

**QUANTIFYING TECTONIC, LITHOLOGICAL, AND CLIMATIC CONTROLS
ON UPLAND RIVER DYNAMICS**

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DECLARATION

Declaration of originality

I declare that this thesis, submitted for the degree of Doctor of Philosophy at Imperial College London and Associateship of the Royal School of Mines, is original research that has been completed by me, and carried out under the supervision of Dr Alex Whittaker, Dr Rebecca Bell and Prof Gary Hampson. All published and unpublished material used in this thesis has been appropriately referenced. None of this work has previously been submitted to Imperial College London or any other academic institution for a degree or diploma, or for any other qualification.

Ziqiang Zhou

20th April, 2026

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ABSTRACT

Upland rivers shape landscape evolution and release sediments into source-to-sink systems in response to tectonics, lithology, and climate. If and where their dynamics are quantified, channel geometry can be leveraged to provide long-term insights into seismic and erosion hazards. Yet, channel dynamics involve flux-dependent and threshold-stochastic processes that are yet to be fully understood. In this thesis, I probe these processes by examining tectonic, lithological, and climatic controls on channel geometry.

I first take a landscape-scale perspective to test the use of the simple, classic stream power incision model (SPIM) in tectonically active, lithological heterogeneous landscapes. I leverage the well-constrained Gulf of Corinth, Greece, as a natural laboratory to demonstrate that the classic SPIM successfully isolates tectonic and lithological signals. I then narrow my focus to two Corinth rivers to investigate how lithology, through its influence on sediment generation, modulates channel responses to tectonics via flux-dependent processes. I show that, until knickpoints migrate into abrasion-resistant rocks in these rivers, lithology dominates over tectonics in bedload sediment generation via abrasion. I further suggest that lithology-influenced sediment supply determines the excess shear stress required for equilibrium and influences how that excess shear stress is partitioned between channel width and slope adjustments. Finally, I zoom out to examine the role of threshold-stochastic processes on incision by comparing channel width and slope responses to discharge-variability proxies on a global scale. The results challenge the traditional slope-centred perspective and instead emphasise the sensitivity of channel width, likely because bank erosion requires a higher threshold than bed erosion.

In a nutshell, this thesis sheds new light on upland river dynamics in terms of flux-dependent incision and threshold-stochastic processes. In particular, I clarify when the classic SPIM is most effective and highlight the importance of channel-width dynamics, with important implications for assessing seismic and erosion hazards.

CONTENTS

QUANTIFYING TECTONIC, LITHOLOGICAL, AND CLIMATIC CONTROLS ON UPLAND RIVER DYNAMICS	1
ABSTRACT	2
ACKNOWLEDGEMENTS	3
CONTENTS	5
1 INTRODUCTION	16
1.1 Why are fluvial landscapes important?	16
1.2 Research background	18
1.3 Aims and objectives	26
1.4 Thesis outline	28
2 UNRAVELLING TECTONIC AND LITHOLOGICAL EFFECTS ON TRANSIENT LANDSCAPES IN THE GULF OF CORINTH, GREECE ¹	31
Abstract	31
2.1 Introduction	32
2.2 Geological background	34
2.3 Methods	38
2.4 Results	47
2.5 Discussion	59
2.6 Conclusions	70
2.7 Acknowledgements	71
2.8 Supplementary material	72

3 DECODING TECTONIC AND LITHOLOGICAL CONTROLS ON SEDIMENT EXPORTED FROM
FOOTWALL CATCHMENTS: IMPLICATIONS FOR HANINGWALL STRATIGRAPHY ¹ 90

Abstract	90
3.1 Introduction.....	91
3.2 Study area.....	92
3.3 Methods	96
3.4 Results and analysis.....	102
3.5 Discussion	110
3.6 Conclusions.....	116
3.7 Acknowledgements	117
3.8 Supplementary material.....	118

4 SEDIMENT SUPPLY CONTROLS ON CHANNEL MORPHOLOGICAL ADJUSTMENTS TO
TECTONICS AND LITHOLOGY ¹ 128

Abstract	128
4.1 Introduction.....	129
4.2 Flux-dependent incision	130
4.3 Study area.....	132
4.4 Methods	135
4.5 Results	138
4.6. Discussion	142
4.7 Conclusions.....	147
4.8 Acknowledgements	148
4.9 Supplementary material.....	148

5 DISCHARGE VARIABILITY CONTROLS CHANNEL WIDTH BUT NOT SLOPE ¹	153
Abstract	153
5.1 Introduction	154
5.2 Methods	156
5.3 Results	158
5.4 Discussion	160
5.5 Conclusions	162
5.6 Acknowledgments	162
5.7 Supplementary material	162
6 DISCUSSION	173
6.1 Synthesis	173
6.2 Cross-cutting themes	175
6.3 Future work	180
7 CONCLUSIONS	184
8 REFERENCES	187

List of figures

- Figure 2.1. (a) Hillshaded map of Gulf of Corinth showing faults and catchment boundaries. Inset map shows regional setting within the tectonic framework of the Aegean. Catchment abbreviations (blue): AS-Asopos, DE-Deveni, FO-Fonissa, KR-Krios, KT-Krathis, LA-Ladopotous, ME-Meganitis, PH-Phoenix, RA-Razanis, SE-Selinoutous, SK-Skoupeiko, SI-Sithas, VO-Vouraikos, XE-Xerias, ZA-Zapantis. Active fault abbreviations (red): AIG-Aigion fault, DER-Deveni fault, E. ELK-East Eliki fault, E. XYL-East Xylokastro fault, LYK-Lykoporia fault, N. KIA-North Kiato fault, PER-Perachora fault, PSA-Psathopyrgos fault, NER-Neo Erineos fault, W. ELK-West Eliki fault, W. XYL-West Xylokastro fault. Inactive faults are shown as dashed black curves. (b) cross-section A-A' in (a) showing tectono-sedimentary evolution of the study area (from Ford et al., 2017). 36
- Figure 2.2. (a) Simplified geological map of the studied catchments in the Gulf of Corinth, shown with faults (active-red curves, inactive-dashed black curves) and knickpoints (see section 4.2), and (b) compiled uplift rates (see supplementary Table 2.S1) plotted along the dashed line B-B' in (a). 37
- Figure 2.3. The relationship between topography and lithology. Left panels are maps of (a) k_{sn} and (c) hillslope angle. Right panels show box plots of (b) k_{sn} and (d) hillslope angle for the lithologies in the study area, with colours and lithology abbreviations in accordance with Figure 2.2. 47
- Figure 2.4. Box-and-whisker plots showing along-strike variations (cf. Figure 2.2) in (a) channel steepness and (b) hillslope angle, along with the distribution of the faults. 49
- Figure 2.5. (a-e) Long profiles of trunk rivers crossing faults, coloured by lithology (cf. Figures 2.2 and 2.3). (f) Along-strike variation in knickpoint height. The character corresponds to catchments shown in Figure 2.1. In profiles below 950 m, the lower, middle, and upper sets of tectonic knickpoints are referred to as 'knick1' (blue squares), 'knick2' (red triangles) and 'knick3' (yellow diamonds), respectively, and prominent lithological knickpoints are shown as grey circles; in profiles above 950 m (grey bars), knickpoints are marked as white circles; more details are provided in section 3.2 and supplementary material. Fault abbreviations: AIG-Aigion fault, DER-Deveni fault, E. ELK-East Eliki fault, LYK-Lykoporia fault, N. KIA-North Kiato fault, PER-Perachora fault, NER-Neo Erineos fault, W.ELK-West Eliki fault, W.XYL-West Xylokastro fault. 51
- Figure 2.6. Box-and-whisker plots show lithology-grouped k_{sn} (a) and slope (b) for the lowest trunk segment in catchments with 'knick2' knickpoints. In the same segment, variations between different lithologies are used to evaluate the impact of lithology; for the same lithology, variations across different segments are used to evaluate the impact of tectonics (cf. Figure 2.3 and Figure 2.4). Note the lowest segment for VO and KR is the segment below the 'knick1' knickpoint. 55
- Figure 2.7. (a) Boxplot of the range of relative erodibility (i.e., the reciprocal of limestone-normalized k_{sn}) for different lithologies; (b) Comparison between observed (open circles) and

modelled (red crosses) positions of migrating knickpoints; more in supplementary material.	56
Figure 2.8. Optimized θ (θ_{opt}) for post-ca.780 ka river profiles in both a uniform erodibility (blue) and lithology-dependent erodibilities (orange) scenarios.	57
Figure 2.9. (a) Knickpoint height plotted against knickpoint initiation age for all identified tectonic knickpoints; (b) Along-strike age variations for knickpoints in both ‘knick1’ and ‘knick2’ sets, with distribution of the faults.	58
Figure 2.10. Possible fault growth and linkage history reconstructed from knickpoint initiation ages and stratigraphic thickness variations from ca. 780 ka to present (modified from Nixon et al., 2016 and Gawthorpe et al., 2022). Opaque red triangles and blue squares in (a-f) are the modelled knickpoint positions at the start of each time slice, and their transparent counterparts in (f) mark the current knickpoint positions. Note the areas where faults are interpreted to grow and link are highlighted by the two rectangles. Key faults considered here are marked in red, with their yet-to-be-activated segments shown as dashed lines.	67
Figure 2.11. Schematic diagram showing contrasting landscape responses to faults growth and linkage under high uplift and low erodibility (a), high uplift and high erodibility (b) and low uplift and high erodibility (c). Note the three catchments that differ in uplift rate and erodibility have similar drainage area.	69
Figure 2.S1. Uplift rate variations across fault strike for the Sithas/SI (a) and the Vouraikos/VO (b) (modified from de Gelder et al., 2019 and Zondervan et al., 2020b). Note the linear decay model with a rate of $\sim 0.08 \text{ mmyr}^{-1}\text{km}^{-1}$ (dashed red lines) is mostly effective within the range of $\sim 12 \text{ km}$ upstream of the fault, beyond which the decay rate decreases significantly. In the SI, the fulcrum position is likely located $>30 \text{ km}$ upstream of the fault using the decay rate of $\sim 0.025 \text{ mmyr}^{-1}\text{km}^{-1}$; In the VO, the potential fulcrum position is likely located $>25 \text{ km}$ upstream of the fault where alluvial fill appears (see their Figure 2 in Zondervan et al., 2020b).	72
Figure 2.S2. Knickpoints in the XE catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...	73
Figure 2.S3. Knickpoints in the RA catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same color are from the same stage. ...	74
Figure 2.S4. Knickpoints in the ZA catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...	75
Figure 2.S5. Knickpoints in the AS catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...	76

Figure 2.S6. Knickpoints in the SI catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...77

Figure 2.S7. Knickpoints in the FO catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...78

