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6
7 **Complex erosional response to uplift and rock strength contrasts in**
8 **transient river systems crossing an active normal fault revealed by ^{10}Be**
9 **and ^{26}Al cosmogenic nuclide analyses**

10
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24
25 **Abstract**

26 Understanding the influence of bedrock lithology on the catchment-averaged
27 erosion rates of normal fault-bounded catchments, and the effect that different
28 bedrock erodibilities have on the evolution of transient fluvial geomorphology
29 remain major challenges. To investigate this problem, we collected 18 samples for
30 ^{10}Be and ^{26}Al cosmogenic nuclide analysis to determine catchment-averaged

31 erosion rates along the well-constrained Gediz Fault system in western Türkiye,
32 which is experiencing fault-driven river incision owing to a linkage event ~ 0.8 Ma,
33 and has weak rocks overlying strong rocks in the footwall. Combined with existing
34 cosmogenic data, we show that the background rate of erosion of the pre-incision
35 landscape can be constrained as < 92 mMyr⁻¹ and erosion rates within the
36 transient reach vary from 16 – 1330 mMyr⁻¹. Erosion rates weakly scale with unit
37 stream power, steepness index and slip rate on the bounding fault, although
38 erosion rates are an order of magnitude lower than slip rates. However, there are
39 no clear relationships between erosion rate and relief or catchment slope. Bedrock
40 strength is assessed using Schmidt hammer rebound and Selby Rock Mass
41 Strength Assessments; despite a 30-fold difference in erodibility there is no
42 difference in the erosion rate between strong and weak rocks. We argue that for
43 the Gediz Graben the strong lithological contrast effects the ability of the river to
44 erode the bed resulting in a complex erosional response to uplift along the graben
45 boundary fault. Weak co-variant trends between erosion rates and various
46 topographic factors potentially result from incomplete sediment mixing or pre-
47 existing topographic inheritance. These findings indicate that the erosional
48 response to uplift along an active normal fault is a complex response to multiple
49 drivers that vary spatially and temporally.

50 **Keywords:** Turkey, Türkiye, Active faulting, cosmogenic nuclides, rock strength,
51 detachment-limited.

52

53 **1. Introduction**

54 The role of climate, tectonics and lithology on the evolution and form of bedrock
55 (detachment-limited) streams is well known. The effect of tectonics, in particular
56 the effect of variable uplift rates (i.e., Wobus et al., 2006; Kirby and Whipple,
57 2012; Whittaker, 2012; Whittaker and Boulton, 2012), and climate gradients
58 (D'Arcy and Whittaker, 2014; Adams et al., 2020) on the rates and patterns of
59 incision have been widely reported. Until recently the role of lithology and rock
60 strength have attracted less attention and many studies have sought to remove
61 or minimise this variable by choosing study areas with little rock variation (e.g.
62 Miller et al., 2012; Ortega et al., 2013; Regalla et al., 2013; Snyder et al., 2000).
63 However, landscape evolution modelling (Forte et al., 2016; 2020; Perne et al.,

64 2017, Darling et al., 2020; Mitchell and Yanites, 2021) and field investigations at
65 the landscape (Bernard et al., 2019; Zondervan et al., 2020) and catchment scale
66 (Sklar and Dietrich, 2001; Duvall, 2004; Whittaker et al., 2007; Kent et al., 2021;
67 Gailleton et al., 2021; Peifer et al., 2021) have increasingly investigated the
68 importance of lithology on river incision and fluvial geomorphology. Yet, there are
69 still uncertainties in how bedrock properties influence catchment scale erosion,
70 and how such characteristics can be effectively measured in the field.

71 Furthermore, while a number of studies have directly compared catchment-
72 averaged erosion rates (CAER) to bedrock channel properties (i.e., Safran et al.,
73 2005; Harkins et al., 2007; Ouimet et al., 2009; DiBiase et al., 2010; Cyr et al.,
74 2010; Abbühl et al., 2011; Miller et al., 2013; Bellin et al., 2014; Kober et al.,
75 2015). Relatively few studies have determined CAER along the strike of an active
76 fault. For example Densmore et al. (2009) studied two faults in the western USA,
77 the 18 km long Sweetwater fault and the 130 km Wassuk fault. Along neither fault
78 were CAER found to be proportional to uplift rates nor to various topographic
79 measures. Densmore et al. (2009) attributed the uncoupling of erosion from fault
80 displacement to the influence of inherited high relief topography and the
81 widespread occurrence of mass wasting. In contrast, Rossi et al. (2017) reported
82 26 erosion rates along a normal fault system in Baja California demonstrating a
83 positive trend between CAER with slope and channel steepness. Roda-Boluda et
84 al. (2019) also showed a linear relationship between CAER and the footwall
85 component of fault throw rate from 15 samples taken from a series of catchments
86 crossing an active normal fault system in southern Italy. In all these studies the
87 footwalls of the studied faults are composed of metamorphic or igneous rocks with
88 limited reported lithological variability at a regional scale.

89 This lithological homogeneity of existing research areas is significant, as the
90 modelling of Forte et al. (2016) suggests that the presence of lithological contacts,
91 where rock strength changes from strong to weak, will profoundly influence the
92 response rates of an incising river system. For example, their modelling suggests
93 that when soft rocks overlie hard rocks (along a contact dipping at 20 - 35°
94 downstream) the lithological contact becomes an important and persistent
95 topographic feature in the landscape. Interestingly, although the geological
96 boundary moves downstream over time, the model suggests that erosion rates
97 above and below the boundary should diverge. The soft rocks downstream erode

98 at the imposed uplift rate, but the underlying hard rocks erode at a rate lower
99 than the regional uplift rate (Forte et al., 2016). The difference in the strength
100 and bedrock erodibility between the hard and soft rocks controls the magnitude
101 of difference between erosion and uplift rate, and also the duration of the
102 landscape adjustment. Subsequent modelling studies by Perne et al. (2017),
103 Darling et al. (2020), Wolpert and Forte (2021) and Mitchell and Yanites (2021)
104 are broadly consistent with Forte et al. (2016)'s results. Although the more
105 complex interbedded hard-soft rock scenarios of Darling et al. (2020)'s model
106 indicate that in such cases the harder rocks may erode quicker than the soft rocks.
107 A further implication of Forte et al.'s (2016) landscape evolution model is that
108 CAER, determined from cosmogenic radionuclides (CRN - commonly ^{10}Be), maybe
109 affected by the relative enrichment of material from the harder rocks in the detrital
110 sediment. Consequently, CAER would be perturbed or amplified because of the
111 lithological variation.

112 Therefore, there is a knowledge gap in our understanding of how erosion rates
113 change along faults with lithologically variable footwall geology. There is also the
114 requirement to empirically test the results of models such as Forte et al. (2016),
115 Perne et al. (2017) and Darling et al. (2020) in regions with complex geology to
116 assess the applicability of these models to real systems.

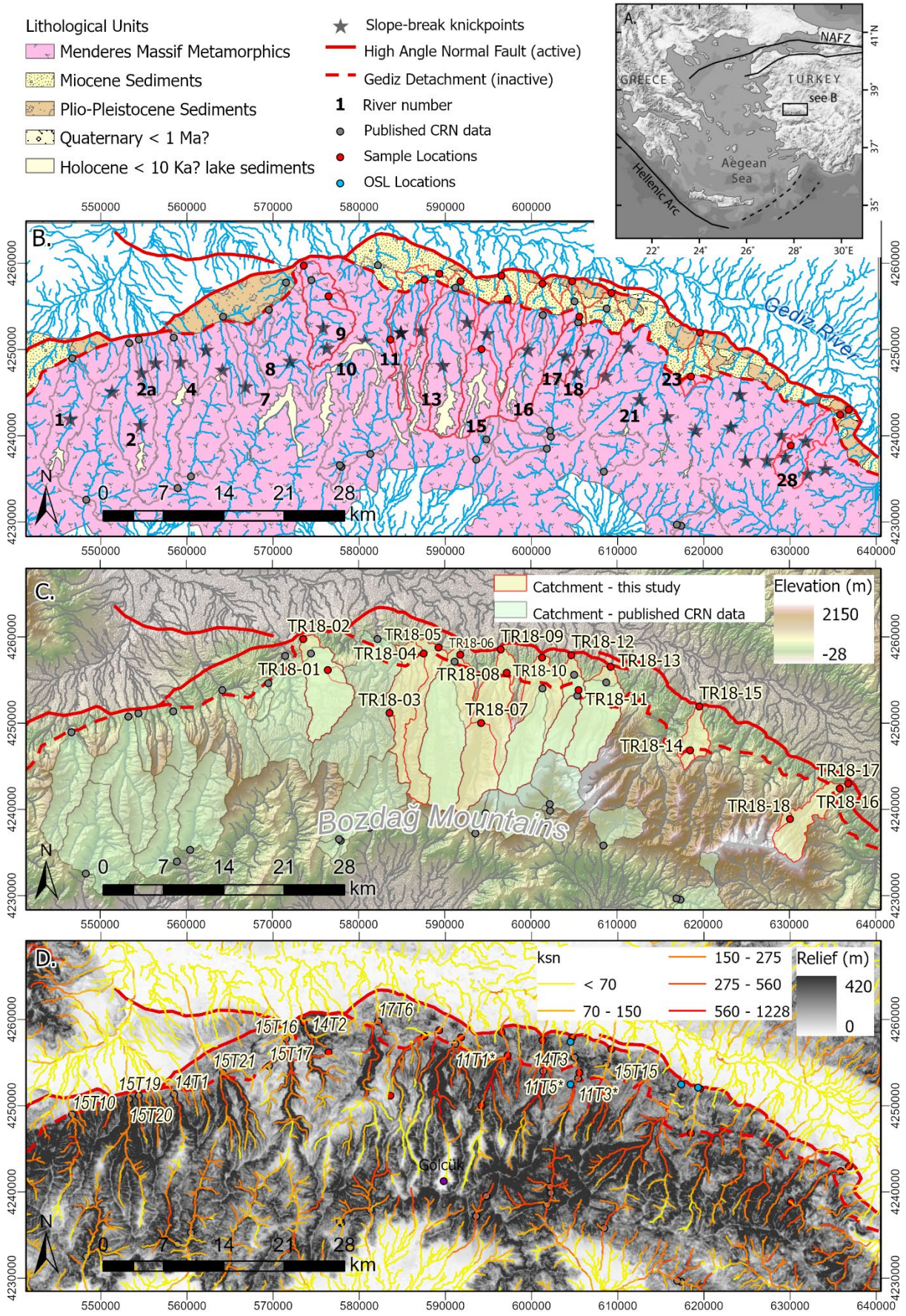
117 Here, we use the well-constrained Gediz fault system (western Türkiye) as a
118 natural laboratory to study the landscape response to fluvial incision across a
119 strong lithological contrast (soft rocks over hard rocks) in the footwall of an active
120 normal fault. As the geologic and geomorphic evolution of the region is well
121 understood and constrained (i.e., Seyitoğlu and Scott, 1996; Seyitoğlu et al.,
122 2002; Bozkurt, 2003; Bozkurt and Sözbilir, 2004; Çiftçi and Bozkurt, 2009a; Öner
123 and Dilek, 2011; Kent et al., 2016; 2017; 2021), we can use the area to test the
124 model predictions of Forte et al. (2016) and investigate the role that strength
125 contrasts play in the evolution of transient landscape responses to base-level fall.
126 This is achieved through a suite of new ^{10}Be and ^{26}Al CRN samples to determine
127 CAER along the strike of the boundary faults combined with published cosmogenic
128 data (Buscher et al., 2013; Heineke et al., 2019) and geomorphic indices (Kent et
129 al., 2021). Catchment-averaged erosion rates are quantified using ^{10}Be and ^{26}Al
130 so that the potential effect of sediment storage can be excluded, thus allowing

131 accurate exposure and denudation histories to be calculated (c.f., Bierman et al.,
132 1999; Granger and Muzikar, 2001; von Blanckenburg, 2005).

