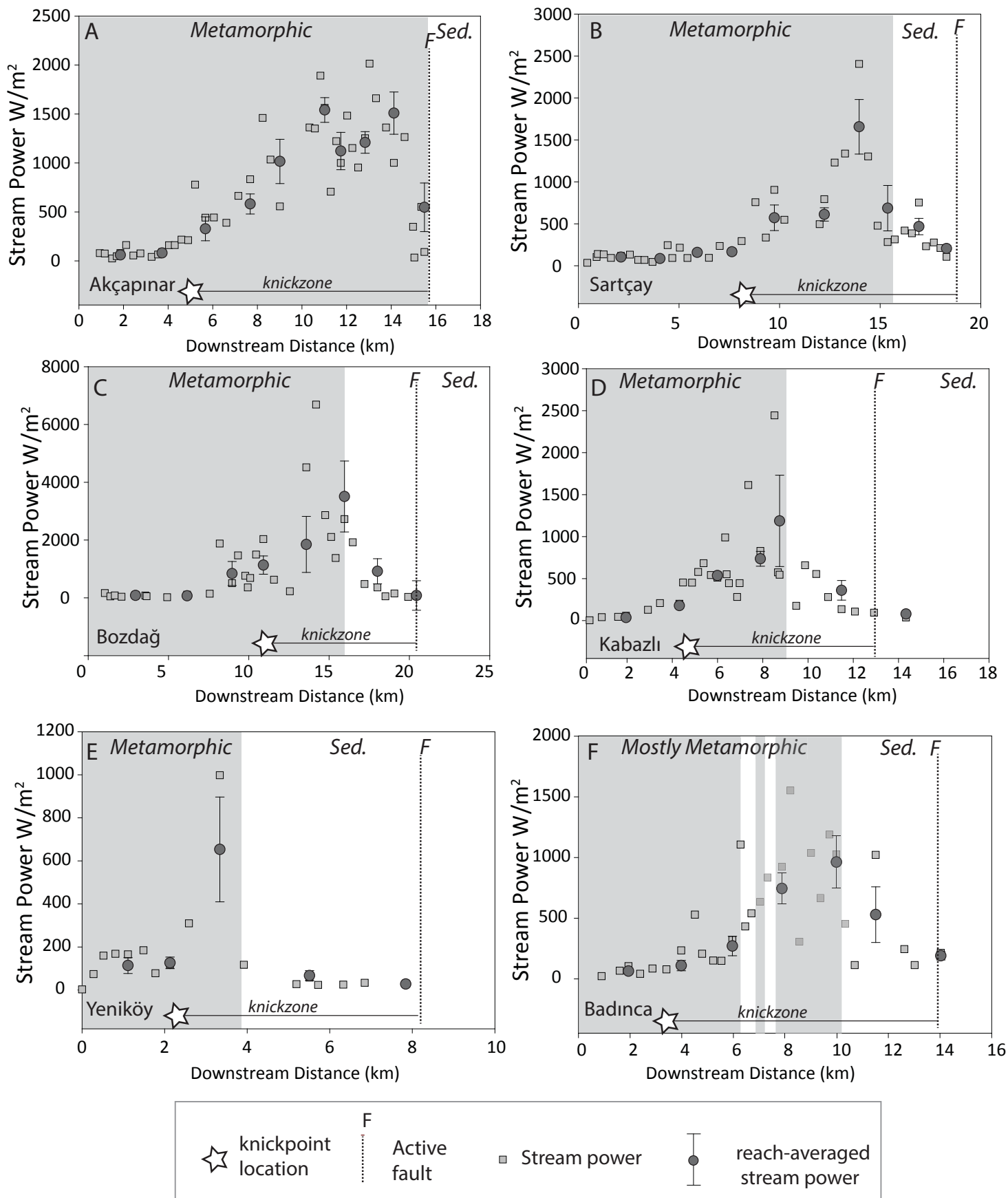


Figure 7