Figure 2.S8. Knickpoints in the SK catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...79

Figure 2.S9. Knickpoints in the DE catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...80

Figure 2.S10. Knickpoints in the SK catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...81

Figure 2.S11. Knickpoints in the KR catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...82

Figure 2.S12. Knickpoints in the LA catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...83

Figure 2.S13. Knickpoints in the VO catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...84

Figure 2.S14. Knickpoints in the KE catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...85

Figure 2.S15. Knickpoints in the SE catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...86

Figure 2.S16. Knickpoints in the ME catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...87

Figure 2.S17. Knickpoints in the PH catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage. ...88

Figure 3.1. Gulf of Corinth in maps of hillshaded relief (a), lithology (b), and precipitation (c), with active faults and the two study catchments highlighted. Active fault abbreviations (red):

AIG-Aigion fault, DER-Deveni fault, E. ELK-East Eliki fault, E. XYL-East Xylokastro fault, LYK-Lykoporia fault, N. KIA-North Kiato fault, PER-Perachora fault, PSA-Psathopyrgos fault, NER-Neo Erineos fault, W. ELK-West Eliki fault, E. XYL-East Xylokastro fault (after [Nixon et al., 2024](#)). Geological unit abbreviations: *p-con sed.* – poorly consolidated sediment; *cong.* – conglomerate; *sand.-silt.* – sandstone-siltstone; *lime.* – limestone. 93

Figure 3.2. Main characteristics of two study catchments in maps (a, b) and river profiles (c, d), shown with trunk channels (red), key tributaries (black) and field localities marked by white, yellow, and red dots (sediment grain size measurement in trunk, tributary, and on hillslopes, respectively), as well as cyan squares (rock strength measurement). First-order variations in uplift rates are estimated using the linear back-tilting model derived from well-constrained faults in the Corinth rift ([de Gelder et al., 2019](#); [Zondervan et al., 2020](#)). The triangles mark the boundary between the upstream and downstream zones in the trunk channel. 95

Figure 3.3. Field photographs illustrating hillslope and channel processes (A, C, E), their corresponding sediments (B, D, F), rock strength measurements (G), and sieving (H). 97

Figure 3.4. Schmidt hammer rebound values (A) and uniaxial compressive strength (B) of the four geological units in the studied catchments, shown with field photos (C, D, E, F). Red text refers to the mean representative Schmidt hammer rebound values taken nearby ([Table 3.S1](#)). 102

Figure 3.5. The Phoenix river k_{sn-p} map (a), gradient map (b) and mass-wasting map (c). Tributary-averaged k_{sn-p} , gradient (with standard deviations as error margins; d, e), and mass-wasting density (f), along the trunk channel. 103

Figure 3.6. The Sithas river k_{sn-p} map (a), gradient map (b) and mass-wasting map (c). Tributary-averaged k_{sn-p} , gradient (standard deviations as error margins; d, e), and mass-wasting density (f), along the trunk channel. 104

Figure 3.7. Representative tributaries in map view (A, C, E, G, I, K) and their trunk river profiles (B, D, F, H, J, L). Tributaries from the Phoenix and Sithas catchments are shown in the left (A-F) and right (G-L) panels, respectively (see [Figure 3.5](#) and [Figure 3.6](#) for locations and legends). 105

Figure 3.8. Sediment delivery from weathering (solid lines) and mass-wasting (dotted dashed lines) processes on hillslopes and from tributary channels (dashed lines) (A), shown with photos of their characteristic sediments (B, C, D, E). 106

Figure 3.9. Trunk-channel sediment grain size for both the Sithas and Phoenix rivers. The shaded area represents the grain size range of hillslope sediments available for channel transport. 106

Figure 3.10. Downstream variations in trunk-channel sediment D_{84} for the Phoenix (A) and Sithas (B) catchments, shown with hillslope gradients in a 1,000 m-wide swath (pink bands) and key tributary junctions (dashed lines) along trunk channels coloured by lithology. Note, the Wolman grain size is used for both sieved (red squares) and Wolman (purple diamonds) localities to enable direct comparisons ([Figure 3.S3](#)). The eastern and western tributaries are

marked in red and blue text, respectively. The minimum, mean, and maximum hillslope gradients in a 1,000 m-wide swath along the trunk channel are presented. 107

Figure 3.11. Trunk-channel sediment D_{84} versus swath hillslope gradient. Hillslope gradients were calculated as the mean gradient over a 300 m-long and 1,000 m-wide swath centred on the trunk-channel localities. 109

Figure 3.12. Schematic diagram illustrating process-based tectonic and lithological controls on exported sediment grain size (A, B) and their speculative implications for stratigraphic grain size evolution (C, D). In catchments where knickpoints have not yet developed, hard and fractured bedrock triggers pervasive mass wasting, and lithology-induced coarse hillslope sediment can directly coarsen trunk-channel sediment via coupled hillslopes and indirectly through tributary input (A); an upward-coarsening stratigraphy (IEFB onwards) is expected if continued incision unroofs the soft bedrock and exposes the hard and fractured bedrock (C). In catchments where tectonic knickpoints are developed within soft bedrock, mass wasting is concentrated downstream of the knickpoints, and tectonically induced coarsening, limited by lithology-dependent abrasion, does not extend into trunk-channel sediments (B); a characteristic stratigraphy with coarse sediment (OCD onwards) being preceded by fine sediment (PO-OCD) is expected due to the time required for knickpoints to propagate into hard bedrock (D). 114

Figure 3.S1. Field localities measuring rock strength (cyan squares), tributary sediments (yellow circles), hillslope sediments (red circles), and trunk sediments (white circles) in both the Phoenix (A) and Sithas (B), as well as adjacent Vouraikos (C) and Skoupeiko (D) catchments where identical bedrock is well-exposed. See Table 3.S1 and 3.S2 for raw data. 117

Figure 3.S2. Typical mass wasting events (red polygons) mapped from Google Earth: (A) mass wasting concentrated on *sand-silt*-dominated hillslopes along steepened channels in the Sithas, where tectonic knickpoints are present, (B) a zoomed-in view of the mass-wasting scars, (C) pervasive mass wasting on the *lime-* or *flysch*-dominated hillslopes in the Phoenix, and (D) a zoomed-in view of the mass-wasting scars and deposits. 118

Figure 3.S3. D_{84} -Wolman versus D_{84} -sieve. The solid and dashed red lines represent the best-fit lines, with and without considering the potential outlier (grey square). Note that the outlet localities from Watkins et al. (2020) for the two study rivers were also included (Table 3.S1). 118

Figure 3.S4. Ternary diagram of lithological proportions in tributary catchments. Note the colour and size of the symbols indicate the impact of tributary input on trunk sediments (Table 3.S1). 119

Figure 4.1. Two competing, hump-shaped relative sediment flux functions—dashed (Sklar and Dietrich, 2004) and solid (Turowski et al., 2007)—are fitted to short-term experimental data from an erosion mill, in which sediments of different lithologies are input to abrade an underlying bedrock disk (scatters on grey axes; Sklar and Dietrich, 2001). Here, Turowski et al.'s (2007) function fits the scatters better than Sklar and Dietrich's (2004), but its significance is limited by differences in timescale and hydrodynamics between the experimental mill and natural rivers. Identifying the correct function is crucial, as the two predict distinct river

dynamics (see main text). Note the Q_s/Q_t range can be conceptually divided into zones 1, 2, and 3, based on the gradient or sensitivity of the function to Q_s/Q_t variations. 130

Figure 4.2. Gulf of Corinth in maps of hillshaded relief (a), and lithology (b), with active faults (red lines) and the two study catchments highlighted. Abbreviations: p-con sed. – poorly consolidated sediment; cong. – conglomerate; sand-silt – sandstone-siltstone; lime. – limestone. AIG-Aigion fault, DER-Deveni fault, E. ELK-East Eliki fault, E. XYL-East Xylokastro fault, LYK-Lykoporia fault, N. KIA-North Kiato fault, PER-Perachora fault, PSA-Psathopyrgos fault, NER-Neo Erineos fault, W. ELK-West Eliki fault. 132

Figure 4.3. Main characteristics of the Phoenix (a, c, e) and Sithas (b, d, f) rivers: (a, b) trunk and tributary channels cutting across different lithologies and normal faults; (c, d) trunk-channel river profiles, coloured by lithology from [Figure 4.2](#), shown with first-order variations in uplift rates. Dashed red lines and grey bands represent the mean and range of fault uplift rate constraints, with their upstream decrease based on the linear back-tilting model derived from two well-constrained faults in Corinth ([see main text](#)); (e, f) low-flow stage field observations ([see locations in a and b](#)). Note both rivers exhibit the transition between the upstream zone, where sediment input comes directly from adjacent coupled hillslopes, and the downstream zone, where tributary input dominates. The upstream-downstream transition coincides with the boundary between hard and soft bedrock in the Phoenix river and with a knickpoint in the Sithas river. 133

Figure 4.4. Field photos illustrating bankfull flood markers (A), width (B), slope (C), and depth (D). 135

Figure 4.5. Bankfull width (grey) / wideness (red) (a, b), slope (grey) / steepness (red) (c, d), and width:depth, W/D (e, f) along the lithology-coloured trunk-channel river profiles for the Phoenix (a, c, e) and Sithas (b, d, f) rivers. Note the Phoenix river is dominated by width adjustments, whereas the Sithas river is dominated by slope adjustments. The onshore faults ([Figure 4.3a](#)) are marked as red lines, and the four transparent markers in the Phoenix river represent channels interpreted to be close to transport-limited alluvial conditions (see main text for details). See [Figure 4.3](#) for lithology legend. 138

Figure 4.6. Downstream variations in excess shear stress along the lithology-coloured trunk-channel river profiles for the Phoenix (a) and Sithas (b) rivers, with the locations of tributary confluences marked (dashed blue lines). Note the sensitivity of excess shear stress to tributary input differs significantly between the Phoenix and Sithas rivers. The four transparent markers in the Phoenix river represent channels interpreted to be close to transport-limited alluvial conditions (see main text for details). See [Figure 4.4](#) for lithology legend. 140

Figure 4.7. A schematic diagram illustrating the ‘tools and cover’ effect of sediment in shaping channel equilibrium geometry. Note, channels a and c exhibit similar channel geometry despite significant differences in relative sediment flux, because both have limited contrast in erosional efficacy between the channel bed and banks. Furthermore, this contrast does not change significantly when there is an increase in Q_s/Q_t , as indicated by dash arrows. In contrast, channel b, where the bank efficacy exceeds bed efficacy, exhibits a much wider

channel. The increasingly shallow gradient of the curve in the cover domain causes a Q_s/Q_t increase to reduce bank efficacy more than bed efficacy, as indicated by dash arrows. 143