133

134 **2. Study area**

135 The Gediz (also known as the Alaşehir) Graben is located in western Anatolia
136 (Figure 1) forming an arcuate, asymmetric graben ~ 150 km in length. The Bozdağ
137 Range to the south is uplifted along the southern graben-bounding normal fault
138 and rises to over 2000 m in elevation. The ~ N-S extension forming this horst and
139 graben structure has been ongoing since early Miocene times, probably as the
140 result of roll-back along the Hellenic subduction zone (Okay and Satır, 2000; ten
141 Veen et al., 2009) and can be divided into two main phases (Bozkurt and Sözbilir,
142 2004). Initial extension caused uplift along the now-inactive low-angle north-
143 dipping Gediz detachment fault (Gessner et al., 2001; Seyitoğlu et al., 2002; Ring
144 et al., 2003). The Gediz detachment fault presently dips to the N-NE at up to 32°
145 and is gently corrugated along its strike (Sozbilir, 2012; Bozkurt and Sozbilir,
146 2012). The detachment forms the boundary between the Menderes Massif
147 metamorphic rocks and overlying syn-tectonic sedimentary rocks (Figure 2). In
148 the footwall, the Menderes Massif metamorphic core complex is composed mainly
149 of Palaeozoic greenschist to amphibolite-facies schists, augengneisses, and
150 paragneisses (Gessner et al., 2001; Ring et al., 2003).

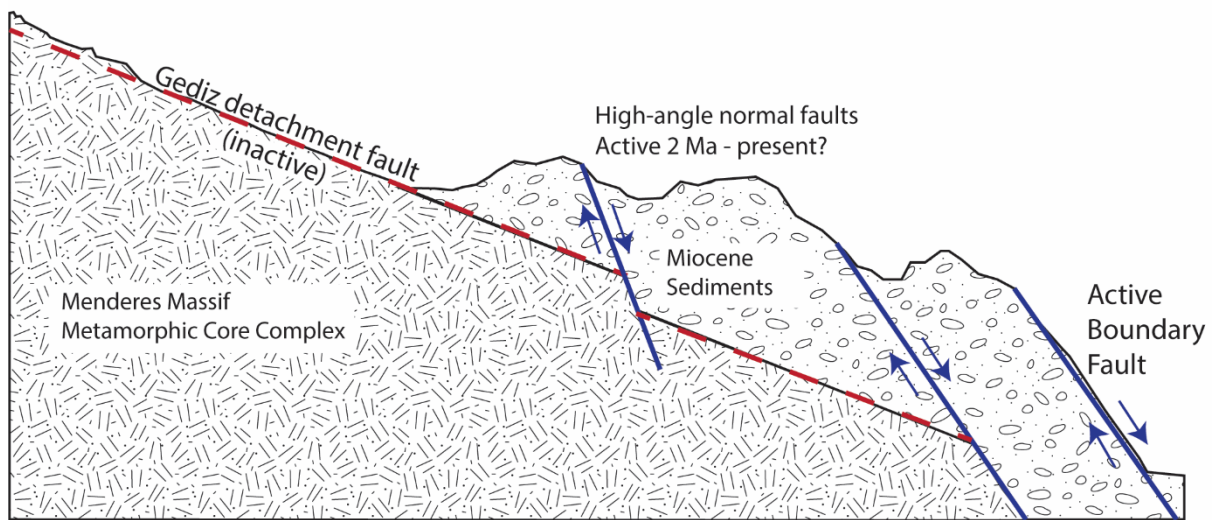


152 *Figure 1A). regional location map showing the location of the Gediz Graben in Western*
153 *Anatolia; B). geological map of the study area. Geological units are simplified from Kent*
154 *et al. (2021) with additional mapping of Holocene lake deposits from Süzen et al. (2006).*
155 *Numbers in bold indicate rivers sampled for CRN either in this study (table 1) or by*
156 *Heineke et al. (2019) or Buscher et al. (2013) (table 3), rivers mentioned by name in the*
157 *text are 9 - Akçapınar; 15 - Bozdağ; 16 - Gümüşçay; 17 - Kabazlı; 21 - Kavaklıdere; 23*
158 *- Yeniköy. Stars show location of slope-break knickpoints; C). topographic map of the*
159 *study area (ALOS World 3D 30 m DEM) showing the sample locations with sample numbers*
160 *collected during this study; D). relief map of the study area showing the steepness index*
161 *of the rivers and the location of CRN samples collected by Buscher et al. (2013) indicated*
162 *by * and Heineke et al. (2019) Also shown are the location of five OSL dates reported by*
163 *Kent (2015) (unlabelled - blue) and the approximate location of the C14 date of Sullivan*
164 *(1988) labelled as Gölcük. Following the cessation of slip on the Gediz detachment fault*
165 *at ca. 2 Ma (Buscher et al., 2013), strain stepped northwards (basinwards) onto high angle*
166 *faults. These include the presently active normal fault forming the range front fault*
167 *(Figures 1 and 2) to the present-day topographic graben (Çiftçi and Bozkurt, 2009a). In*
168 *the uplifted footwall of the active fault are friable sedimentary rocks deposited originally*
169 *on the hangingwall of the Gediz detachment. These sedimentary units, comprised mainly*
170 *of early Miocene to Pliocene-aged alluvial fan and fluvial sandstones and conglomerates,*
171 *unconformably overlie and derive from the metamorphic basement (e.g., Purvis and*
172 *Robertson, 2004; 2005; Çiftçi and Bozkurt, 2009b).*

173

174 Quaternary sediments are variable in extent across the Bozdağ range (Figure 1).
175 Fragments of river terraces have been reported by Kent (2015) along three rivers
176 - the Yeniköy, Kavaklıdere and the Kabazlı (Figure 1B). These river terraces are
177 of small spatial extent with OSL dates of five samples (Figure 1D) from the fine-
178 grained facies of only one, well-developed, terrace level indicating aggradation
179 between ~ 84 - 7.5 ka (Kent, 2015). However, in the headwaters of several of
180 the larger river systems fluvial and lacustrine fine-grained sediments up to 170 m
181 thick can be found (Süzen et al., 2006). Sediment cores from Gölcük Lake (Figure
182 1D) yielded ¹⁴C dates of ≤ 10 ka (Sullivan, 1988) suggesting deposition during the
183 Holocene to Pleistocene but ages of the older sediments are not constrained. These
184 deposits are thought to have formed owing to 1 - 2° of rotation on the graben
185 boundary fault during the Holocene resulting in slope reduction, lake formation,
186 and sediment deposition (Süzen et al., 2006).

187 Across the Bozdağ Range, transverse bedrock rivers flow northwards into the
188 Gediz Graben across the southern boundary fault. The rivers are generally deeply
189 incised with prominent knickpoints and gorges upstream of the active fault. The
190 slope-break knickpoints are not coincident with lithological boundaries (Kent et
191 al., 2017) and are interpreted to mark the upstream extent of transient wave of
192 river incision. Incision was caused by an increase in slip on the graben bounding
193 fault as a result of the fault linkage of three initial fault segments $\sim 0.6 - 1$ Ma
194 (Kent et al., 2016; 2017). As a result of this linkage, present day throw rates (the
195 vertical component of the slip rate) are now thought to be higher than the long-
196 term average, with rates of up to 2 ± 0.2 mmyr⁻¹ calculated for the centre of the
197 fault array (Kent et al., 2017).



198
199 *Figure 2. Simplified cross-section of the northern margin of the Bozdağ Horst showing the*
200 *relationship between low and high-angle faults (adapted from Kent et al., 2016).*

201 Kent et al. (2021) selected six of the transverse rivers to investigate the
202 lithological controls on transient river behaviour. For simplicity, Kent et al. (2021)
203 used two broad groupings of rock types; metamorphic and sedimentary in their
204 quantitative analyses. Rivers were chosen to investigate differences in the
205 proportion of metamorphic to sedimentary bedrock reaches (100% metamorphic
206 in the Akçapınar River through to $\sim 50\%$ along the Yeniköy River; Figure 1B) and
207 differences in uplift rate as a result of activity along the graben boundary fault.
208 Here we continue to use these two broad lithologic groups to allow comparisons
209 to this previous work.

211 **3. Methods**

212 **3.1. Sample collection and CRN**

213 Eighteen samples of river sand from the active riverbed or sediment bars were
214 collected from nine catchments draining northwards across the Gediz Graben
215 boundary fault in May 2018 (Figures 1c and 3). The rivers were selected because
216 either they had previously been sampled by Buscher et al. (2013) or were one of
217 the six rivers studied in detail by Kent et al. (2021). Overall, a nested sampling
218 strategy was adopted so that ten samples were collected from the range front
219 where the rivers cross the active normal fault. On the easternmost river, two
220 samples were collected ~ 2 km apart to assess downstream mixing and
221 reproducibility. The remaining eight samples collected further upstream at either
222 the lithological boundary between the sedimentary and metamorphic rocks or
223 upstream of the knickpoint. Five of these eight samples were collected upstream
224 of identified slope-break (tectonic) knickpoints identified by Kent et al. (2017) and
225 the final three samples were collected at the low-angle detachment that forms the
226 lithological boundary enabling comparison to published datasets. A further CRN
227 dataset was published by Heineke et al. (2019) bringing the total number of
228 samples analysed in the Gediz region to 33.

229 The eighteen samples collected here were sieved to 2 mm in field and further
230 sieved to the 250-500 μm size fraction in the lab. Standard magnetic separation
231 to concentrate the quartz fraction of the sample using a Franz magnetic separator
232 was undertaken at the University of Plymouth. Subsequently samples were
233 chemically leached using diluted HF, and between 16 and 20 grams of clean quartz
234 cores were dissolved at SUERC together with ~ 0.29 grams of the CIAF-PH9 in-
235 house ^9Be carrier solution ($[\text{Be}] = 849 \pm 12$ ppm) following the procedure of Child
236 et al. (2000). ^{10}Be and ^{26}Al concentrations were measured by the 5-MV NEC
237 Pelletron accelerator mass spectrometer (AMS) at SUERC (Xu et al., 2010).

238 The results were input into the online CRONUS-Earth calculator v 3.0 (Balco et al.,
239 2008) using the LSDn scaling, a sample density of 2.65 g cm^3 and NIST_27900
240 and Z92-0222 standardisations for ^{10}Be and ^{26}Al , respectively. Mean catchment
241 elevation and shielding were derived from the ALOS World3D 30 m DEM, which
242 has been shown to extract more accurate hydrological networks than other
243 comparable global DEMs (Boulton and Stokes, 2018) using ArcGIS Pro 2.6.2 and
244 TopoToolBox functions (Schwanghart and Scherler, 2014). Similarly, catchment

245 mean slope and relief over a 150 m radius were extracted using standard GIS
246 tools.



247
248 *Figure 3 – Field photos showing landscapes and sampling in the Gediz region: A) View of*
249 *the downstream reach of the Akçapınar River – a river characterised by 100%*
250 *metamorphic bedrock, B) Sampling in the knickzone of the Bozdağ River, C) Sampling in*
251 *the sedimentary reach of the Gümüşçay, note the well lithified Miocene clastic bedrock, D)*
252 *Vertical step knickzone on the Kabazlı River at the boundary between the metamorphic*
253 *basement and the sedimentary cover.*

254 Burial ages were derived from ^{10}Be and ^{26}Al data following the same principles as
255 Granger and Muzikar (2001). This method allows solving of both the erosion rate
256 corresponding to the initial ^{10}Be and ^{26}Al concentrations, and the average burial
257 time after the exhumation of the quartz grains. To make them consistent with
258 CRONUS v.3 results, scaled concentrations, spallation and muon production rates,
259 and attenuation lengths were calculated as in Rodés (2021).

260 We also recalculated the ^{10}Be sample concentrations reported in Buscher et al.
261 (2013) and Heineke et al. (2019) for our study area using the same parameters
262 stated above (e.g., using topographic shielding and a sample density of 2.65 gcm^3
263 and CRONUS v 3). Note that Heineke et al. (2019) did not apply a topographic
264 shielding and used a sample density of $2.2 - 2.5\text{ gcm}^3$ in addition to using v 2.3

265 of the CRONUS-Earth calculator, which results in differences in the erosion rates
266 stated here compared to those reported in the original papers. Neither of these
267 previous studies included ²⁶Al concentrations, so corrections for sediment
268 reworking or burial cannot be determined for these previously published CRN data.