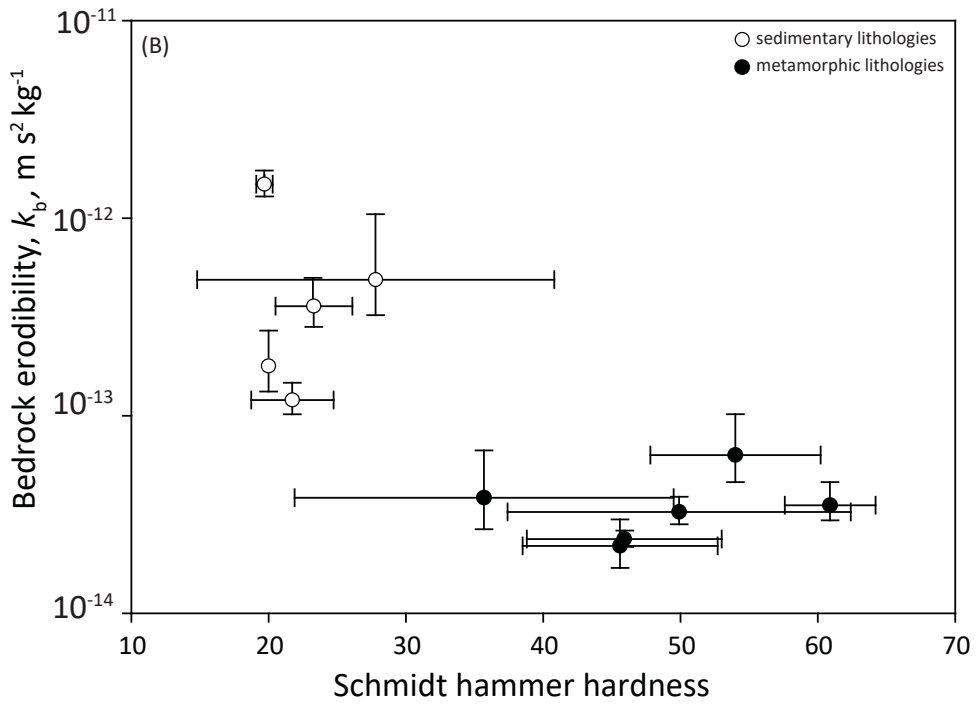
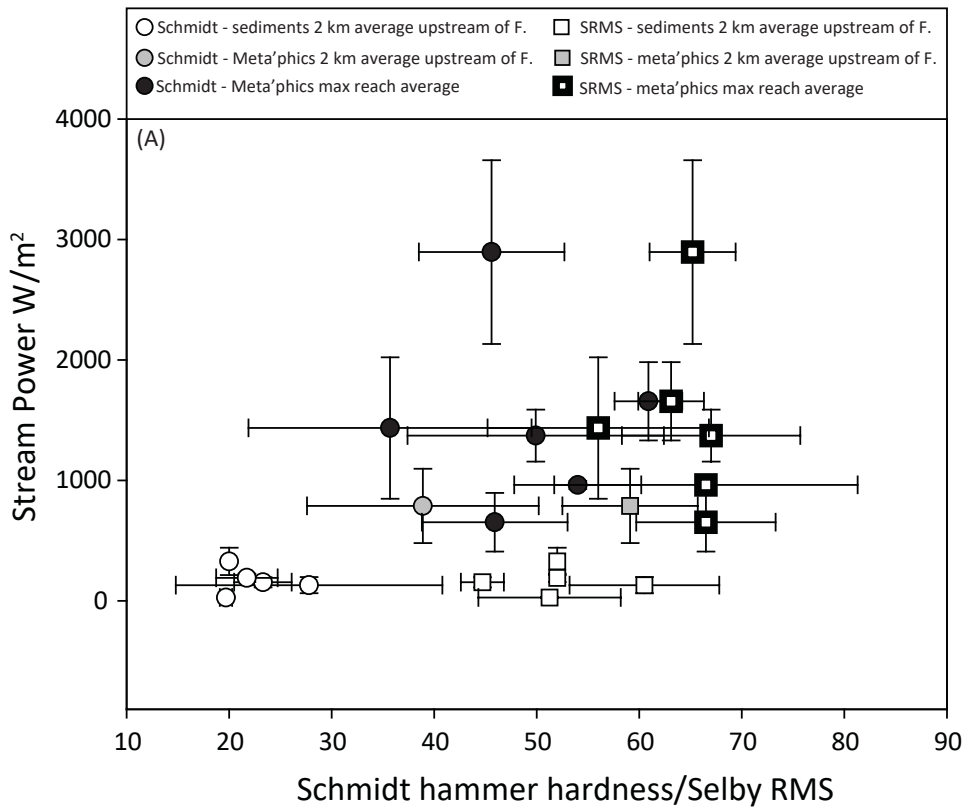