Figure 5.1. Erosion thresholds and their impacts on threshold-exceeding floods. (A, B) Threshold discharges for the channel bed, Q_{bed} (A) and for banks, Q_{bank} (B). (C) Flood frequency-magnitude relation described by a stretched exponential distribution, whose shape parameter c characterises discharge variability. (D) Cross-sectional variations in shear stress cause Q_{bank} to be higher than Q_{bed} for a given critical discharge. Here, an increase in discharge variability (e.g. from $c = 1$ to $c = 0.5$) redistributes more of the total discharge into bigger events, allowing more frequent threshold-exceeding floods (e.g. the area below $c = 0.5$ is greater than that below $c = 1$ for Q_{bank}). Note, the area increase (termed erosion enhancement) caused by a given increase in discharge variability depends on the magnitude of threshold discharges, allowing channel width in actively incising rivers to be more sensitive to discharge variability than slope (Figure 5.S1). 154

Figure 5.2. Locations of selected rivers on (A) global mean annual precipitation map (Fick and Hijmans, 2017), (B) global aridity index map (Trabucco and Zomer, 2009). Aridity index is the ratio of mean annual precipitation to potential evapotranspiration. 156

Figure 5.3. Channel width versus drainage area, coloured by aridity index (AI). Each data point represents the mean width at a given location, with standard deviation shown as error bars. In total, ~2350 width measurements were made at 522 locations across 72 rivers. 157

Figure 5.4. k_{sn} and k_{wn} versus (A) erosion rate E , and k_{sn-E} and k_{wn} versus (B) mean annual precipitation MAP and (C) aridity index AI , where k_{sn-E} is k_{sn} normalised by the predicted values from the best-fit function in (A). 159

Figure 5.S1. (A) Two synthetic flood distributions with identical total discharge but distinct variability, parameterized by c , (B) total threshold-exceeding discharge for both distributions, and (C) their difference as a function of threshold discharge Q_c . Gray lines in A and boxes in B and C constrain the common range of threshold discharges for $c = 0.4$, based on return times reported by Lague (2014). For actively incising rivers ($Q_c < 10$), channel width is expected to be more sensitive to discharge variability than slope, because the bank threshold is higher than the bed threshold, as detailed in the main text (indicated as Q_{bed} and Q_{bank} in C). 163

Figure 5.S2. Illustration of channel width measurement in an example river (A), with five measurements taken at the location corresponding to 50% of the total drainage area (B). ...165

Figure 5.S3. Channel width versus drainage area, coloured by erosion rate (A) and mean annual precipitation (B). 166

List of tables

Table 2.1. Data associated with the knickpoints identified in the 16 trunk rivers.	54
Table 2.S1. Compiled uplift rates in the south coast of the Gulf of Corinth.	71
Table 3.1. Basic characteristics of study catchments.	94
Table 3.S1. Schmidt hammer rock strength.	120
Table 3.S2. Sediment grain size.	124
Table 4.S1. Bankfull channel geometry, discharge, excess shear stress, and transport capacity.	149
Table 4.S2. Tributary bedload-lithology proportion and drainage area relative to the trunk channel.	151
Table 5.S1. Tectonic, climatic, lithological, and topographic characteristics of the selected rivers.	168
Table 5.S2. P-value and power-law exponents of the relations between channel geometry and tectonic and climatic parameters.	171

1 INTRODUCTION

1.1 Why are fluvial landscapes important?

Much of Earth's surface is shaped by rivers that carve into bedrock and redistribute water, sediment, and nutrients (e.g. Allen, 2008; Best, 2019). In this thesis, upland rivers are defined to include both bedrock rivers *stricto sensu* (Whipple, 2004) and flux-dependent rivers (Whipple and Tucker, 2002). These rivers are responsible for generating approximately 80–90% of topographic relief (e.g. Whipple, 2004), setting the base level for adjacent hillslopes (e.g. Gallen et al., 2011), and releasing sediment cascades into source-to-sink systems (e.g. Allen, 2008). In doing so, their channel slope and width approach an equilibrium with tectonic, climatic, and lithological forcings (e.g. Wobus et al., 2006; Stark et al., 2006). The link between channel geometry and the forcings, if quantified, can be used to assess seismic and erosion hazards (e.g. Kirby et al., 2008; Barneveld et al., 2025) and model landscape dynamics (e.g. Yanites, 2018; Whipple et al., 2022). Nonetheless, quantifying the impacts of tectonics, lithology, and climate on channel geometry in upland rivers is non-trivial.

In the mid-twentieth century, channel slope was first used routinely to infer river erosive power for a given drainage area (Hack, 1957; Flint, 1974). This approach was further formalised by Howard and Kerby (1983), who quantified the relationship between channel slope, drainage area, and erosion rate in 'badland' settings. In recent decades, area-normalised slope or steepness has been shown to correlate well with uplift rates in steady-state catchments (e.g. Portenga and Bierman, 2011). This has supported its use as a proxy for fault uplift rates that can be used to refine tectonic models and assess seismic hazards (e.g. Kirby and Whipple, 2012). For example, the seismic hazard associated with the devastating Mw 7.9 earthquake that struck Sichuan, China in 2008 – unresolved in short-term geodetic data – is nevertheless apparent in channel steepness patterns (Kirby et al., 2008). In most existing analyses, however, the effect of lithology on channel steepness is not explicitly accounted for (e.g. Miller et al., 2013; Gallen and Wegmann, 2017). This ad hoc simplification of uniform rock type is concerning given that failure to incorporate lithological effects may lead to misinterpretations (e.g. Gailleton et al., 2021). This concern could be alleviated if competing

tectonic and lithological effects on channel steepness were properly isolated in geomorphic studies.

Mechanically, river incision is driven mostly by impacts of moving bedload sediment (e.g. [Turowski et al., 2023](#)). This mechanism allows lithology to control channel geometry via its influence on sediment generation (e.g. [Thaler and Covington, 2016](#); [DiBiase et al., 2018](#)), as sediment supply governs both the shear stress required for erosional equilibrium (e.g. [Cowie et al., 2008](#)) and its partitioning between channel slope and width adjustments (e.g. [Yanites and Tucker, 2010](#)). It follows that lithology-influenced sediment supply may complicate efforts to infer tectonic activity from channel slope (e.g. [Whittaker et al., 2007](#)) and model landscape evolution (e.g. [Attal et al., 2008](#)). Thus, insights into sediment-mediated lithological control on channel slope and width are much needed and can only be gained where bedload sediment supply is well-constrained.

In essence, channel slope and width are shaped by the time-integrated effects of bed and bank erosion during threshold-exceeding floods. The threshold mechanism allows increases in discharge variability to result in enhanced erosion (e.g. [Molnar, 2001](#); [DiBiase and Whipple, 2011](#)). It has direct societal relevance, because intense erosion exacerbates flood hazards (e.g. [Thapa et al., 2024](#)) and jeopardises infrastructure by bank failure and riverbed degradation (e.g. [Barneveld et al., 2025](#)). These risks become even more urgent considering that ongoing climate change may increase discharge variability ([IPCC, 2023](#); [McLeod et al., 2024](#)). However, the nature, direction, and magnitude of cumulative channel erosion over the coming decades remain poorly understood (e.g. [Slater and Singer, 2013](#)). This can be aided by understanding how modern river slope and width respond to discharge variability.

This thesis addresses tectonic, lithological and climatic controls on fluvial landscape evolution by combining DEM, field, remote sensing analyses. I first focus on rivers draining into the Gulf of Corinth, Greece as a natural laboratory to unravel the relative importance of tectonics and lithology in shaping long-term topography (Chapter 2), and to examine the sediment-mediated lithological control on channel geometry (Chapters 3 and 4). I then use carefully selected rivers across the world to investigate the response of river erosion to discharge

variability (Chapter 5). Together, these studies provide new insights into how channel geometry can be leveraged to interpret landscape dynamics and geohazards.

1.2 Research background

To address this important topic, I review the theoretical and empirical foundations that frame the research and motivate the approaches adopted in this thesis. This section begins with the formulation of stream-power incision laws or models, which form the basis for subsequent analyses. It is followed by three subsections synthesising key developments in tectonic, lithological, and climatic controls on upland river dynamics. Additional background, including study-area descriptions and methods, is provided in the subsequent research chapters respectively.

1.2.1 Stream-power incision models

To date, the ‘default’ framework for modelling upland river dynamics is the stream-power incision model (SPIM) (Whipple et al., 2013). Its classic formulation originated in ‘badland’ settings in the 1980s, when Howard and Kerby (1983) expressed erosion rates as a function of drainage area and channel slope. In these highly erodible settings, erosion thresholds are readily exceeded during most rainfall events and sediment supply (Q_s) is much lower than transport capacity (Q_t). However, such conditions are not representative of all geomorphic settings. The effects of moving sediment, incision thresholds, and stochastic flood variability have therefore been incorporated into subsequent SPIM formulations (e.g. Tucker and Bras, 2000; Whipple and Tucker, 2002). In their most general form, SPIMs can be expressed in terms of shear stress, such that erosion rate (E) scales as a power law with bed shear stress (τ_b) and is modulated by bedrock erodibility (k_e), sediment effects ($f(Q_s/Q_t)$), and an erosion threshold (τ_c) (Whipple, 2004).

$$E = k_e f(Q_s/Q_t) (\tau_b^a - \tau_c^a) \quad (1.1)$$

where Q_s/Q_t is relative sediment flux, and the exponent a depends on the mechanics of erosion. In this case, bed shear stress can be approximated as:

$$\tau_b = k_t (Q/W)^\alpha S^\beta \quad (1.2)$$

where k_t , α , and β are constants set by flow-resistance relations, Q is water discharge, and W is channel width, S is channel slope (Whipple, 2004). Equations 1.1 and 1.2, when combined with empirical relations among drainage area (A), discharge, and channel width, yield a generalised SPIM formulation that incorporates both sediment effects and threshold–stochastic dynamics (see Whipple, 2004 for details). In practice, however, this general formulation is often simplified by necessity. Below, I outline three SPIM simplified variants tailored to specific geomorphic settings or research objectives.

Firstly, in landscape-scale analyses, long-term sediment supply and stochastic discharge distributions are difficult to constrain. As a result, most studies treat $f(Q_s/Q_t)$ as constant (e.g. Whipple and Tucker, 1999), and approximate a stochastic discharge distribution by a constant effective discharge (i.e. bankfull discharge; e.g. Lague et al., 2005) that markedly exceeds critical shear stress (e.g. Howard and Kerby, 1983). In addition, channel width is often assumed to scale solely with water discharge (e.g. Whipple and Tucker, 1999). The three simplifications above lead to the classic SPIM, in which erosion rate is given as a power-law function of drainage area and channel slope:

$$E = KA^mS^n \quad (1.3)$$

where K is an erodibility coefficient, and m and n are positive constants (Whipple and Tucker, 1999). Here, the erodibility coefficient depends on rock resistance to erosion, roughness, channel width, and precipitation, amongst other controls (e.g. Gallen, 2018). This model is used in Chapter 2 (see section 1.2.2).