269

270 **3.2. Sediment (un)mixing**

271 In the Bozdağ catchments studied samples were taken at the catchment outlet,
272 at the major lithological boundary and in five locations above the slope-break
273 knickpoint. This sampling strategy allows the erosion rates above (un-incised) and
274 below (incised) the slope-break knickpoint to be deconvolved assuming that the
275 same amount of quartz-bearing sediment is produced in both parts of the
276 watershed. The sediment mixing is determined using the approach of Granger et
277 al. (1996), as the CRN records the average erosion rate for the entire contributing
278 catchment area. Therefore, the erosion rate between two sample points (a
279 'subcatchment') can be determined by correcting for the upstream sediment flux
280 according to:

$$281 \quad E_b = \frac{(E_c \times A_c) - (E_a \times A_a)}{A_b} \quad (\text{e.q. 2})$$

282 Where E (mMyr⁻¹) is the erosion rate of a catchment with area A (m²), with
283 subscripts indicating different subcatchments (Figure 4), where *c* is the entire
284 catchment and *a* and *b* are the upstream and downstream subcatchments,
285 respectively. In this study a *single* common value for the upstream erosion rate
286 E_A is used for all catchments owing to; a) the limited data on the CAER above the
287 knickpoint, b) the assumption that this area represents a low relief and low erosion
288 rate landscape formed prior to the uplift causing the present transient river
289 incision.

290 ArcGIS Pro 2.6.2 was used to calculate the areas used in the unmixing
291 calculations. The knickpoint finder tool in TopoToolBox (Schwanghart and
292 Scherler, 2014; Stolle et al., 2019) was used to identify the highest knickpoint
293 along all tributaries in the study area using a tolerance of 30. These were then
294 used as pour points for the watershed tool, the results of which were then summed
295 to determine the total unincised area in each river catchment, which is then

296 subtracted from the total catchment area calculated in the same way for the
 297 sample locations.

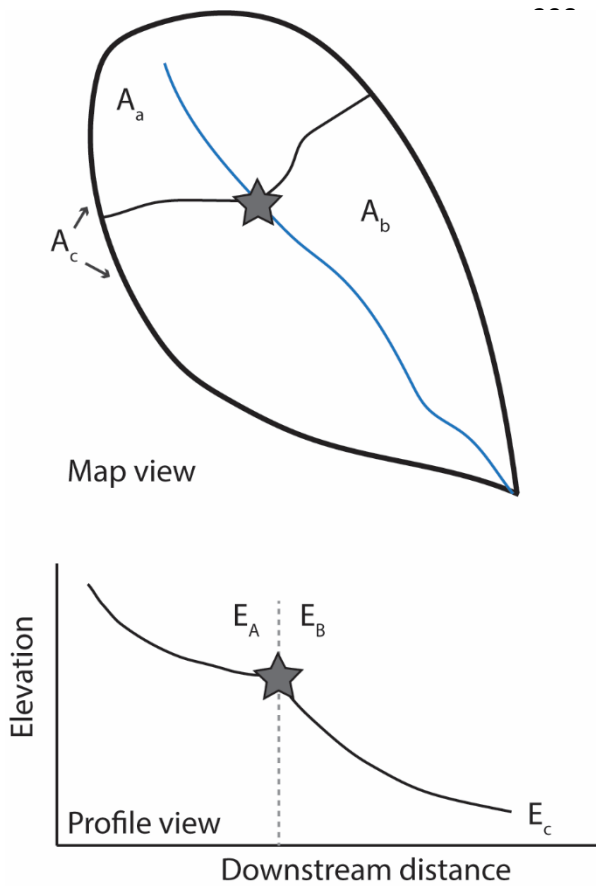


Figure 4 – Conceptual diagram showing how different erosional zones add together to define total erosion rate at sample location. Top, a map view of a two zone mixing model showing the catchment areas above, A_a, and below, A_b, the knickpoint comprising the total catchment area A_c. Below, a topographic profile showing how the different zones relate to the transient river long profile with the samples collected at the knickpoint (star), E_A, and at the river mouth, E_C, allowing the determination of the erosion rate of only the transient, incising reach E_B (modified from Rosenkranz et al. (2018)).

315

316 **3.3. Calculation of unit stream power**

317 Geomorphic indices were calculated using ArcGIS Pro and TAK (Forte and Whipple,
 318 2018), k_{sn} values were determined with a $\Theta_{ref} = 0.45$ following Kent et al
 319 (2017;2021) and the profiler function. While the choice of reference concavity
 320 can impact the resultant K_{sn} values, Gailleton et al., (2021) demonstrated this is
 321 not significant.

322 Kent et al. (2021) constrained the rock strength (using Schmidt hammer rebound
 323 and Selby Rock Mass strength) and specific bedrock erodibility, E , using the unit
 324 stream power model (c.f., Whittaker et al., 2007; Attal et al., 2011; Zondervan et
 325 al., 2020):

$$326 \quad E = k_b \omega = k_b \frac{\rho g Q S}{W} \quad (\text{e.q. 1})$$

327 where the unit stream power, ω represents energy dissipation per unit channel
328 area on the bed with units of Wm^{-2} , ρ is the density of water, g is the acceleration
329 due to gravity, Q is the water discharge (m^3s^{-1}), S is local channel slope (m/m)
330 and W the channel width (m) as measured in the field. Consequently, specific
331 bedrock erodibility, k_b , has units of $\text{ms}^2\text{kg}^{-1}$, representing the inverse of stress (c.f.
332 Yanites et al., 2017).

333 Kent et al. (2021) demonstrate that the metamorphic rocks are around twice as
334 hard as the sedimentary rocks. This difference is reflected by the bedrock
335 erodibility, which was calculated as $2.2 - 6.3 \times 10^{-14} \text{ ms}^2\text{kg}^{-1}$ in the metamorphic
336 rocks. In contrast, bedrock erodibility values in the sedimentary units were 5 to
337 30 times larger (i.e., 5 to 30 times weaker) at 1.2×10^{-13} to $1.5 \times 10^{-12} \text{ ms}^2\text{kg}^{-1}$
338 (Kent et al., 2021). Significantly, stream power was shown to scale with fault
339 throw rate in the metamorphic rocks but not in the sedimentary units; potentially
340 because the weaker sedimentary rocks themselves directly influence the fluvial
341 processes and long-term erosional dynamics.

342 However, values for unit stream power (equation 3) for each river with reported
343 CRN concentrations are required. Using the regional Q to A relationship
344 determined using field measurements for the six rivers detailed in Kent et al.
345 (2021), the estimate of Q for each river is found by extracting cumulative
346 catchment area downstream along each sampled river at 100 m intervals using
347 ArcGIS Pro 2.6.2 and the ALOS World 3D30 DEM. Similarly, the elevation is
348 extracted at each point allowing the determination of local channel slope over each
349 100 m interval. The vertical accuracy of the AW3D30 DEM is $< 5 \text{ m}$ (Tadono et
350 al., 2016). As field-derived measurements of width are not available for all rivers,
351 width is calculated using the scaling relationships of Finnegan et al. (2005) and
352 Whittaker et al. (2007) as well as using Kent et al.'s (2021) local hydraulic scaling
353 relationship (see supplemental methods for more detail). These estimates of width
354 are then used to derive the downstream distribution of unit stream power, ω , for
355 each river. The maximum stream power was found for each river, and an average
356 of the three stream powers taken. The error reported is the 2σ value on these
357 values.

358

359 **3.4. Rock strength and erodibility measurements**

360 *In situ* rock strength measurements can be used to estimate bedrock erodibility,
361 which is related to the inverse of the lithologies tensile strength (Sklar and
362 Dietrich, 2001). However, tensile strength measurements are difficult to measure
363 in the field and as a result the Schmidt hammer is commonly utilised owing to the
364 ease of use and portability (e.g., Goudie, 2016). Kent et al. (2021) used an N-
365 type Schmidt hammer to characterise average bedrock uniaxial compressive
366 strength for each lithological unit. Additionally, information on fracture
367 characteristics was collected to calculate the semi-quantitative Selby Rock Mass
368 strength – SRMS (Selby, 1980).

369 Twenty Schmidt hammer readings were taken at 130 locations along the six study
370 rivers, the majority of which are from the metamorphic basement. At only eight
371 locations could the Schmidt hammer reliably return a rebound value for the
372 sedimentary rocks. At another 28 sites the exposed bedrock was too weak to
373 accurately characterise the strength using this method and was recorded as
374 having a rebound strength of < 20 (the effective limit of the Schmidt hammer),
375 allowing the SRMS to be determined even where bedrock is very weak. Schmidt
376 hammer rebound and SRMS are then averaged for the ~ 2 km upstream of the
377 CRN sample locations where possible.

378

379 **4. Results**

380 **4.1. ^{10}Be and ^{26}Al concentrations and catchment-wide erosion** 381 **rate**

382

383 The ^{10}Be concentrations measured in the new samples range from 1.3 – 10.0 x
384 10^4 atoms g^{-1} , while there were between 1.6 – 96.4 x 10^4 atoms g^{-1} of ^{26}Al (Table
385 1). These values compare well to previously reported CRN concentrations of ^{10}Be
386 in the range 1.5 – 13.7 x 10^4 atoms g^{-1} (Supplemental Table 1) from sediment in
387 rivers mainly draining the metamorphic basement (Buscher et al., 2013; Heineke
388 et al., 2019).

389

390 Table 1. New ^{10}Be and ^{26}Al analytical and derived erosion rate data with no corrections for
 391 subcatchments or sediment recycling/burial.

Sample	Location			River No (Kent)	Distance along strike (km)	Mean Catchment elevation (m)‡	Topographic shielding	Measured concentrations*						Denudation rates (no corrections)†			
	Latitude (°N)	Longitude (°E)	River Name					^{10}Be Production rate (at/g/yr)	^{26}Al Production rate (at/g/yr)	^{10}Be concentration (Atoms g ⁻¹)	Uncertainty in ^{10}Be concentration (Atoms g ⁻¹)	^{26}Al concentration (Atoms g ⁻¹)	Uncertainty in ^{26}Al concentration (Atoms g ⁻¹)	^{10}Be (m Myr ⁻¹)	Internal uncertainty	^{26}Al (m Myr ⁻¹)	Internal uncertainty
TR1801	38.44701	27.877054	Akçapınar	9	35.4	787	0.9898	7.6806	53.9639	19935	1483	107543	14843	236.0	17.6	316.0	43.6
TR1802	38.48295	27.843433	Akçapınar	9	35.4	704	0.9750	7.0490	49.5853	29153	1527	207696	21774	150.0	7.9	150.0	15.8
TR1803	38.4048	27.957544	Sart Çay	11	53.4	1002	0.9738	9.0573	63.4609	152808	4001	964210	51577	36.1	1.0	40.1	2.2
TR1804	38.46693	28.003768	Sart Çay	11	53.4	844	0.9746	7.9411	55.7586	16803	1251	68007	9544	290.0	21.6	517.0	72.7
TR1805	38.47295	28.024298	Sart Çay	11	53.4	854	0.9743	8.0088	56.2268	13521	1218	16163	6201	363.0	32.8	3060.0	1180.0
TR1806	38.4651	28.052133	Çaltılı	13	56.3	1018	0.9608	9.0620	63.4898	97728	3043	405368	27218	56.3	1.8	94.5	6.4
TR1807	38.39361	28.079372	Bozdağ	15	60.3	1301	0.9663	11.4862	80.1738	92364	3127	531175	34773	72.9	2.5	88.3	5.8
TR1808	38.4457	28.114133	Bozdağ	15	60.3	1215	0.9639	10.6928	74.7231	85558	2677	508773	33300	73.9	2.3	86.8	5.7
TR1809	38.47007	28.106199	Bozdağ	15	60.3	1108	0.9640	9.8025	68.5974	74938	2245	462547	30666	78.1	2.4	88.6	5.9
TR1810	38.46108	28.160952	Gümüş Çayı	16	65.4	1149	0.9594	10.0898	70.5714	26596	1521	124819	14650	222.0	12.7	339.0	39.9
TR1811	38.42648	28.208946	Kabazlı	17	69	1347	0.9770	12.0506	84.0655	100108	3106	586261	36870	70.2	2.2	83.6	5.3
TR1812	38.46332	28.200208	Kabazlı	17	69	997	0.9736	9.0205	63.2126	17481	1299	65090	9569	310.0	23.0	598.0	87.9
TR1813	38.45042	28.252572	Yeşilkavak	18	73.7	983	0.9616	8.8043	61.7107	45207	1988	287656	20621	117.0	5.2	130.0	9.4
TR1814	38.36203	28.356265	Yeniköy	23	85	795	0.9775	7.6208	53.5361	17675	1518	59004	8148	265.0	22.8	576.0	79.6
TR1815	38.40754	28.369389	Yeniköy	23	85	526	0.9769	6.0361	42.5557	19759	1384	42834	7648	196.0	13.7	661.0	118.0
TR1816	38.31973	28.553925	Badınca	28	105.1	1051	0.9574	9.2513	64.7776	38592	1806	300619	26206	142.0	6.7	129.0	11.3
TR1817	38.32456	28.565227	Badınca	28	105.1	1034	0.9579	9.1283	63.9307	45977	2250	240144	19537	118.0	5.8	160.0	13.0
TR1818	38.28846	28.487471	Badınca	28	105.1	1165	0.9642	10.2264	71.4968	22853	1559	153479	15142	262.0	17.9	277.0	27.4