Figure 8

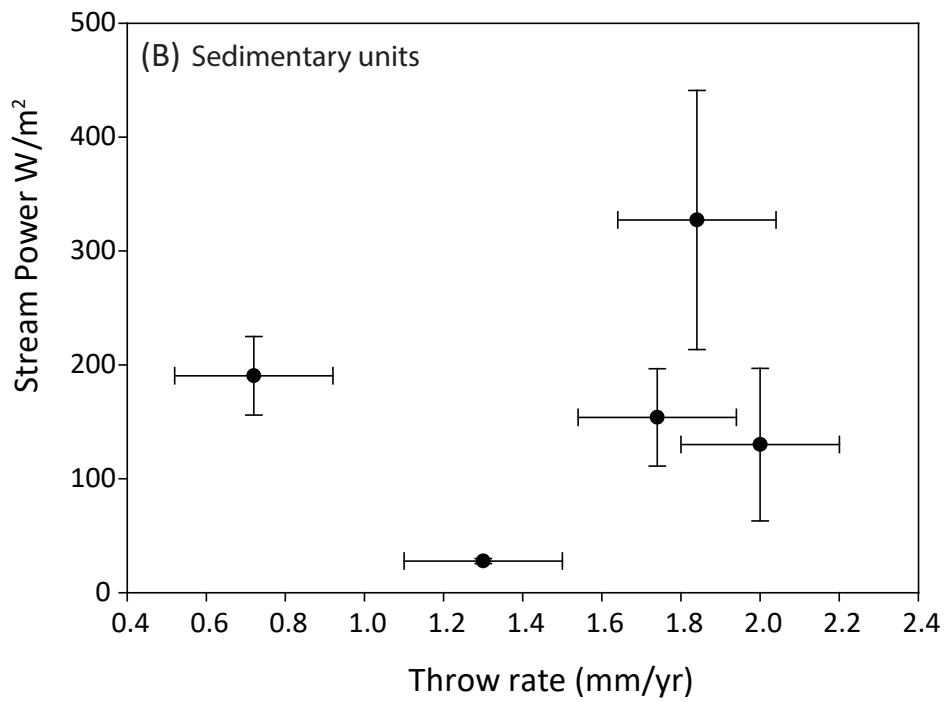
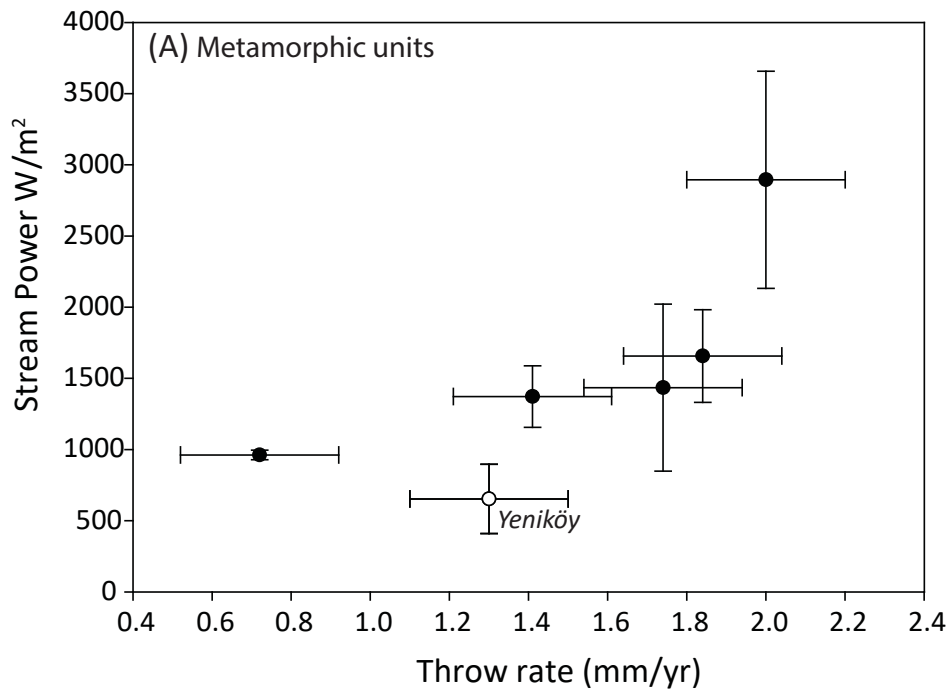


Figure 9