Secondly, to constrain the role of sediment supply, the bankfull-discharge and negligible-critical-shear-stress simplifications are often taken to highlight the term $f(Q_s/Q_t)$. Because sediment supply can influence channel width (e.g. Yanites and Tucker, 2010), width should be explicitly incorporated. Together, these considerations yield the flux-dependent variant of SPIM:

$$E = k_e f(Q_s/Q_t) [k_t (Q/W)^\alpha S^\beta]^\alpha \quad (1.4a)$$

$$\text{or } E = K' f(Q_s/Q_t) (Q/W)^{m'} S^{n'} \quad (1.4b)$$

where K' is $k_e k_t^a$, m' is αa , and n' is βa . The functional form of $f(Q_s/Q_t)$, being fundamental to flux-dependent dynamics, has been explored by theoretical (e.g. Turowski et al., 2007), experimental (e.g. Sklar and Dietrich, 2001; Johnson and Whipple, 2010; Chatanantavet and Parker, 2008), and field (e.g. Cowie et al., 2008; Hobley et al., 2011; Johnson et al., 2009) studies. In general, these studies favour a dual role of sediment in influencing river incision (cf. Beaumont et al., 1992): incision is enhanced at a moderate sediment supply but is inhibited when sediment supply becomes excessive, a behaviour referred to as the “tools and cover” effect. Two formulations of $f(Q_s/Q_t)$ incorporate this behaviour but differ in their sensitivity to Q_s/Q_t (see Sklar and Dietrich, 2004; Turowski et al., 2007); however, field evidence is yet to distinguish between them. This model motivates the work in Chapters 3 and 4 (see section 1.2.3).

Thirdly, in efforts to quantify the role of discharge variability, $f(Q_s/Q_t)$ is often assumed constant to emphasise the erosional threshold and the stochastic distribution of floods (e.g. Tucker and Bras, 2000). Here, channel width is often prescribed as a function of discharge but can vary with instantaneous flood discharge based on at-a-station hydraulic geometry (e.g. Lague et al., 2005). In the stochastic-threshold variant of SPIM, long-term erosion is computed by integrating the instantaneous incision rate over the discharge distribution for flows exceeding the erosion threshold (e.g. Lague, 2005). For simplicity, I do not present the derivation of the model but highlight its underlying principles here. In the stochastic-threshold model, the magnitude of erosion thresholds and their impact on regulating the proportion of threshold-exceeding floods are central to the resulting erosion dynamics. This is well-illustrated by how thresholds and stochastic floods can combine to reproduce a non-linear relationship between channel steepness and erosion rate (e.g. Lague et al., 2005; DiBiase and Whipple, 2011; Snyder et al., 2003). This behaviour arises because when channel slope is low and thresholds are high, only extreme events exceed thresholds and contribute to erosion; increases in channel slope increase the stream power of individual floods, as in other SPIM variant, but also lowers the erosion threshold and therefore increasing the proportion of erosive floods. As a result, erosion rates become more sensitive to increases in channel slope

at higher slopes than at lower slopes. The principles of this model are used in Chapter 5 (see section 1.2.4).

An alternative framework for understanding upland river dynamics is provided by mechanical erosion models (Venditti et al., 2020). In contrast to SPIMs, mechanical models are built upon specific erosional processes such as abrasion and plucking, allowing the “tools and cover” effect to be explicitly incorporated (e.g. Sklar and Dietrich, 2004; Chatanantavet and Parker, 2009). These models have been tested and validated at reach or smaller scales (e.g. Cook et al., 2013; Li et al., 2023), but upscaling them to landscape scales remain challenging due to difficulties in constraining Q_s , identifying dominant erosion processes across heterogeneous landscapes, and meeting their high computational demands (e.g. Venditti et al., 2020). In this thesis, I aim to investigate long-term, landscape-scale upland river dynamics; consequently, I use SPIMs as a practical framework for my analysis.

1.2.2 Lithological versus tectonic controls on topography

As I outlined at the start of this thesis, channel steepness correlates well with erosion rate at local scales (e.g. Kirby and Whipple, 2012). In global compilations, however, channel steepness can differ by more than an order of magnitude for the same erosion rate (e.g. Portenga and Bierman, 2011; Kirby and Whipple, 2012; Lague, 2014). A possible explanation is variation in bedrock lithology, via its influence on rock resistance to erosion (e.g. Sklar and Dietrich, 2001; Gallen, 2018). Indeed, lithological control on topography via erodibility has been well-demonstrated, especially in post-orogenic areas (e.g. Clementucci et al., 2022; Dethier et al., 2014; Scharf et al., 2013; Bernard et al., 2019; Haag et al., 2025). This then raises the question: what is the relative importance of tectonics and lithology in shaping topography in tectonically active settings? This question is important not only to understanding landscape evolution (e.g. Yanites et al., 2017) but also to assessing seismic hazards (e.g. Kirby, 2008) in tectonically active and lithologically heterogeneous landscapes.

This has motivated many studies to estimate lithological erodibility, but this has proven challenging in tectonically active landscapes (e.g. Gailleton et al., 2021). For example, deriving lithological erodibility from rock strength (Sklar and Dietrich, 2001) can be difficult, as

measurements at centimetre to metre scales may not adequately capture heterogeneity and fracturing (e.g. [Gallen et al., 2015](#); [Turowski et al., 2023](#)). Landscape-scale lithological erodibility can be calibrated using numerical modelling (e.g. [Stock and Montgomery, 1999](#); [Yanites et al., 2017](#)) or field measurements (e.g. [Kent et al., 2021](#); [Zondervan et al., 2020a](#)), but both methods require robust and extensive constraints on erosion rates that are often unavailable at a landscape scale. In cases where erosion rates are relatively uniform, such as post-orogenic areas, channel steepness has been used to estimate lithological erodibility (e.g. [Gallen, 2018](#); [Clementucci et al., 2022](#); [Zondervan et al., 2020b](#)). This simple approach not only yields landscape-scale erodibility estimates but also relies only on readily available datasets. It could therefore be a useful tool if proven effective in tectonically active settings.

In studies that have derived lithological erodibility, both its magnitude and range differ markedly. For example, two- to five-fold differences in lithological erodibility have been reported in some studies (e.g. [Zondervan et al., 2020a](#); [Yanites et al., 2017](#); [Gallen, 2018](#)), in stark contrast to the orders-of-magnitude contrasts inferred in other work ([Sklar and Dietrich, 2001](#); [Stock and Montgomery, 1999](#)). In some studies, lithological erodibility is even suggested to exhibit negligible variation (e.g. [Miller et al., 2012](#); [Racano et al., 2023](#)). This inconsistency between different studies highlights the need to test the effectiveness of estimated lithological erodibility. In tectonically active settings, lithology-calibrated topographic metrics are expected to record tectonic processes, if the estimated lithological erodibility is effective. In other words, the effectiveness of estimated erodibility can be evaluated by comparing lithology-calibrated metrics with independent constraints on tectonic processes; however, such comparisons remain rare, in part because areas with independent tectonic constraints are limited. The validated erodibility also provides a benchmark against which erodibility values obtained from different approaches can be evaluated.

1.2.3 Flux-dependent lithological control on channel geometry

Lithology may control channel geometry via its influence on bedload sediment supply (e.g. [Thaler and Covington, 2016](#); [DiBiase et al., 2018](#); [Carr et al., 2023](#)). This is well illustrated by [Thaler and Covington \(2016\)](#), who showed that channel steepness does not correlate with local bedrock lithology, but with boulder size and the extent of boulder cover within channels,

both of which are set by hard rock upstream. Likewise, [Duvall et al. \(2004\)](#) and [Johnson et al. \(2009\)](#) found that channels crossing hard bedrock before entering softer units exhibit higher steepness than those directly incising soft bedrock. In particular, they suggested that the elevated channel steepness is driven by the need to transport the bedload sediment generated by the hard bedrock. In contrast, [DiBiase et al. \(2018\)](#) and [Carr et al. \(2023\)](#) argued that the grain size of sediment generated by bedrock dominates channel steepness. In general, these field studies demonstrate that lithology can exert nonlocal effects on channel steepness, either via sediment flux or grain size; however, few direct constraints on bedload sediment supply and channel width limit further progress in this regard.

In tectonically active and lithologically heterogeneous areas, both tectonics and lithology can control bedload sediment generation (e.g. [Sklar et al., 2017](#)). For example, lithology influences sediment production through rock-strength-dependent fragmentation and abrasion (e.g. [Attal and Lavé, 2009](#); [Allen et al., 2015](#); [Marc et al., 2021](#); [Dingle et al., 2017](#)). In contrast, uplift rate increases can enhance coarse sediment production by promoting landslides and reducing weathering time (e.g. [Whittaker et al., 2010](#); [Attal et al., 2015](#)). Nevertheless, how tectonics and lithology interact to control sediment generation remains poorly understood. This limits progress in characterising non-local lithological effects on channel geometry. It is also crucial from a stratigraphic perspective, given that sediment generation determines the material available for deposition. Indeed, bedrock unroofing has been invoked to explain stratigraphic changes (e.g. [Carroll et al., 2006](#); [McArther et al., 2013](#)), but this mechanism cannot be explicitly tested without directly characterising lithological influence on sediment generation in tectonically active settings (e.g. [Attal and Lavé, 2006](#)).

Where sediment generation or supply can be characterised, the non-local effect of lithology on channel width may be clarified. Numerical models suggest that lithology-influenced sediment supply can control the relative importance of channel slope and width adjustments in accommodating external forcing, such as tectonic uplift (e.g. [Turowski et al., 2008](#); [Yanites, 2018](#)). This appears consistent with the complex channel width response observed in field (e.g. [Whittaker et al., 2007](#); [Amos and Burbank, 2007](#); [Yanites et al., 2010](#)) and sandbox experiments (e.g. [Turowski et al., 2006](#)), as argued by [Yanites and Tucker \(2010\)](#). This

modelling result, however, remains untested because direct constraints on sediment supply are often lacking, and two competing $f(Q_s/Q_t)$ formulations that are central to flux-dependent dynamics are yet to be resolved (Sklar and Dietrich, 2004; Turowski et al., 2007). This thus underscores the need to examine the role of sediment supply in modulating upland river incision.