* ^{10}Be and ^{26}Al concentrations were measured by the 5-MV NEC Pelletron accelerator mass spectrometer (AMS) at SUERC (Xu et al., 2010). Measured ^{10}Be is normalised to the NIST_27900 st
 † Denudation rates were calculated using the online CRONUS-Earth calculator v 3.0 (Balco et al., 2008) using the LSDn scaling and a sample density of 2.65 gcm⁻³.
 ‡ Sample elevation and shielding were derived from the ALOS World3D 30 m DEM

392

393

394 Therefore, apparent denudation rates range between 36 to 363 mMyr⁻¹ and 40 to
 395 3060 mMyr⁻¹ for ^{10}Be and ^{26}Al , respectively. However, the denudation rates
 396 estimated from both nuclides agree within error for < 30% of the samples. These
 397 samples show $^{26}\text{Al}/^{10}\text{Be}$ ratios in the range 6.2 – 7.8. The samples with a larger
 398 deviation between the derived denudation rates of each nuclide have significantly
 399 depleted $^{26}\text{Al}/^{10}\text{Be}$ ratios of < 5.2 (Table 2). In a two-isotope diagram (Figure 5),
 400 44% of data points cluster in the 0 – 0.5 Ma burial zone; 17% in the 0.5 – 1 Ma
 401 burial zone and 39% of points in the > 1 Ma burial zone. These data indicate that
 402 a simple exposure/denudation history, without taking into account sediment
 403 storage, is incorrect for the majority of samples and implies that sediment
 404 reworking from the alluvial plain and/or the uplifted sediments is contributing a
 405 significant component of the transported bedload in many rivers (c.f. Granger et
 406 al., 1996).

407

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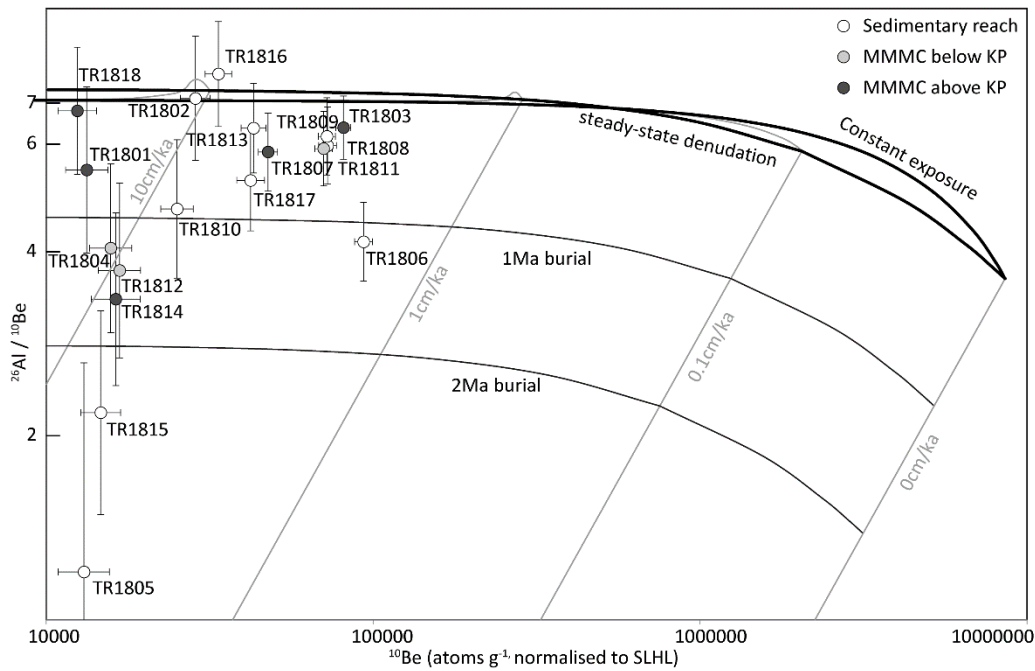
409

410 *Table 2. ²⁶Al/¹⁰Be ratios, burial age and recalculated total catchment erosion rates based*
 411 *upon burial corrections.*

Sample name	²⁶ Al/ ¹⁰ Be ratio	Burial age (Ma)	Burial erosion rate (m/Myr)	corrected erosion rate (m/Myr)
TR1801	5.4±0.9	0.60 ± 0.41	174 ± 45	
TR1802	7.1±0.8	0.00 ± 0.34	149 ± 19	
TR1803	6.3±0.4	0.21 ± 0.25	32 ± 5	
TR1804	4.0±0.6	1.19 ± 0.41	159 ± 42	
TR1805	1.2±0.5	4.40 ± 1.03	40 ± 19	
TR1806	4.1±0.3	1.05 ± 0.27	33 ± 6	
TR1807	5.7±0.4	0.39 ± 0.27	60 ± 11	
TR1808	5.9±0.4	0.33 ± 0.27	63 ± 11	
TR1809	6.2±0.5	0.26 ± 0.27	69 ± 12	
TR1810	4.7±0.6	0.87 ± 0.36	144 ± 33	
TR1811	5.9±0.4	0.36 ± 0.26	59 ± 10	
TR1812	3.7±0.6	1.35 ± 0.43	157 ± 42	
TR1813	6.4±0.5	0.22 ± 0.28	105 ± 20	
TR1814	3.3±0.5	1.60 ± 0.42	119 ± 33	
TR1815	2.2±0.4	2.49 ± 0.48	56 ± 16	
TR1816	7.8±0.8	0.00 ± 0.18	139 ± 10	
TR1817	5.2±0.5	0.62 ± 0.30	86 ± 17	
TR1818	6.7±0.8	0.12 ± 0.34	248 ± 45	

412

413 Therefore, the ²⁶Al data allows the calculation of an average burial history and the
 414 determination of a new erosion rate taking into account the depletion of the ¹⁰Be
 415 and ²⁶Al concentration during the time that the quartz grains were buried (Table
 416 2). This calculation gives 'burial-corrected' erosion rates of 32 to 248 mMyr⁻¹ for
 417 the study area catchments. Unfortunately, a similar calculation cannot be
 418 undertaken on the existing published CRN datasets (Buscher et al., 2013; Heineke
 419 et al., 2019) as there are no reported ²⁶Al data. As these sites are predominantly
 420 located in the footwall of the detachment fault, where there is little or no outcrop
 421 of sediments, it suggests that sediment storage should be limited for these
 422 samples. However, the presence of Holocene or older sediments in some
 423 catchments is a source of potential error that cannot be accounted for in the
 424 previously published data, and may explain why the published erosion rates are
 425 in general slightly higher than those reported here. This hypothesis is supported
 426 by the ²⁶Al/¹⁰Be ratios of three of samples upstream of the boundary between the
 427 sedimentary rocks and the Menderes Massif metamorphics falling in the > 1 Ma
 428 burial zone (Figure 5).



429

430 *Figure 5. $^{26}\text{Al}/^{10}\text{Be}$ vs ^{10}Be ratio two isotope diagram showing burial model and*
 431 *concentration data scaled to surface production rates (Lal, 1991) for measured samples.*
 432 *Surface muon contributions of $0.99 \pm 0.20\%$ and $1.45 \pm 0.29\%$ were considered for ^{10}Be*
 433 *and ^{26}Al respectively. Samples taken above the slope-break knickpoint are indicated by the*
 434 *grey symbols. MMMC = Menderes Massif Metamorphic core complex. Error bars include*
 435 *analytical and production rate uncertainties.*

436

437 On five rivers, samples were taken at or upstream of the slope-break knickpoint
 438 (TR18-01; 03; 07; 14; 18). These samples represent the denudation rate prior to
 439 landscape rejuvenation and transient river incision, as a result of fault linkage \sim
 440 0.8 Ma (Kent et al., 2017), providing constraints for the unmixing model to
 441 determine the rate of erosion excluding these low erosion rate areas. Samples
 442 TR18-03 and TR18-07 give the lowest burial-corrected erosion rates at 32 and 60
 443 mMyr^{-1} , respectively. Ridge crest erosion rates determined by Heineke et al.
 444 (2019) also fall in the range $\sim 30 - 90 \text{ mMyr}^{-1}$. Whereas, samples TR18-01 and
 445 TR18-18 give much higher rates of 174 and 248 mMyr^{-1} , respectively; while TR18-
 446 14 returns an intermediate value of 119 mMyr^{-1} . Significantly, these latter three
 447 samples have only small catchment areas upstream of the sample point ($1.7 - 3$
 448 km^2), which may be below a threshold for an appropriate size of catchment area.
 449 Additionally, the CAER from ^{10}Be and ^{26}Al nuclides are not within error and

450 consequently indicate variable sediment recycling, which is difficult explain in the
451 metamorphic headwaters. Therefore, given the higher values than for the ridge
452 crests, small catchment areas and incomplete mixing, these latter three samples
453 are not used to determine the erosion rate upstream of the knickpoint. Instead,
454 the average of the other two samples is taken to be representative of the low
455 incision zone and used for all catchments (c.f., Roda-Boluda et al., 2018).
456 Therefore, the average CAER used is 46 mMyr^{-1} above the knickpoints.

457 **4.2. Results from unmixing model**

458 In a landscape experiencing transient river incision, erosion rates above the
459 knickpoint are expected to be lower than below the knickpoint. Therefore, we used
460 a unmixing method (e.g., Granger et al., 1996; Rosenkranz et al., 2018) to
461 remove the influence of such low erosion rates on downstream samples. Using the
462 minimum erosion rate estimate determined above (i.e. 46 mMyr^{-1}), it is possible
463 to derive a quantitative estimate for the erosion rates within the transient reach;
464 i.e., upstream of the active fault and downstream of the knickpoint. This method
465 is applied to both the new burial-corrected CAER and also the previously published
466 CRN datasets (Table 3). The effect of applying this unmixing model is variable
467 depending on the proportion of the total catchment area falling in the low erosion
468 rate zone above the knickpoint, and on the difference between the low erosion
469 rate and the denudation rate determined for the downstream sample (Table 3).
470 For example, where the downstream initial burial-corrected CAER are relatively
471 low (such as on the Bozdağ) the unmixing results in a small increase in CAER
472 (e.g., from 63 mMyr^{-1} to 99 mMyr^{-1}). But where the difference between the
473 assumed upstream erosion rate of 46 mMyr^{-1} and the downstream sample is
474 greater, the final calculated rate is markedly higher. For example, on the
475 Gumuşçay the initial burial-corrected CAER is 144 mMyr^{-1} , which increases to 1330
476 mMyr^{-1} after unmixing; a tenfold increase. For the majority of samples the rates
477 do increase, but a limited number of samples from or close to the lithological
478 boundary result in no or negligible change. This is because the measured rate is
479 close to the low erosion rate value even though the samples are within the
480 knickzone. For one sample - 14T1 (Heineke et al., 2019), this adjustment results
481 in a negative erosion rate. This CAER is not included in further analyses.