1.2.4 Threshold-mediated climate control on channel geometry

Unlike tectonics, climate control on topography has proven challenging to substantiate. For example, observed relationships between erosion rates and mean annual precipitation (*MAP*) are complex, ranging from minimal correlation (e.g. Riebe et al., 2001; Burbank et al., 2003; Godard et al., 2014) to sub-linear (e.g. Bookhagen and Strecker, 2012; Ferrier et al., 2013; D’Arcy and Whittaker, 2014) and near-linear (e.g. Adams et al., 2020; Leonard et al., 2023) scaling. This is partly because stochastic floods comprise events with varying magnitudes and frequencies, and only those exceeding a given threshold contribute to erosion, such that the role of climate is mediated by these thresholds (e.g. Perron, 2017). Erosion rates can be decoupled from total discharge, a metric that *MAP* attempts to constrain (e.g. D’Arcy and Whittaker, 2014), when erosion thresholds are sufficiently high that extreme floods, rather than average floods, dominate erosion (e.g. Lauge et al., 2005; DiBiase and Whipple, 2011). This underscores that the partitioning of total discharge among small, moderate, and extreme floods (i.e., discharge variability) may be as important as total discharge in controlling river erosion, depending on erosion thresholds (e.g. Molnar, 2001; Molnar et al, 2006). Quantifying the impact of discharge variability on erosion becomes increasingly important for human societies, considering that ongoing climate change is likely to increase discharge variability in many rivers worldwide.

The impact of discharge variability on erosion has been illustrated in some well-constrained areas. For example, Carretier et al. (2013) attributed differences between decadal and millennial erosion rates to the contribution of extreme floods with recurrence intervals longer than decadal but shorter than millennial timescales. Their results show that the contribution of such rare extreme floods to millennial erosion increases from 0% in humid zones to over 90% in arid zones. For catchments on Reunion Island in Indian Ocean, Gayer et al. (2019) found

that, although they experience the same extreme rainfall events, erosion rates still differ by an order of magnitude and are higher when prevailing climates are drier. Both field studies support the hypothesised role of discharge variability in controlling erosion, but their scope is limited to a local scale. A global-scale analysis was attempted by [Marder and Gallen \(2023\)](#), but their results are confounded by the trade-off between discharge variability and total discharge (e.g. [Molnar et al, 2006](#)). In general, field evidence for discharge variability control on erosion remains limited at a global scale.

Moreover, discharge variability control on erosion has rarely been examined with respect to channel width (cf. [Lague et al., 2005](#); [DiBiase and Whipple, 2011](#)). [Stark et al. \(2010\)](#) found that bedrock river sinuosity, driven by channel bank erosion, increases with the frequency of extreme rainfall (> 50 mm/day). Likewise, [Leenman et al. \(2025\)](#) found that high discharge variability accelerates alluvial river mobility (e.g. channel migration and avulsions). Both studies suggest that high discharge variability causes enhanced bank erosion, as evidenced by increased channel mobility and sinuosity, and therefore may lead to channel widening. This hypothesis was explored, to my knowledge, only by [Phillips et al. \(2024\)](#), who reported that gauged alluvial rivers across the continental US are wider, when discharge variability increases. Hence, whether this pattern extends to bedrock rivers or beyond the US remains elusive.

1.2.5 Key research gaps

To sum up, erosion in upland river is a theoretically complex process, and the roles of erosion thresholds, stochastic floods, and relative sediment flux are often poorly constrained. In attempts to probe and leverage these controls, the cause-effect relationships between external forcing and channel geometry have been examined. Despite marked progress, three key gaps remained: (i) tectonics and lithology often co-shape landscapes, yet few studies have attempted to unravel them, (ii) sediment supply has been hypothesised to modulate channel responses to external forcing, but the mechanisms remain elusive, and (iii) the effects of erosion thresholds and stochastic floods on channel geometry remain centred on channel slope and are poorly calibrated against field data.

1.3 Aims and objectives

Having reviewed the status of current research addressing tectonic, lithological, and climatic controls on fluvial-dominated landscapes, I have developed four key aims that I will address in this thesis to bridge some of the existing knowledge gaps. These aims are:

- (i) isolating tectonic versus lithological effects on transient landscapes (Chapter 2).
- (ii) characterising tectonic versus lithological effects on sediment generation (Chapter 3).
- (iii) constraining the role of bedload sediment supply in shaping channel geometry (Chapter 4).
- (iv) determining discharge variability control on channel geometry (Chapter 5).

Aim 1 examines tectonic and lithological effects on topographic evolution. This is important because lithological controls on topographic metrics, if not properly accounted for, may obscure or be misinterpreted as tectonic signals. The extent to which the relative influences of tectonic and lithological controls can be disentangled within the framework of the classic SPIM also helps evaluate its applicability in a tectonically and lithologically complex setting. To this end, I examine 16 upland rivers draining into the Gulf of Corinth, Greece, where tectonics, lithology, and climate are well constrained (e.g. [Gawthorpe et al., 2018](#); [Watkins et al., 2019](#)). The specific objectives related to this aim are to:

- (i-a) constrain lithological erodibility and derive lithology-calibrated topographic metrics intended to isolate tectonic signals from lithological effects;
- (i-b) compare these lithology-calibrated metrics with independent tectonic constraints to evaluate the extent to which lithological and tectonic effects are disentangled;
- (i-c) apply the lithology-calibrated metrics, if shown to be effective, to refine current tectonic models.

Aim 2 then turns to the erosional products of topographic evolution (i.e. sediment). This analysis bridges the gap that the hypothesised role of upstream rivers in influencing stratigraphic architecture via sediment generation has not been well-assessed. It also provides

essential constraints on bedload supply for subsequent analyses. I therefore focus on two Corinth rivers in detail to assess the competing influence of tectonics and lithology in sediment generation. The specific objectives are to:

(ii-a) characterise tectonic and lithological effects on hillslope and fluvial processes that both generate and transport sediment;

(ii-b) track downstream variations in sediment grain size from hillslopes to tributaries and ultimately to catchment outlets, to assess the competing influences of tectonics and lithology;

(ii-c) evaluate the potential impact of upland sediment generation on stratigraphy.

Building upon these results, **Aim 3** further analyses how sediment supply modulates the morphological responses of the two Corinth rivers to tectonics and lithology. This work seeks to clarify the conditions under which the classic SPIM can be reasonably used to characterise upland river incision. This issue is particularly important because, (1) given the challenges in constraining Q_s , the classic SPIM has been, and will likely continue to be, a “default” model for interpreting upland river incision, and (2) neglecting flux-dependent dynamics may result in misleading interpretations. To meet this aim, the objectives are to:

(iii-a) characterise channel slope and width responses to known gradients in tectonics and lithology;

(iii-b) compare the sensitivity of excess shear stress to sediment supply variations to distinguish between two competing $f(Q_s/Q_t)$ formulations;

(iii-c) clarify how flux-dependent dynamics may influence the relative importance of channel slope and width adjustments in accommodating variations in tectonics and lithology.

In contrast, **Aim 4** explores discharge variability control on channel geometry at a global scale. This research is motivated by the fact that, despite the fundamental role of channel width in regulating channel slope and river dynamics, its response to discharge variability remains rarely constrained. This is of increasing social importance, given that global warming is expected to increase discharge variability in many rivers in the near future. I focus on well-

constrained rivers across the globe to evaluate the potential impact of discharge variability on erosion (cf. [Marder and Gallen, 2023](#)). The specific objectives are to:

(iv-a) measure channel slope and width and characterise their responses to discharge variability proxy.

(iv-b) constrain the potential impact of climate change on channel bank and bed erosion via influencing discharge variability.

In this thesis, I address aims 1-3 in the Gulf of Corinth, Greece. I use the Corinth rift as a natural laboratory because the area represents a tectonically active, lithologically heterogeneous landscape typical of many regions at Earth's surface, and onshore terraces and offshore high-resolution seismic stratigraphy provide independent tectonic constraints (e.g. [Bell et al., 2008](#); [de Gelder et al., 2019](#); [Fernández-Blanco et al., 2020](#)). Full details of the study area are provided in each research chapter, below, as needed. For aim 4, I selected rivers across the globe based on specific criteria, which are described in Chapter 5.

1.4 Thesis outline

This thesis comprises an introduction, four research chapters, a discussion, and a conclusion. Each of the research chapters is written and structured as a stand-alone research study, but together they form an integrated thesis that addresses the aims and objectives outlined above. As a whole, this thesis provides new insights into tectonic, lithological, and climatic controls on upland river dynamics.

In Chapter 2, I use the Gulf of Corinth as a natural laboratory to isolate tectonic and lithological effects on topographic evolution (**aim 1**). I first estimate lithological erodibility and derive lithology-calibrated topographic metrics. I then compare these lithology-calibrated metrics with independent tectonic constraints to evaluate the effectiveness of the erodibility estimates. Based on this evaluation, I exploit the calibrated metrics to refine the tectonic model. This chapter sets the context for the following analyses. **A version of this chapter has been published in *Basin Research* as Zhou Z., Whittaker, A.C., Bell, R.E., Hampson, G.J. (2024). "Unravelling tectonic and lithological effects on transient landscapes in the Gulf of Corinth, Greece." doi: [10.1111/bre.12901](https://doi.org/10.1111/bre.12901)**

Based on this context, I further investigate how the two drivers—tectonics and lithology—compete to control sediment liberation, as sediments are the direct erosional product of topographic evolution (**aim 2**), in Chapter 3. Focusing on two Corinth rivers, I use a process-based approach to first characterise hillslope and channel processes directly responsible for sediment production and transport, and then to track downstream variations in sediment grain size—from hillslopes to tributaries and ultimately to catchment outlets. The results show that lithology dominates over tectonics in setting sediment grain size and only the hard bedrock contributes to bedload sediment generation in our study areas. **A version of this chapter has been published in *Geological Society of America Bulletin* as Zhou Z., Whittaker, A.C., Hampson, G.J., Bell, R.E., Chandresh, R., Watkins, S.E., Zondervan, J.R. (2026). “Decoding tectonic and lithological controls on sediment exported from footwall catchments: Implications for hangingwall stratigraphy.” doi: [10.1130/B38399.1](https://doi.org/10.1130/B38399.1)**

Building upon these constraints on tectonics, lithology, and sediment supply, I subsequently analyse the role of bedload sediment supply in modulating channel geometry (**aim 3**) in Chapter 4. For the two Corinth rivers that traverse known gradients in tectonics and lithology with distinct sediment supply, I characterise their channel slope and width responses. I then compare their excess shear stress sensitivities to sediment supply variations to distinguish between two competing $f(Q_s/Q_t)$ formulations. This analysis enables a conceptual model that clarifies the relative role of channel slope and width adjustments in accommodating variations in tectonics and lithology. The results, together with Chapter 3, clarify how lithology can control channel geometry via its influence on sediment supply. **A version of this chapter has been published in *Earth and Planetary Science Letters* as Zhou Z., Chandresh, R., Whittaker, A.C., Hampson, G.J., Bell, R.E. (2026): “Sediment supply controls on channel morphological adjustments to tectonics and lithology.” doi: [10.1016/j.epsl.2026.120021](https://doi.org/10.1016/j.epsl.2026.120021)**

In Chapter 5, I explore discharge variability control on channel geometry at a global scale (**aim 4**). I focus on steady-state, metamorphic-rock-dominated rivers with conditions typical of most rivers worldwide, and examine their channel slope and width responses to mean annual precipitation and aridity index, both of which are effective proxies for discharge variability. The results show that channel width, but not slope, correlates with these discharge variability

proxies. Based on these results, I provide unique insights into the impact of discharge variability on landscape dynamics. **A version of this chapter is to be submitted to *Geology* as Zhou et al., Whittaker, A.C., Hampson, G.J., Bell, R.E. (2026): “Discharge variability controls channel width but not slope.”**

Having explored tectonic, lithological, and climatic controls on channel geometry, I return to a wider perspective in Chapter 6 to summarise the key findings of all four research chapters and to discuss future directions. I first synthesise these findings in relation to the aims and objectives of this thesis. I then discuss some important cross-cutting themes that emerge across the chapters. I finished by outlining prospective directions for future research that are directly motivated by the results presented here.