482

483 *Table 3. Parameters used in the unmixing calculations to remove effect of low erosion rate*
 484 *and resultant erosion rates (Eb) for transient reach.*

Sample No	River No	Catchment area (m ²)			Erosion rates (LSDn)(m/Myr)					
		Aa	Ab	Ac	Ea	±	Ec	±	Eb	±
15T10	1	23827405	19380606	43208011	46	14	55	7	65	23
15T20	2	33018658	30633406	63652064	46	14	46	6	45	20
15T19	2a	3713634	3729325	7442959	46	14	82	11	118	26
14T1	4	12346068	4072945	16419013	46	14	33	4	-6	45
15T21	7	5464432	16719167	22183599	46	14	79	10	90	14
15T16	8	37403840	26672597	64076437	46	14	92	14	157	39
15T17	8	3186523	815578.2	4002101	46	14	73	10	180	73
TR18-02	9	25649691	20803431	46453122	46	14	149	19	276	46
14T2	9	25649691	20803431	46453122	46	14	187	37	361	84
17T6	10	68662720	27467993	96130713	46	14	250	39	760	141
TR18-05	11	43799839	26295684	70095523	46	14	40	19	30	56
TR18-04	11	43799839	26295684	70095523	46	14	159	42	347	114
11T1*	13	45492741	35010065	80502806	46	14	60	7	77	24
TR18-06	13	45492741	35010065	80502806	46	14	33	6	16	23
TR18-09	15	47765237	22648803	70414040	46	14	69	12	118	48
TR18-08	15	47765237	22648803	70414040	46	14	63	11	99	45
TR18-10	16	30641229	5400620	59651748	46	14	144	33	1330	373
11T5*	16	30641229	5400620	36041849	46	14	61	6	148	89
TR18-12	17	11032158	16141920	27174078	46	14	157	42	233	71
14T3	17	11032158	11455766	22487924	46	14	151	22	252	45
TR18-11	17	11032158	5741419	16773577	46	14	157	10	370	40
11T3*	17	11032158	5558898	16591056	46	14	63	8	95	37
TR18-13	18	13303890	32870296	46174186	46	14	105	20	129	29
15T15	18	13303890	32870296	46174186	46	14	239	43	317	61
11T4*	18	13303890	13624552	26928442	46	14	126	17	204	36
TR18-15	23	1932233	12999221	14931454	46	14	56	16	57	18
TR18-17	28	15970937	12825476	28796413	46	14	86	17	136	42
TR18-16	28	15970937	12219002	28189939	46	14	139	10	261	29

485

486

487 **4.3. Relationship between CAER and geomorphic indices**

488 These calculations enable the comparison between erosion rates to a number of
 489 geomorphic and geologic measures (Table 4). The burial-corrected mixed rates
 490 (i.e., CAER for the entire catchment) and the burial-corrected unmixed rates for
 491 the transient reaches (with the area upstream of the knickpoint removed) are
 492 compared alongside the recalculated published CAER (Buscher et al., 2013;

493 Heineke et al., 2019) and the published CAER unmixed for the low erosion rate
 494 area, to investigate the relationships between different factors and erosion along
 495 the southern margin of the Gediz Graben.

River Name (Kent)	River No	Sample	¹⁰ Be source	Distance along strike (km)	Catchment area (km ²)	Mean catchment slope (°)	Mean slope above KP (°)	Mean slope below KP (°)	Mean catchment relief (m)	Mean catchment relief above KP (m)	Mean catchment relief below KP (m)	k _{in} upstream of sample (m ^{0.5})	Maximum stream power (Wm ⁻²)	Max incision upstream of sample (m)	Throw rate @fault since 0.7 Ma (mmyr ⁻¹)	Long term throw rate since 2 Ma (mmyr ⁻¹)		
Akçapınar	9	TR1801	This study	35.4	3	12.6	15.16	20.85	60	66	101	184.4	2.7	1987	2314	132	1.41	0.44
Akçapınar	9	TR1802	This study	35.4	46	16.7	15.16	20.85	78	66	101	69.7	1.4	1987	406	356	1.41	0.44
Sart Cay	11	TR1803	This study	53.4	1	8.8	15.17	18.47	77	66	85	161.3	2.5	1792	2080	29	1.84	0.99
Sart Cay	11	TR1804	This study	53.4	5	17.5	15.17	18.47	77	66	85	94.4	2.0	1792	528	209	1.84	0.99
Sart Cay	11	TR1805	This study	53.4	38	17.5	15.6	18.47	79	66	85	94.4	2.0	1792	528	308	1.84	0.99
Çaltılı	13	TR1806	This study	56.3	80	20.5	18.33	24.49	102	90	124	107.8	2.6	2600	704	615	1.91	1.28
Bozdağ	15	TR1807	This study	60.3	34	20.3	18.33	22.02	96	90	107	70.0	0.8	6163	427	564	2	1.42
Bozdağ	15	TR1808	This study	60.3	64	21.5	18.33	22.02	101	90	107	91.7	0.7	6163	818	564	2	1.42
Bozdağ	15	TR1809	This study	60.3	70	19.3	18.33	22.02	99	90	107	100.3	3.3	6163	1198	564	2	1.42
Gümüş Çayı	16	TR1810	This study	65.4	60	21.7	22.41	20.75	106	112	100	123.7	3.2	3986	1060	634	1.86	1.33
Kabazlı	17	TR1811	This study	69	17	24.7	18.97	17.33	84	92	81	133.7	2.7	2376	1454	279	1.74	1.33
Kabazlı	17	TR1812	This study	69	28	17.6	18.97	17.33	84	92	81	104.3	2.2	2376	1439	331	1.74	1.33
Yeşilkavak	18	TR1813	This study	73.7	46	21.1	23.75	19.81	101	117	95	133.7	2.7	3274	1454	355	1.58	1.48
Yeniköy	23	TR1814	This study	85	3	16.7	18.28	16.34	88	88	78	161.3	2.5	1046	2080	211	1.3	0.99
Yeniköy	23	TR1815	This study	85	15	18.4	18.28	16.34	80	88	78	75.8	1.8	1046	2628	240	1.3	0.99
Badınca	28	TR1816	This study	105.1	2	23	24.97	19.4	111	125	93	45.0	1.9	2144	1060	442	0.72	0.59
Badınca	28	TR1817	This study	105.1	28	23	24.97	19.4	110	125	93	123.7	3.2	2144	1060	442	0.72	0.59
Badınca	28	TR1818	This study	105.1	29	21.5	24.97	19.4	105	125	93	38.1	1.8	2144	1416	300	0.72	0.59
Çay Sokak	10	1776	Heineke et al. (2019)	44.6	157	16	14.82	17.39	72	69	83	142.3	1.0	3838	1416	418	1.65	0.87
Armutlu	1	15710	Heineke et al. (2019)	5.6	70	21.9	14.85	26.27	101	71	132	142.3	1.0	930	1416	722	0.7	0.7
Yeşilkavak	18	15715	Heineke et al. (2019)	73.7	3	22.1	23.75	19.81	102	117	95	104.3	2.2	3274	1439	355	1.58	1.48
Başıktaş Der.	8	15716	Heineke et al. (2019)	32.9	104	16	15.17	16.41	73	66	78	26.2	0.8	1203	4343	372	1.35	0.9
Başıktaş Der.	8	15717	Heineke et al. (2019)	32.9	7	9.4	15.17	12.81	73	66	78	165.5	5.4	548	4343	70	1.35	0.9
Kazmpaşa	Za	15719	Heineke et al. (2019)	13.65	12	25.5	15.16	25.63	122	116	127	165.5	5.4	1192	4343	286	0.87	0.59
Yenikuruder	2	15720	Heineke et al. (2019)	12.8	104	24.2	19.06	27.03	114	93	137	75.8	1.8	1023	2628	760	0.87	0.59
İrlamaz Çayı	7	15721	Heineke et al. (2019)	22.7	36	21.7	15.17	23.02	98	77	111	133.7	2.7	1539	1454	289	1.1	0.71
Cevizdere	4	1471	Heineke et al. (2019)	17.7	28	22.4	18.09	24.52	107	101	122	133.7	2.7	1998	1454	533	0.99	0.81
Akçapınar	9	1472	Heineke et al. (2019)	35.4	69	17.1	15.16	20.85	76	66	101	161.3	2.5	1987	2080	356	1.41	0.44
Kabazlı	17	1473	Heineke et al. (2019)	69	1	24.7	18.97	17.33	82	92	81	84.7	2.0	2376	631	331	1.74	1.33
Çaltılı	13	1171*	Buscher et al. (2013)	56.3	80	21.5	18.33	24.49	103	90	124	58.0	1.0	2600	631	615	1.91	1.28
Kabazlı	17	1173*	Buscher et al. (2013)	69	28	18.6	18.97	17.33	84	92	81	138.8	0.7	2141	1282	279	1.74	1.33
Yeşilkavak	18	1174*	Buscher et al. (2013)	73.7	42	24.9	23.75	17.33	102	117	95	138.8	0.7	3274	1282	355	1.58	1.48
Gümüş Çayı	16	1175*	Buscher et al. (2013)	65.4	59	24.7	22.41	19.81	116	112	100	80.5	1.7	3986	1282	634	1.86	1.33

496

497 *Table 4. Geomorphic and geological variables by sample and river.*

498

499 Firstly, if the along strike geomorphic character of the uplifted footwall of the Gediz
 500 Graben boundary fault is examined, it is clear that the mean catchment relief
 501 (Figure 6A), maximum incision (Figure 6B) and mean catchment slopes (Figure
 502 6C) of sampled catchments are variable (Figure 6B) but overall follow the trend in
 503 fault throw rate (Figure 6B) with minima all metrics coinciding with the mapped
 504 fault segment boundaries (dashed lines, Figure 6). Indeed, the clear relationship
 505 along strike of the geomorphic expression of active faulting was partly used by
 506 Kent et al. (2016) to determine long-term uplift rates along the Gediz Graben
 507 boundary fault (Figure 6B). If the relief (Figure 6A) and slope (Figure 6C) above
 508 and below the knickpoints are considered separately, the same overall trends are
 509 apparent but with higher relief and slopes downstream of the knickpoint in the
 510 central and western parts of the range. This result is expected as the transient
 511 wave of incision causes gorge formation and hillslope steepening as it propagates
 512 through the river system. In the eastern part of the range this relationship is
 513 apparently inverted with higher slopes and relief above the knickpoint. Although,
 514 fewer data are available in this zone.