I concluded this thesis in Chapter 7 by providing a short and clear summary of the main findings.

In all four research chapters, the pronoun “we” is used rather than “I” to acknowledge the contributions of co-authors. However, I confirm that I conceived the research, carried out the data analysis, produced the figures and tables, and wrote the manuscripts. The contributions of co-authors Dr Alex Whittaker (Chapters 2–5) and Dr Rebecca Bell (Chapters 2-5) included PhD supervision, field assistance, as well as review and editing of manuscripts (i.e., the research chapters). The contributions of co-author Prof Gary Hampson (Chapters 2-5) included PhD supervision and manuscript review and editing. The contributions of co-authors Reha Chandresh (Chapters 3-4), Dr Stephen Watkins (Chapter 3) and Dr Jesse Zondervan (Chapter 3) included assistance with fieldwork and data provision.

2 UNRAVELLING TECTONIC AND LITHOLOGICAL EFFECTS ON TRANSIENT LANDSCAPES IN THE GULF OF CORINTH, GREECE ¹

Abstract

Landscapes are the integrated product of external forcings such as tectonics, climate, and bedrock. In principle, hard bedrock with low erodibility can steepen rivers in a similar way to tectonic uplift. A key challenge in geomorphic analysis is thus separating the tectonic and lithological effects on landscapes. To address this, we focus on multiple rivers that are transiently incising through contrasting lithologies in the Gulf of Corinth, Greece, where tectonic history is broadly well-constrained. We first exploit topographic metrics and river long profiles to demonstrate that landscapes are responding to both tectonics and lithology. In particular, the long profiles are divided into knickpoint-bounded segments, and at this scale, channel steepness is shown to be more sensitive to lithology than the entire catchment, possibly due to relatively uniform erosion rate at the segment scale. We then use segment-scale steepness variations between different lithologies to constrain their relative erodibilities ($K_{lme} : K_{cong.} : K_{sand-silt} : K_{p-con sed.} = 1 : 2 : 3 : 4$), which are further converted into actual lithology-dependent erodibilities by modelling a well-constrained, ca. 700 ka knickpoint in the Vouraikos catchment. The effectiveness of lithology-dependent erodibilities is supported by the observation that river long profiles, only when calibrated by lithological erodibility, obtain optimal concavities that fall within the theoretical range. Finally, we use lithology-calibrated metrics to provide new geomorphic constraints on the timing and magnitude of tectonic perturbations in these catchments. These geomorphic results are interpreted in conjunction with previous studies to shed new light on fault growth and linkage history in the Gulf of Corinth. Our study therefore provides a topographic analysis-based approach to quantify lithological effects on transient catchments, with important implications for other lithologically heterogeneous landscapes.

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2.1 Introduction

Landscape evolution is characterized by the competing influences of external forcings (e.g., tectonics and climate) and intrinsic characteristics (e.g., bedrock erodibility) (Jansen et al., 2010; Yanites et al., 2017; Kent et al., 2021). It has been widely demonstrated that topographic metrics, such as channel steepness and hillslope angle, can be used to infer uplift rate (e.g., Ouimet et al., 2009; DiBiase et al., 2010; Miller et al., 2012), decipher precipitation variations (e.g., Ferrier et al., 2013; D'Arcy and Whittaker, 2014; Adam et al., 2020; Marder and Gallen, 2023) or to constrain bedrock erodibility (e.g., Jansen et al., 2010; Bursztyn et al., 2015; Gallen, 2018) in some carefully chosen study areas. Where boundary conditions are less well-constrained, however, the fact that topography records an integrated response to multiple forcings inhibits our ability to decode useful information from landscapes (Whittaker, 2012; Kirby and Whipple, 2012). For example, higher channel steepness can be interpreted to reflect higher uplift rates, lower precipitation rates, less erodible bedrock, or a combination of these three factors (e.g., Gailleton et al., 2021b; Takahashi et al., 2022; Leonard et al., 2023). Although such ambiguities have long been acknowledged, unravelling these processes from landscapes remains an outstanding challenge (e.g., Schoenhohm et al., 2004; Wobus et al., 2006; Yanites et al., 2017; Kent et al., 2021).

A common and relatively simple framework to unravel tectonic, climatic, and lithological controls is the stream power law, in which river incision rate, E is modelled as a function of bedrock erodibility, K , upstream drainage area, A and local slope, S :

$$E = KA^mS^n, \quad (2.1a)$$

$$K = K_pP^m, \quad (2.1b)$$

where m and n are positive constant exponents (Whipple and Tucker, 1999; Leonard et al., 2023). If erosion rates balance uplift rates (i.e., steady state), then the effects of tectonics are translated directly into parameter E . Additionally, the impact of climate is often described by changes in precipitation rate, P (Perron, 2017; Adams et al., 2020). Recent studies have demonstrated that bedrock lithology (among other factors) dominates the spatial variability of the erodibility value (K_p) (Bursztyn et al., 2015; Zondervan et al., 2020b; Kent et al., 2020;

[Clementucci et al., 2022](#)). Thus, the effect of spatially variable lithology on topography can be calibrated if erodibility variations between different lithologies (i.e., relative erodibilities) are estimated (e.g., [Gallen, 2018](#); [Campforts et al., 2020](#); [Gailleton et al., 2021b](#)). In particular, lithology-calibrated metrics may yield key constraints on landscape evolution and its drivers, as well-illustrated in post-orogenic areas ([Bernald et al., 2019](#); [Gailleton et al., 2021](#)). For example, [Gallen \(2018\)](#) has used lithology-dependent erodibilities to constrain the timing of base level fall in the post-orogenic, southern Appalachian Mountains.

In tectonically active areas, however, the impact of lithology in modulating topography is less often and less well incorporated in geomorphic analysis. In many instances, lithological variations in erodibility were neglected to extract the presumably dominant tectonic signal (e.g., [Miller et al., 2012](#); [Gallen and Wegmann, 2017](#)). This simplifying assumption arises mostly from a weak correlation between lithology and topographic metrics averaged over a region, where complexities such as variations in erosion rate exist (e.g., [Racano et al., 2023](#)). In contrast, field studies focusing on segment-scale topography often demonstrate a strong lithological control (e.g., [Goode and Wohl, 2010](#); [Zondervan et al., 2020b](#); [Kent et al., 2021](#)). These studies thus imply that the impact of lithology on topography may be underestimated in some broad-scale analysis, given the mixed nature of these averaged metrics. This has motivated a few studies using segment-scale steepness variations to estimate lithological variations in erodibility and thereafter calibrate topographic metrics; yet their estimations and calibrations often lack assessment by independent constraints (e.g., [Shi and Shao, 2014](#); [Takahashi et al., 2022](#)). In short, the extent to which lithology modulates topography and the extent to which such modulation can be incorporated explicitly to refine the detection of tectonic signals remain elusive in areas of active faulting.

To bridge this gap, we focus on the Gulf of Corinth, where tectonic history is broadly well-constrained and where contrasting lithologies are exposed (e.g., [Bell et al., 2009](#); [Nixon et al., 2016](#); [Ford et al., 2017](#)). In this study, we first characterize how topographic metrics respond to tectonics and lithology at the catchment- and segment- scale and use segment-scale differences in channel steepness to estimate relative erodibilities for different lithologies. Then, we evaluate whether our estimated relative erodibilities can account for spatial

variability in lithology to first order. Finally, we use lithology-calibrated metrics in conjunction with previous studies to refine fault growth and linkage history in the Gulf of Corinth. The results enable us to quantify lithological effect through a simple topographic analysis, with important implications for tectonic interpretations of topographic metrics in lithologically heterogeneous landscapes.

2.2 Geological background

The W-E trending Corinth rift in central Greece initiated within a diffuse zone between the Eurasian and Aegean-Anatolian plates at around 4-5 Ma (Figure 2.1a; Armijo et al., 1999; Ford et al., 2013; Gawthorpe et al., 2018). This N-S extension, superimposed on the Hellenide thrust sheets, was characterized by continued northward fault migration and increasing extension rates over time (Figure 2.1b; Ford et al., 2017; Gawthorpe et al., 2018). The modern extension rates of 5-15 mm/year make it one of the fastest extending rifts in the world (Bell et al., 2008). The Gulf of Corinth and Alkyonides represent the current active part of the rift, with the Ionian Sea to the west and the Aegean Sea to the east (Figure 2.1a; Ford et al., 2017). Their episodic connection to the ocean in geological time is fundamentally controlled by subsiding sills (i.e., the west Rion sill, the east Corinth Isthmus) and fluctuating sea level (Gawthorpe et al., 2022).

In general, the tectono-stratigraphic evolution of the Corinth Rift can be divided into two main phases (Figure 2.1b; Ford et al., 2017; Gawthorpe et al., 2018). The first rift phase (from 5.0–3.6 to 2.2–1.8 Ma) was mostly characterized by small and widely distributed onshore faults. In these shallow and discretized depocenters, drainages including antecedent rivers deposited the “Lower Group” with alluvial to lacustrine sediments to the west (Ford et al., 2017) and Gilbert deltas to the east (Gawthorpe et al., 2018). The second rift phase can be further divided into two substages (Figure 2.1b; Ford et al., 2013; Nixon et al., 2016). In the early substage, the rift underwent a ~15km northward fault migration and significantly increased extension rates at ca. 2.2-1.8 Ma (Ford et al., 2013; Gawthorpe et al., 2018). A range of N- and S- dipping faults controlled the deposition of onshore sediments belonging to the “Middle Group” and the offshore seismic unit 1 (Nixon et al., 2016; Ford et al., 2017). The proximal Middle Group consisted of coarse-grained Gilbert deltas with elevation up to ~950 m (Rohais et al., 2007; Backert et al., 2010; Fernández-Blanco et al., 2019). In contrast, finer-grained

sediments such as turbidites and basinal mudstones were deposited in more distal offshore areas, some of which are now onshore (Cullen et al., 2019; Gawthorpe et al., 2022). At ca. 780 ka (Figure 2.1b; Gawthorpe et al., 2022), there is marked fault migration northwards (Ford et al., 2013), strain localization on N-dipping faults, and a basin-wide unconformity offshore (Nixon et al., 2016). After this time (from ca. 780 ka to present), depocenters migrated to the current south coast of the Gulf where the onshore “Upper Group” and offshore seismic unit 2 have been deposited (Figure 2.1b; Hemelsdaël and Ford, 2014; Nixon et al., 2016). Notably, fault interaction and the activation of faults in locations further north (e.g., the PSA fault; Figure 2.1a) occurred again at ca. 400 ka (Ford et al., 2017), and offshore depocenters started to link and a local unconformity developed in the western Gulf at ca. 340 ka (Nixon et al., 2016). In this contribution, we particularly focus on the landscape evolution in the south coast after ca. 780 ka.

Along with northward-migrating faults, pre-rift basement and syn-rift strata are systematically exposed across the fault strike in the study area. The syn-rift stratigraphy, including proximal coarse-grained alluvial deposits and deltas, distal sandstone and siltstone, as well as poorly consolidated sediment, is controlled by northward-migrating faults and therefore typically occurs to the north of the faults (i.e., hangingwall, Figure 2.1b, Watkins et al., 2020). In contrast, the pre-rift basement, characterized by complex lithologic units deformed during the Hellenide Orogeny, is preferentially exposed to the south of the faults (i.e., footwall, Figure 2.1b, Bornovas and Rondogianni-Tsiambaou, 1983; Armijo et al., 1996). Existing work has argued that rivers in the region may record active tectonics (e.g., Demoulin et al., 2014). Zondervan et al. (2020b) illustrated how a knickzone in the Vouraikos catchment (i.e., VO in Figure 2.1) represents a wave of incision upstream of the East Eliki fault that has clearly migrated through uplifted Gilbert Delta stratigraphy that is known to be younger than ca. 780 ka. However, field studies also argue that differing rock types may have influenced the style of landscape response to active faulting (e.g., Zondervan et al., 2020b), and have markedly affected the grain size characteristics of sediment exported to the Corinth basin (Watkins et al., 2020). In particular, some studies have provided limited constraints on bedrock erodibility in terms of K in Equation 2.1a. For instance, Zondervan et al. (2020b) have exploited the unit

stream power model ($n=1$) to constrain the bedrock erodibility for limestone ($\sim 1.6 - 2 \times 10^{-6} \text{ yr}^{-1}$) and conglomerates ($\sim 6 \times 10^{-6} \text{ yr}^{-1}$) in the Vouraikos catchment (i.e., VO in Figure 2.1). By comparing the modelled ($n=1$) and estimated sediment volumes in the Corinth basin, Pechlivanidou et al. (2019) have classified outcropping rock types in this region into bedrock of high ($4 \times 10^{-6} \text{ yr}^{-1}$), intermediate ($2.5 \times 10^{-6} \text{ yr}^{-1}$) and low erodibility ($1.5 \times 10^{-6} \text{ yr}^{-1}$). In addition, Gallen and Fernández-Blanco (2021) have used a Bayesian approach to invert the river long profiles and marine terraces in four eastern catchments and found the best-fit erodibility ranges from $1.6 - 6.2 \times 10^{-6} \text{ yr}^{-1}$ for $m = 0.5$ and $n = 1$. However, the relative impact of this lithological variation across the southern margin of the Gulf has not been explored.

Modern observations show catchment-averaged annual precipitation rate in our study area is $\sim 740 \text{ mm/yr}$, ranging from $\sim 540\text{--}880 \text{ mm/yr}$ (Watkins et al., 2019), although there are point-based variations in rainfall along the rift axis of ca. a factor of 2. Temporal climate changes during glacial-interglacial cycles have been recorded by onshore geomorphic markers such as terraces (e.g., Armijo et al., 1996; Palyvos et al., 2008) and by offshore alternating high-amplitude reflective and non-reflective seismic packages in seismic unit 2 (Nixon et al., 2016; Watkins et al., 2019; Gawthorpe et al., 2022). Model- and pollen-based constraints on paleoclimate for the region suggest the precipitation rate at last glacial maximum of ca. 700 to 1000 mm/yr, approximately -5% to +20% of the modern (Watkins et al., 2018; Gallen and Fernández-Blanco, 2021). However, further constraints on spatial and temporal variations in paleoclimate are currently lacking.

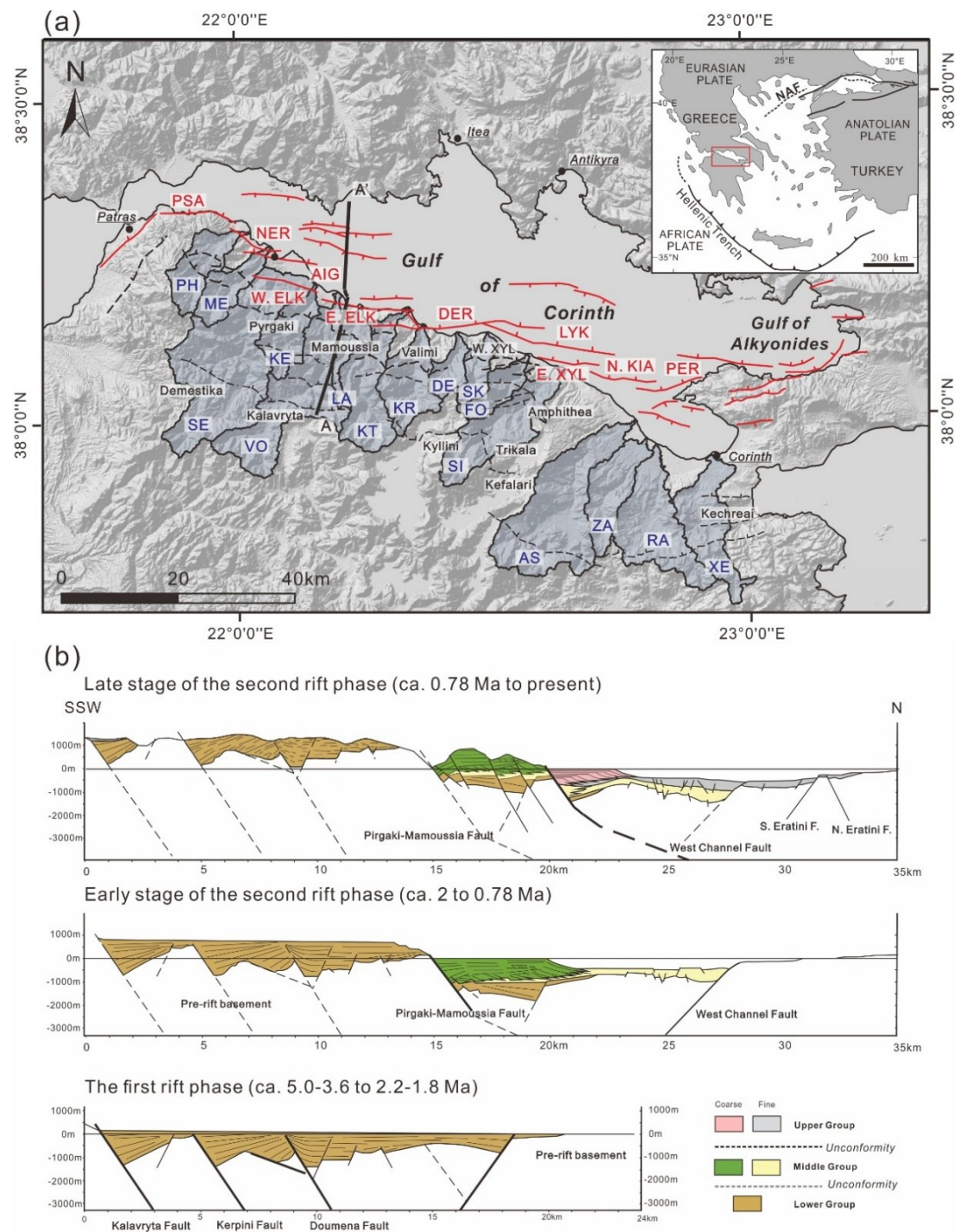


Figure 2.1. (a) Hillshaded map of Gulf of Corinth showing faults and catchment boundaries. Inset map shows regional setting within the tectonic framework of the Aegean. Catchment abbreviations (blue): AS-Asopos, DE-Deveni, FO-Fonissa, KR-Krios, KT-Krathis, LA-Ladopotis, ME-Meganitis, PH-Phoenix, RA-Razanis, SE-Selinoutos, SK-Skoupeiko, SI-Sithas, VO-Vouraikos, XE-Xerias, ZA-Zapantis. Active fault abbreviations (red): AIG-Aigion fault, DER-Deveni fault, E. ELK-East Eliki fault, E. XYL-East Xylokastro fault, LYK-Lykopia fault, N. KIA-North Kiato fault, PER-Perachora fault, PSA-Psathopyrgos fault, NER-Neo Erineos fault, W. ELK-West Eliki fault, W. XYL-West Xylokastro fault. Inactive faults are shown as dashed black curves. (b) cross-section A-A' in (a) showing tectono-sedimentary evolution of the study area (from Ford et al., 2017).