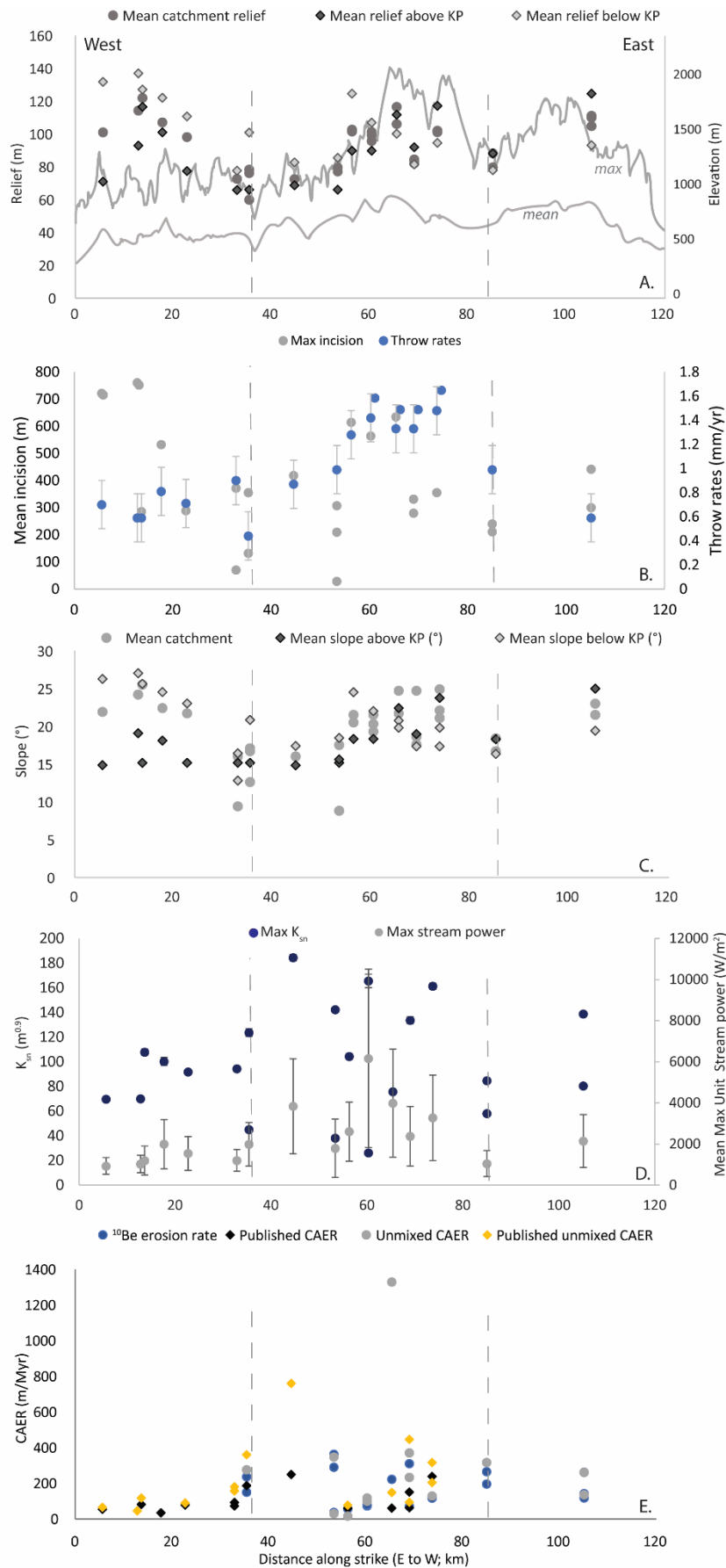
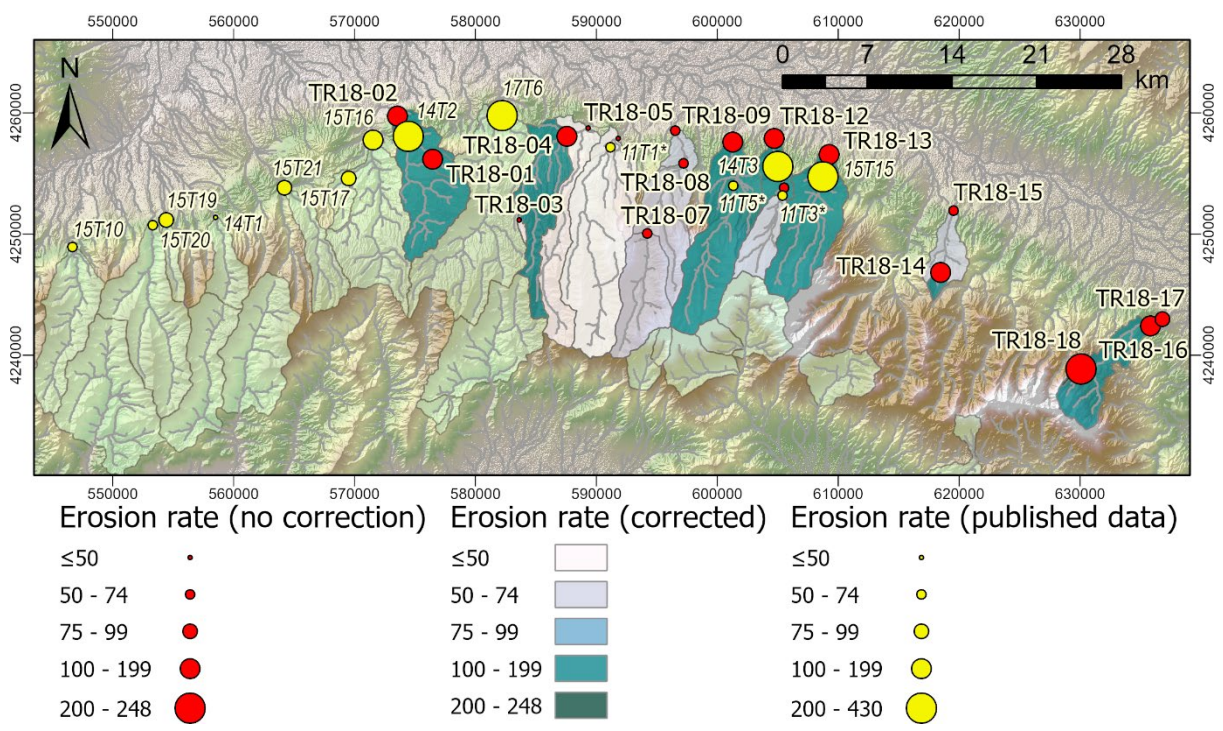


Figure 6. Along strike trends in geomorphic variables and CAERs. Dashed lines show fault segment boundaries from Kent et al. (2018). A) Catchment relief (mean whole catchment, mean above and below the tectonic knickpoint, and elevation mean and maximum swath profiles; B) Channel incision in the transient reaches and long-term throw rates (Kent et al., 2017); C) Total mean catchment slope and mean slope above and below the knickpoint; D) Normalised steepness index and maximum unit stream power, and E) Catchment-averaged erosion rates. Note: error bars are shown where greater than symbol size.

549 When the normalised steepness index in the transient reach is plotted along strike
 550 then the highest steepness indices are present in the centre of the fault array
 551 (Figure 6D), where current fault slip rates are highest. Maximum stream powers
 552 also cluster within the central fault segment, although it is important to
 553 acknowledge that lower values of steepness index and stream power are also
 554 present in the central part of the fault zone.

555 When the along strike trends in CAER are considered there is an increase from the
 556 westernmost sample (54.5 mMyr⁻¹) into the centre of the range (250 mMyr⁻¹) for
 557 both the raw CAER and burial-corrected rates (Figures 6E and 7). However, rates
 558 then decrease again along two large river systems in the centre of the range
 559 (TR18-06 – Catili and TR18-09 – Bozdağ) before increasing again along the
 560 eastern part of the range. This decrease in erosion rates in the centre of the fault
 561 appears unexpected given these catchments are experiencing the highest uplift
 562 rates. When the unmixed CAER are plotted (Figure 6E), a clearer pattern of lower
 563 rates at the fault tips and higher rates in the centre of the range appears although
 564 the CAER in the centre of the fault are still generally subdued.



565
 566 *Figure 7. Map of showing the catchment-averaged erosion rates along the Gediz Graben*
 567 *Boundary Fault. Yellow circles show previously published data (Buscher et al., 2013;*
 568 *Heineke et al., 2019); while red circles show rates derived here but without correction for*
 569 *sediment storage and recycling. Rates corrected for these factors are shown by the shading*
 570 *of the catchment areas.*

571 Interestingly, there are also differences in the CAER along individual sampled river
572 systems with both decreasing and increasing erosion rates downstream being
573 present (Figure 7). For example, and as expected, CAER increases along the
574 Kabazlı River from 59 mMyr⁻¹ upstream of the Gediz Detachment fault to 157
575 mMyr⁻¹ at the boundary fault (Figure 7). By contrast, along the Badınca River
576 (samples TR18-16 to TR18-18; easternmost river), burial-corrected erosion rates
577 decrease downstream from ~ 250 mMyr⁻¹ in the headwaters to 86 mMyr⁻¹
578 upstream of the boundary fault. These data suggest that CAER do not scale simply
579 with tectonic rates (c.f. Roda-Boluda et al., 2019) and may be influenced by
580 factors such as sediment storage and contrasts in bedrock erodibility, which we
581 evaluate below.

582 Secondly, the different CAER can also be compared with a range of topographic
583 metrics that have previously been shown to correlate positively with erosion rates
584 in previous studies such as relief and slope (i.e., Abbühl et al., 2011; Miller et al.,
585 2013; Bellin et al., 2014; Kober et al., 2015). However, when the burial-corrected
586 mixed rates (but not unmixed for low erosion rate areas) and published CAER data
587 are plotted against mean catchment slope, topographic relief (150 m radius) and
588 maximum incision depth upstream of the sample site there are no trends
589 (Supplemental Figure 1).

590 By contrast, when these erosion rates are compared to the maximum upstream
591 unit stream power there is a significant ($P < 0.05$) positive linear trend with erosion
592 rate in the published data from Heineke et al., (2019)(Figure 8A; $r^2 = 0.8$). There
593 are also significant ($P < 0.05$) positive linear ($r^2 = 0.6 - 0.9$) relationships between
594 erosion rates and steepness index for the published data of Buscher et al. (2013)
595 and Heineke et al. (2019)(Figure 8B) and a weak but significant linear relationship
596 between erosion rates and throw rate on the graben boundary fault (figure 8C;
597 $r^2 = 0.2$). It is also noticeable that CAER expressed as m/Ma are lower than the
598 slip rates on the basin bounding fault, particularly towards the centre of the fault,
599 where displacement rates are 2 mm/yr (i.e. 2000 m/Ma) (Figure 6B).

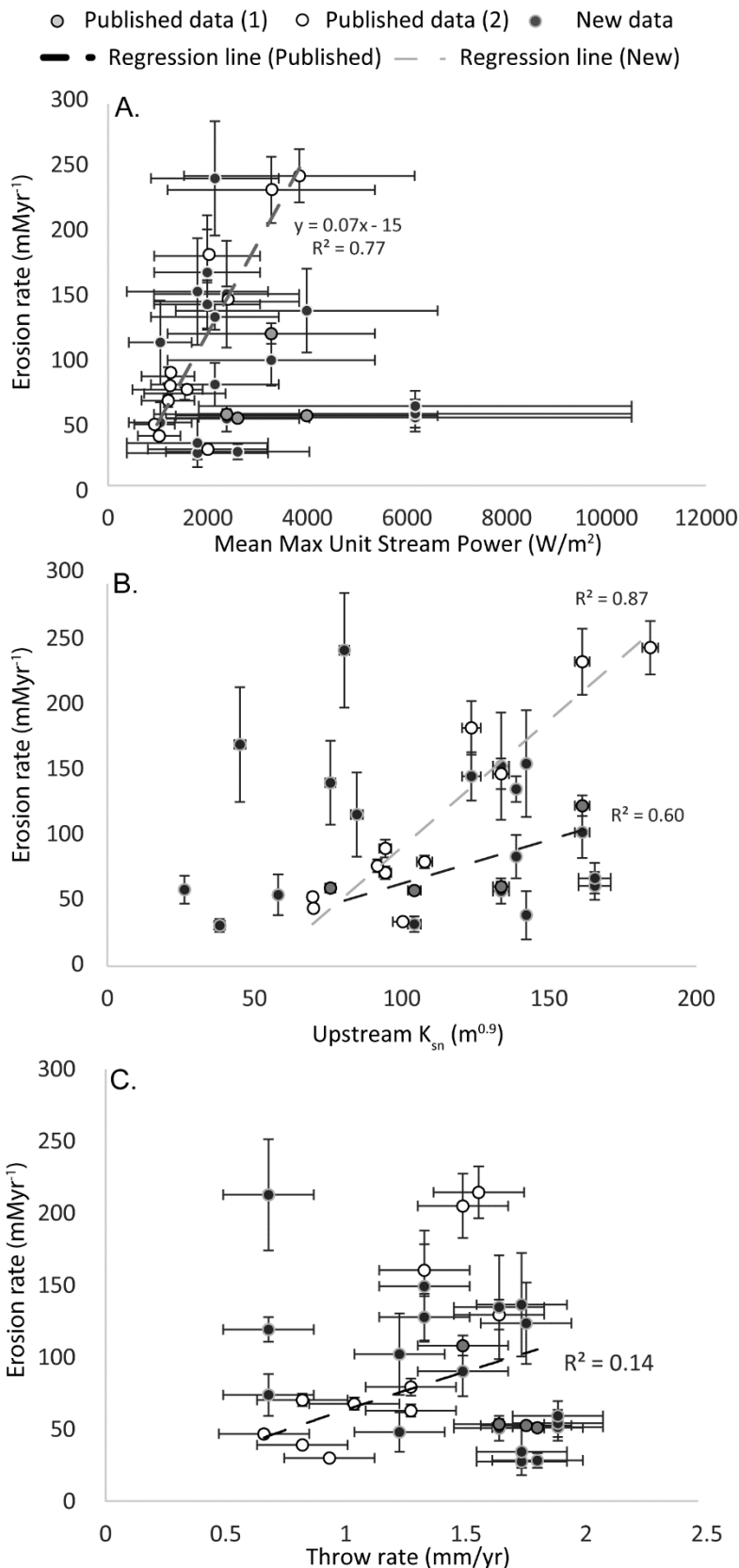


Figure 8. Comparison of geomorphic variables A) mean maximum unit stream power and B) normalised steepness index upstream, and C) throw rate on the Gediz Graben Boundary Fault against catchment-averaged erosion rates for previously published data (1: Buscher et al., 2013; 2: Heineke et al., 2019) with internal uncertainty and for all samples collected here corrected for burial and sediment storage with calculated errors but not unmixed further.

628

629 Thirdly, the unmixed CAER that represent erosion rates only in the transient reach
 630 of the rivers can be compared with the same metrics. When these rates (which

631 include published data as well as the new data determined here) are plotted
632 against mean catchment slope, topographic relief and maximum incision depth
633 upstream of the sample location, again there are no clear or significant trends
634 (Supplemental Figure 2). However, when unmixed CAER are compared to the
635 upstream maximum unit stream power there is a broad positive trend but with
636 only a very weak correlation (Figure 9A). Although when the Bozdağ samples are
637 removed as potential outliers, because this river has very high stream power yet
638 low erosion rates in the centre of the fault, a significant ($P < 0.05$) linear
639 regression line with an $r^2 = 0.25$ can be plotted. Similarly, there is no trend
640 between Ksn and CAER, but if the Gumusi cay sample is excluded as an outlier,
641 there is weak ($r^2 = 0.2$) but significant ($P < 0.05$) positive relationship between
642 erosion rates and steepness index with the best fit regression being an exponential
643 trend (Figure 9B). When all unmixed CAERs are plotted against fault throw rate
644 there is no trend; however, when the samples from the detachment are removed
645 so that only samples close to or at the boundary fault are retained there is a weak
646 ($r^2 = 0.1$) but not significant ($P > 0.05$) positive power law relationship between
647 these two variables (Figure 9C).

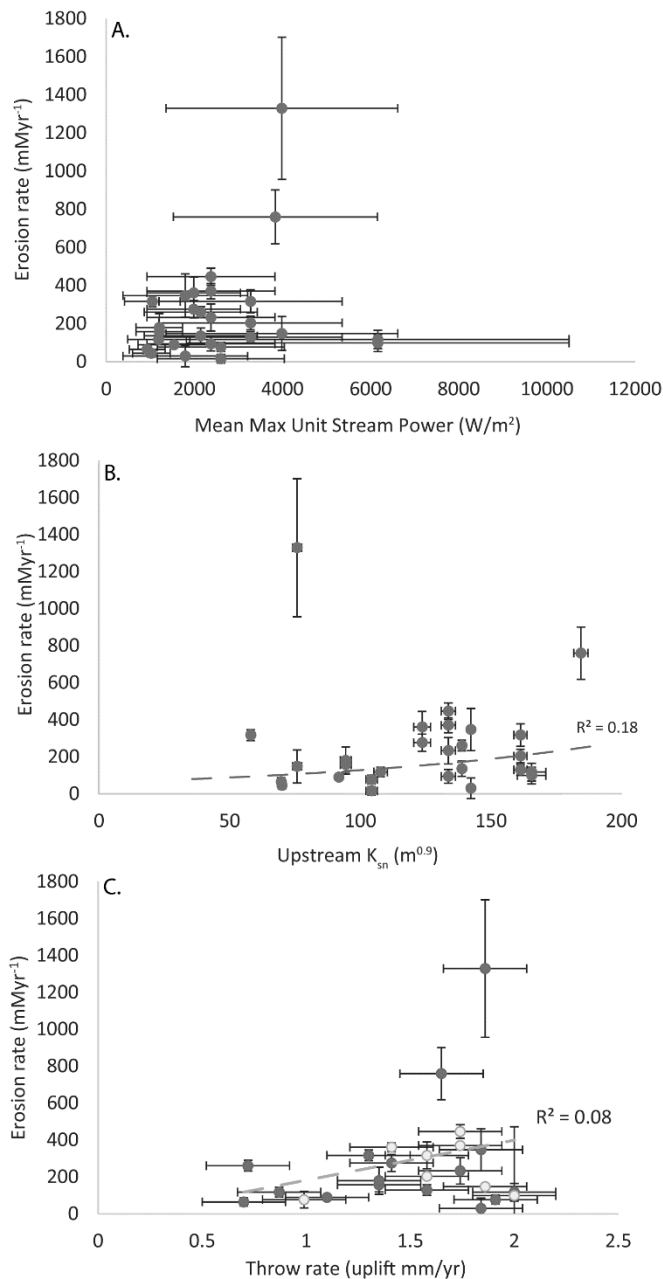


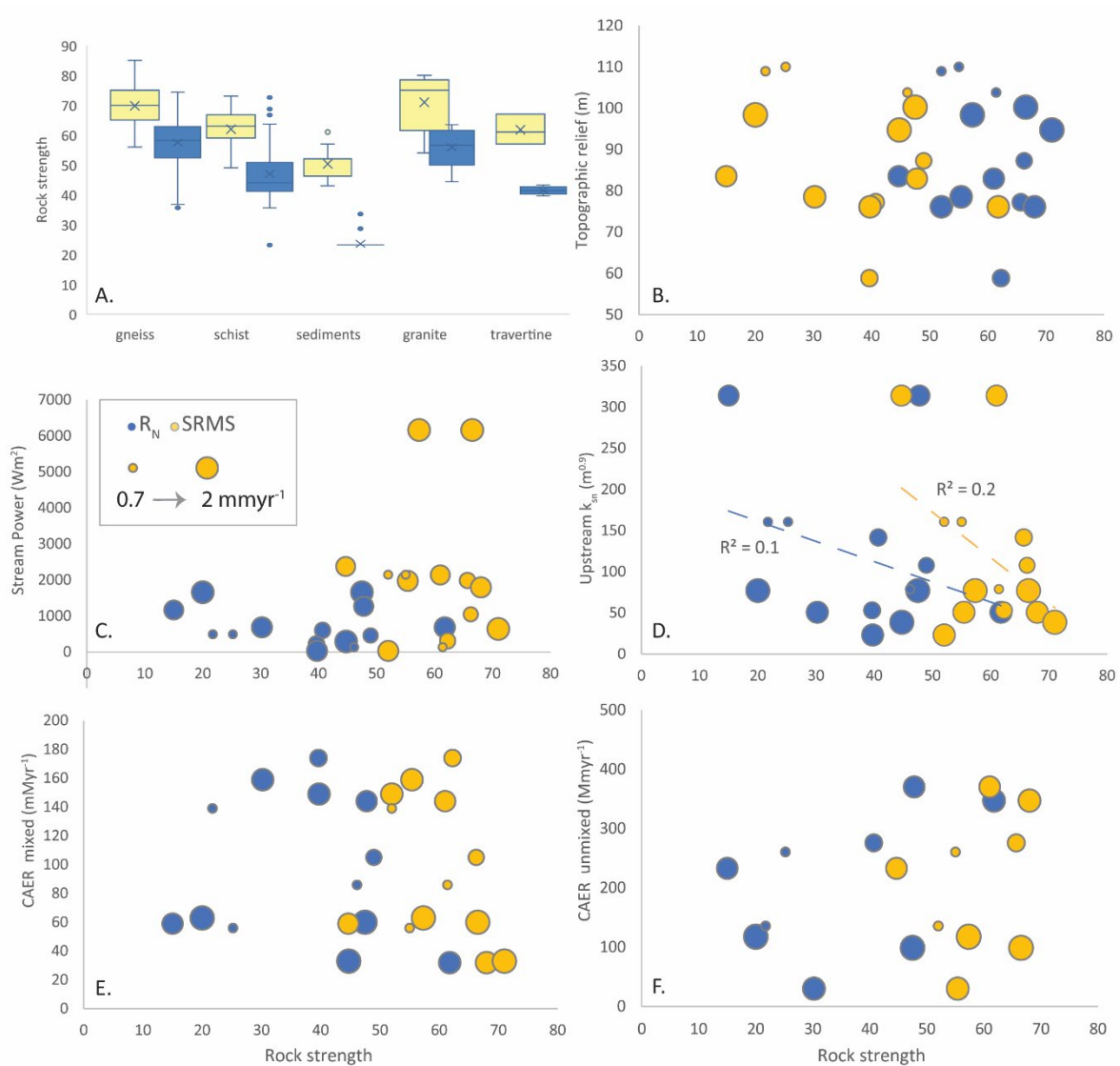
Figure 9. Comparison of geomorphic variables A) maximum stream power and B) upstream steepness index, and C) throw rate on the Gediz Graben Boundary Fault against catchment-averaged erosion rate for previously published data and for samples collected here unmixed to remove the effect of the low erosion rate areas above the knickpoint. On C data have been separated into samples at the range front (dark) and at the detachment fault (light) to investigate the potential difference in erosion rates depending on the bedrock lithology.

671

672 **4.4. Relationship between rock strength, geomorphology and**
 673 **erosion rates**

674 In order to assess the impact that the different bedrock lithologies have on the
 675 geomorphic response in the study region, the erosion rates for the different
 676 catchments can be compared to measurements of bedrock strength. The bedrock
 677 of the Bozdağ range can be broadly divided into the metamorphic lithologies of
 678 the Menderes Massif and the unconformably overlying Miocene and younger
 679 sediments. The metamorphic rocks are primarily composed of moderately strong
 680 to strong (c.f., Selby, 1980) schists, gneisses and granites where the SRMS > 60

681 (Figure 10A) (c.f. Kent et al., 2021). By contrast, the syn-tectonic sandstones
 682 and conglomerates are weak to very weak (SRMS < 50). Therefore, if rock
 683 strength is the main control on CAER then the harder metamorphic rocks should
 684 be eroding at a lower rate than the softer sediments.



685

686 *Figure 10. A) Total Schmidt hammer rebound and SRMS for the main lithologies present*
 687 *in the study area. Schmidt hammer and SRMS calculated over 2 km upstream of the*
 688 *sample locations on the six main study rivers plotted against: B) topographic relief; C)*
 689 *maximum stream power; D) upstream normalised steepness index; E) catchment-*
 690 *averaged burial corrected erosion rates , and F) unmixed erosion rates for the transient*
 691 *reach of the rivers. On B-E the size of the circle proportionally represents the throw rate*
 692 *at the range front where the largest circles equal 2 mm/yr.*

693

694 Across the study region the strong metamorphic rocks are located south of the
695 Gediz Detachment in the upland regions of the Bozdağ range, while the weak
696 sedimentary rocks are mainly to the north, i.e., a soft over hard transition as
697 represented in many landscape evolution models (e.g., Forte et al., 2016).
698 Interestingly though when both measures of rock strength upstream of sample
699 locations are compared to geomorphic variables such as relief (Figure 10B) and
700 stream power (Figure 10C) there are no trends between the variables. This
701 suggests that rock strength alone does not control relief or stream power. By
702 contrast, there is a weak ($r^2 \leq 0.2$) negative linear relationship between rock
703 strength (SRMS and Schmidt hammer rebound) and the upstream steepness index
704 suggesting that rivers are on average less steep when the rocks are harder.
705 However, this is not significant for either RMS or Schmidt Hammer rebound ($p >$
706 0.05) and is the opposite of the relationship that we would expect where the river
707 is steeper in harder rocks. Furthermore, when CAERs are compared to the
708 upstream rock strength, there is no clear relationship either for mixed or unmixed
709 rates with both strong and weak rocks resulting in a similar range of CAERs
710 (Figures 10 E and F). Finally, there are no clear trends of these variables with
711 uplift rate on the fault as indicated by the size of the symbols on figures 10B-F.

712

713 **5. Discussion**

714 ***5.1. What controls erosion rates along the margin of the Gediz*** 715 ***Graben?***

716 The geomorphology of the Bozdağ Range is shaped by the uplift along the Gediz
717 Boundary fault and concomitant incision of the bedrock rivers resulting from the
718 linkage of the boundary faults at ~ 0.8 Ma (Kent et al., 2017). Therefore, it is
719 expected that there should be scaling relationships between various landscape
720 metrics, uplift and erosion, similar to other regions around the world. For example,
721 many studies show a positive relationship between CAER and catchment slope
722 (i.e., Bellin et al., 2014; Rossi et al., 2017; Rosenkranz et al., 2018; Roda-Boluda
723 et al., 2019) as well as positive relationship with channel steepness (Harkins et
724 al., 2007; Cyr et al., 2010; DiBiase et al., 2010; Miller et al. 2013; Bellin et al.,
725 2014; Rossi et al., 2017), which has been shown to be linear at low rates and
726 steepness but becoming non-linear above a threshold steepness index. Related to
727 landscape steepness is relief, which can either be measured as topographic relief

728 across the catchment, or following Roda-Boluda et al. (2018) as maximum incision
729 depth (i.e., maximum local relief) along the river. In both measures, CAER have
730 previously been shown to have a positive relationship with these factors. For
731 example, Bellin et al. (2014) demonstrated a positive linear relationship with relief
732 and Roda-Boluda et al. (2018) a positive power law relationship with maximum
733 incision depth. This is not unexpected assuming little pre-existing topography, as
734 areas of higher relief will have had more material eroded than areas of lower relief
735 over the same time span, thus erosion rates should be higher where relief is
736 higher. Though it is important to note that in general hillslopes have longer
737 response times than rivers to changes in base-level (Simpson and Schlunegger,
738 2003; Schlunegger et al., 2013).

739 Unexpectedly, these trends appear not to hold true along the Bozdağ range either
740 locally or catchment-wide, with no strong trends between erosion rates and
741 average catchment slope, catchment relief or incision depth in either burial
742 corrected CAER or unmixed for just the transient reach. While there are weak
743 positive relationships observed in the data between CAER and normalised
744 steepness index in the channel upstream of the sample point, this varies between
745 a linear relationship for the whole CAER (not significant) and a weak but significant
746 exponential for the transient reach only. The strongest and most significant of
747 these weak trends is the linear relationship between the stream power and CAER
748 (both burial-corrected and unmixed) albeit with larger uncertainties on the stream
749 power data. These last two observations indicate that at the catchment scale and
750 at the precision of our data, the rivers are broadly in line with a simple form of the
751 stream power law, which is linear and $n = 1$ (Whipple and Tucker, 1999) where E
752 $\approx KA^mS^n$ and is consistent with the analyses of Kent et al. (2021).

753 When CAER are compared to throw rates it is striking that erosion rates are around
754 an order of magnitude lower than uplift rate. Given the presence of knickpoints
755 and a documented transient landscape response (Kent et al., 2021, 2017)
756 demonstrating that this region is not in topographic steady state, this relationship
757 is to be expected. As a result the Bozdağ region will be experiencing surface uplift
758 (Figures 8C and 9C). Yet, there are only weak positive relationships between the
759 throw rate and CAERs, when corrected for sediment storage and for the presence
760 of low relief zones. Additionally, it is striking that these relationships are only
761 significant for the burial-corrected CAERs not for the unmixed CAERs. However,

762 this apparent contradiction is consistent with the documented fault linkage. After
763 a fault linkage event, the highest erosion rates should be present in the linkage
764 zones where the previous minimum in fault throw (as these were the tips of
765 individual faults) have had to rapidly increase to achieve the ideal fault profile
766 (Kent et al., 2016), higher uplift rates will also result in increased erosion in these
767 zones. This will also result in the part of the fault with the highest slip rates
768 experiencing lower erosion rates and as a result in the transient reaches throw
769 rate will not scale with erosion rate. Interestingly, at a catchment level the CAERs
770 do scale with throw rate but the correlation is weak perhaps suggesting that prior
771 to fault linkage throw rate did correlate with erosion rate.

772 A number of factors may cause the scatter and the weak correlations in these
773 data, which we explore below. One complication to consider is that the results
774 could be affected by sediment storage or non-uniform erosion as a result of
775 landsliding (e.g., Binnie et al., 2006; Kober et al., 2012; Roda-Boluda et al., 2018).
776 Neither of these factors appear to be likely along the Bozdağ range as firstly, the
777 potential effect of sediment storage has been corrected through the inclusion of
778 ²⁶Al CRN data. Secondly, there is little evidence for significant landsliding in the
779 study region to deliver material with sufficiently low ¹⁰Be concentrations to perturb
780 the measured river sediment concentrations. Incomplete sediment mixing could
781 also explain the scatter in the data, while the measured CRN concentrations of
782 repeat samples along several river systems are within 2σ error, we have limited
783 data across the entire range to fully assess this issue, which has been shown to
784 be a complicating factor in mountainous catchments elsewhere (Binnie et al.,
785 2006).

786 Alternatively, the presence of inherited topography may play a significant role in
787 the landscape response to uplift (c.f., Densmore et al., 2009). This explanation is
788 supported by the clear imprint of the fault segments in the topographic metrics
789 and the observation that in the eastern part of the range higher slopes and relief
790 are found upstream of the tectonic knickpoint (Figure 6), despite transient river
791 incision downstream of the knickpoint. Therefore, inherited topography might
792 explain the disconnect between erosion rates and catchment wide variables such
793 as slope and relief and potentially the variability in the CAER derived from the five
794 samples taken from at or above the knickpoint. Yet if this explanation was the
795 only confounding factor, the unmixed CAER data should show stronger correlations

796 with stream power and steepness index in particular, as the effect of low relief/low
797 erosion rate zones have been accounted for in this calculation, and burial-
798 corrected CAER for the whole catchments might be expected to show relationships
799 with catchment mean slope or relief, which they do not. Therefore, another
800 explanation for the spread in the data could be the influence of a strong lithological
801 contrast within the catchments, which is discussed further below.

802

803 **5.2. The role of rock strength and lithology**

804 A number of recent models have explored the impact of lithological variability on
805 river evolution and erosion rates that could be used to understand the
806 relationships between CAER and the topographic metrics. Forte et al.'s (2016)
807 model of using two distinct lithologies is highly applicable to the Gediz Graben.
808 Their work demonstrated that when soft rocks overlie hard rocks along
809 downstream dipping contact, the lithological contact becomes an important and
810 persistent topographic feature in the landscape with the contact's dip-slope being
811 preserved. This can clearly be seen in the study area as the Gediz Detachment is
812 a pervasive feature along much of the range, and in many interfluvial areas the
813 detachment is well preserved with little evidence of deep erosion.

814 Indeed, the presence of a very strong but thin cataclasite band found along the
815 low-angle Gediz Detachment was used by Heineke et al. (2019) to explain the
816 presence of low erosion rates and gentle slopes. In addition, they proposed that
817 'weak' phyllites and schists result in higher erosion rates in the centre of the range.
818 The results presented here do not support this latter point, as lower CAERs are
819 found in the centre of the range (Figure 7) and figures 10E and F show the CAER
820 are invariant with rock strength upstream of the sample location despite a two-
821 fold difference in strength between the sedimentary and metamorphic rocks
822 overall (Figure 10A) and associated differences in erodibility (Kent et al., 2021).
823 This contradiction speaks to the difficulty in accurately constraining rock strength
824 and erodibility in the field, determining the best categorisation, and linking such
825 data to observed changes in fluvial behaviour and erosion rates (e.g., Bursztyn et
826 al., 2015; Zondervan et al., 2020).

827 In addition, Forte et al.'s (2016) landscape evolution model also suggests that
828 although the lithological boundary moves downstream over time, the erosion rates

829 above and below the boundary will diverge. The soft rocks downstream will erode
830 at the imposed uplift rate while the underlying hard rocks erode at a rate lower
831 than the regional uplift rate. Another implication of Forte et al.'s (2016) landscape
832 evolution model is that CAER would be perturbed or amplified downstream as a
833 result of the lithological variation. We see that erosion rates of the underlying hard
834 metamorphic rocks are eroding at rates lower than inferred uplift rates (Figure
835 9C), consistent with the landscape evolution model outputs. However, the erosion
836 rates in the sedimentary bedrock reaches are also much lower than uplift rates at
837 the graben boundary fault (Figure 9C), and only weakly and not significantly scale
838 with throw rates on the fault.

839 Interestingly, Kent et al. (2021) demonstrated that stream power scales with uplift
840 rate in the metamorphic bedrock reaches of their six study rivers. But uplift does
841 not scale with stream power in the sedimentary reaches where sediment transport
842 appears to be more important, resulting in a difference in the fluvial response in
843 these reaches owing to the abundance of sedimentary material entering the river
844 system (Kent et al., 2021). Therefore, while erosion rates in the sedimentary
845 reaches still weakly scale with the uplift rate the influence of sediment transport
846 and hybrid or transport-limited nature of these lower reaches causes the erosion
847 rate to be lower. In this study area the lithological control on landscape evolution
848 is therefore manifested not as bedrock erodibility but in variable fluvial responses
849 that are not captured in a detachment-limited landscape evolution model. A key
850 challenge for the future is to understand how the spatially variable erosion rates
851 captured here are integrated over time to produce a coherent relief and sediment
852 flux signal.

853

854 **6. Conclusions**

855 Eighteen samples were collected for ^{10}Be and ^{26}Al cosmogenic nuclide analysis and
856 combined with a further 15 previously published ^{10}Be concentrations (Buscher et
857 al., 2013; Heineke et al., 2019) to determine catchment-averaged erosion rates
858 along strike of the well-constrained Gediz Fault system in western Türkiye. This
859 area features a significant lithological contrast where soft sediments overlie hard
860 metamorphic rocks along a moderately dipping downstream contact, a series of
861 north-flowing rivers are incising through this contact as a result of uplift along the

862 fault at rates of up 2 mmyr⁻¹ and a fault-linkage event at ~0.8 Ma (Kent et al.,
863 2017). This natural laboratory allows the results of recent landscape evolution
864 models investigating the role of such lithological contrasts to be tested. The
865 background rate of erosion of the pre-incision landscape is determined as 46 ± 46
866 mMyr⁻¹ and erosion rates within the transient reach vary from 16 – 1330 mMyr⁻¹.
867 Although, erosion rates weakly scale with unit stream power, steepness index and
868 slip rate on the bounding fault, there are no clear relationships between erosion
869 rate and relief or catchment slope. Catchment-wide and within the transient reach
870 erosion rates are an order of magnitude lower than slip rates for both metamorphic
871 and sedimentary reaches and despite a 30-fold difference in erodibility there is no
872 difference in the erosion rate between strong and weak rocks. This finding is at
873 odds with the results of landscape evolution modelling and is likely owing to the
874 influence of sediment transport on fluvial dynamics in the sedimentary reaches,
875 i.e., some rivers are not completely detachment-limited. While the weak
876 relationships between other variables remain unexplained but maybe the result of
877 incomplete sediment mixing or the influence of pre-existing topography prior to
878 the onset of the current incisional phase. These findings indicate that the erosional
879 response to uplift along an active normal fault is a complex response to multiple
880 drivers that vary spatially and temporally.

881

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886 erosion rates using both ¹⁰Be and ²⁶Al can be found at:
887 <https://github.com/angelrodes/banana1026>. Figure 5 was plotted using
888 CosmoCalc (Vermeesch, 2007) available for download from
889 <https://www.ucl.ac.uk/~ucfbpve/cosmocalc/>

890

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