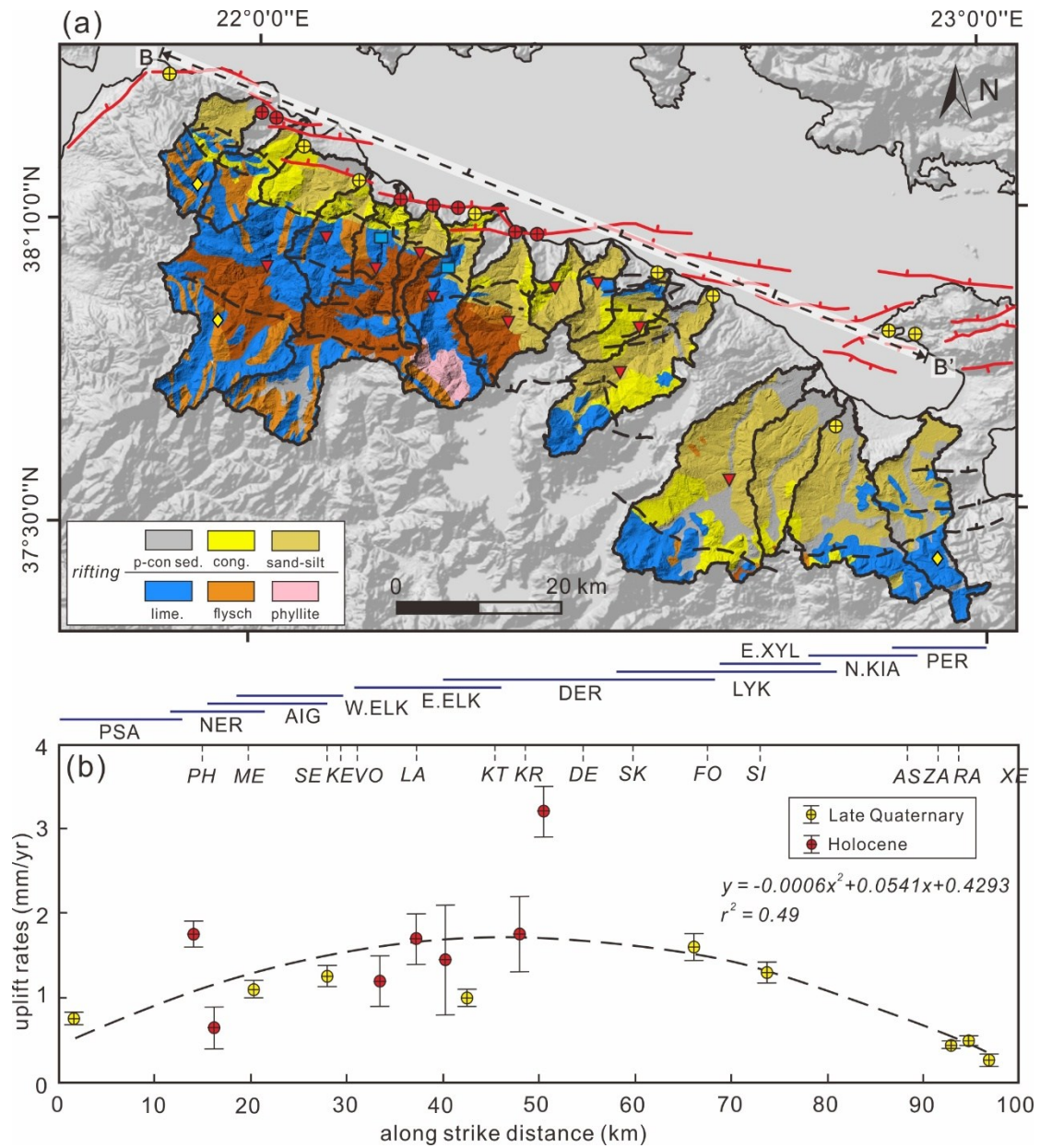


Figure 2.2. (a) Simplified geological map of the studied catchments in the Gulf of Corinth, shown with faults (active-red curves, inactive-dashed black curves) and knickpoints (see section 4.2), and (b) compiled uplift rates (see supplementary Table 2.S1) plotted along the dashed line B-B' in (a).

2.3 Methods

2.3.1 Lithological and tectonic frameworks

Here we digitized the distribution of multiple rock types in our study area, including pre-rift (Mesozoic) limestones and cherts, phyllites-quartzites, sandy turbidite flysch, and syn-rift (Neogene-Quaternary) alluvial to fluvial conglomerate, deltaic conglomerate, sandstone-siltstone and poorly consolidated sediment, as mapped by [Bornovas and Rondogianni-](#)

Tsiambaou (1983) (Armijo et al., 1996; Ford et al., 2013; Gawthorpe et al., 2018). The erodibility value in Equation 2.1a embeds a range of factors including erosional resistance, climatic conditions, among others (Whipple, 2004; Gallen, 2018). In our study area, rock strength and channel width that control erosional resistance are suggested to be closely associated with lithology (Zondervan et al., 2020b) and glacial-interglacial precipitation change is relatively limited (see section 2). In addition, rock strength likely also correlates with the age of non-igneous rocks as they tend to become more consolidated by diagenesis over time (Campforts et al., 2020). Thus, we hypothesise that rock types with significantly different lithologies and/or ages would have variable erodibility values. To make a first-order estimate of erodibility variations, we grouped rock types into six main units: 1) phyllite; 2) flysch; 3) limestone (lime); 4) conglomerates (cong.); 5) sandstone and siltstone (sand-silt); 6) poorly consolidated sediment (p-con sed.) (Figure 2.2a). This grouping is mainly based on lithofacies descriptions and age, and we hypothesise variations within the units to be smaller than variations between them (e.g., Kent et al., 2021).

In order to constrain the uplift framework, we used the locations of active and major inactive faults from Nixon et al. (2016) and Gawthorpe et al. (2018). Here a NW-SE line that approximates the fault strike (Figure 2.2a) was used to enable further along strike analysis. For example, the distributions of active faults were projected along the line (Figure 2.2). Along this line, we also compiled and plotted Holocene and late Quaternary uplift rates from previous studies (e.g., Collier et al., 1992; Pirazzoli et al., 1994; Armijo et al., 1996; Stewart and Vita-Finzi, 1996; Flotte et al., 2001; Houghton et al., 2003; De Martini et al., 2004; McNeil and Collier, 2004; Pirazzoli et al., 2004; McNeil et al., 2005; Pechlivanidou et al., 2019; supplementary Table 2.S1) (Figure 2.2b). In general, uplift rates are higher near the centre and lower towards the eastern and western ends of the southern margin of the Gulf (Figure 2.2b; Pirazzoli et al., 2004; Bell et al., 2009). In contrast to these active faults, few constraints have been provided for the currently inactive faults. Barret et al. (2019) have estimated a sedimentation rate of ~ 0.65 mm/yr for the Selinous delta which was deposited on the immediate hangingwall of the western Pyrgaki-Mamoussia fault (the dashed black fault in between the SE and KT in Figure 2.1). Using $\sim 1:2-3$ as the uplift to subsidence ratio (Bell et al.,

2018), we estimated a footwall uplift rate of $\sim 0.32\text{--}0.21$ mm/yr for the fault. Finally, to approximate how uplift rate decays as a function of distance away from a fault due to the back-tilting of fault-bounded blocks we focused on the Vouraikos and Sithas rivers (VO and SI in Figure 2.1) where uplift rates are well-constrained across fault strike (de Gelder et al., 2019; Zondervan et al., 2020b). We found a linear decay model with a rate of ~ 0.08 mmyr $^{-1}$ km $^{-1}$ can broadly reproduce uplift rate variations across fault strike within the range of ~ 12 km upstream of the fault, beyond which the decay rate significantly decreases (supplementary Figure 2.S1). This fulcrum position where uplift rate approaches zero is likely located >25 km upstream of the fault (supplementary Figure 2.S1) and we assumed this is applicable to other faults in this study.

2.3.2 Topographic and river profile analysis

Topographic data acquired from the ASTER digital elevation model (DEM) were merged and projected in WGS 1984 UTM Zone 34N. The ASTER DEM has a ~ 30 m horizontal resolution and a ~ 17 m vertical resolution (Gesch et al., 2016). Following Watkins et al. (2019), we used a threshold of 300 pixels (0.27 km ^{2}) to derive a drainage network, and 16 catchments along the southern margin of the Gulf of Corinth with drainage areas larger than 5 km ^{2} were then extracted in ArcMap.

Topographic and river profile analysis were performed using the MATLAB-based TopoToolbox (Schwanghart and Scherler, 2014). To explore the potential relationship between tectonics, lithology, and topography, we calculated channel steepness and hillslope angle by pixel. For steady-state rivers, we can solve Equation 2.1a such that:

$$S = (U/K)^{1/n} A^{-m/n}, \quad (2.2)$$

where $(U/K)^{1/n}$ is the channel steepness (k_s), and m/n is the concavity (θ) (Snyder et al., 2000). The topographic gradient, S , defined as the derivative of elevation, z , with respect to downstream distance, x (i.e., $S = dz/dx$) can be further integrated to give:

$$z = z(x_b) + \left(\frac{k_s}{A_0^\theta} \right) \chi, \quad (2.3)$$

$$\chi = \int_{x_b}^x \left(\frac{A_0}{A_x} \right)^\theta dx, \quad (2.4)$$

where x_b is base level, χ is an integral quantity and A_0 is a reference drainage area (Perron and Royden, 2013; Bernard et al., 2019). To calculate normalized channel steepness (k_{sn}), two approaches are often used to choose a reference concavity (θ_{ref}) that approximates the m/n value (Wobus et al., 2006; Gailleton et al., 2021a; Leonard et al., 2023). The first approach uses the optimal concavity (θ_{opt}), which is determined by imposing a best-fit linear regression line in a logarithmic slope-area plot or maximizing the collinearity among river profiles in a z - χ plot (e.g., Gailleton et al., 2021a). However, the θ_{opt} cannot be used to constrain the m/n value unless erodibility and uplift rate can be considered uniform (Whipple, 2000, see more in section 3.3). The alternative approach, however, exploits the expectation that the m/n value should fall within a theoretical range of 0.35-0.6 (Whipple, 2000; Kirby and Whipple, 2012). The narrow concavity range enables an assumed θ_{ref} near 0.5 to not differ significantly from the actual value (e.g., Leonard et al., 2023), leading to relatively accurate χ and k_{sn} (Gailleton et al., 2021a). In our case, we expect erodibility would vary spatially with lithology units and so we employed an assumed $\theta_{ref} = 0.5$ to calculate k_{sn} (e.g., Demoulin et al., 2015; Leonard et al., 2023). Here we first used the “crs” function ($\tau_{crs} = 0.5$, $K_{crs} = 20$) in TopoToolbox to reduce inherent errors in elevation values (z) along river profiles. This function smooths the elevation values using nonparametric quantile regression, where τ_{crs} specifies the quantile, and K_{crs} determines the degree of smoothing (Schwanghart and Scherler, 2017). We then used $\theta_{ref}=0.5$ to determine χ (Equation 2.4) and calculated k_{sn} as the derivative of the z - χ plot, with A_0 set to 1 km² (Equation 2.3). In addition, we derived hillslope angle using a D8 algorithm (Schwanghart and Scherler, 2014). We evaluated the extent to which topography is responding to tectonics and/or lithology by visually and statistically comparing k_{sn} and slope with the established lithological and tectonic frameworks. We do not explicitly consider rainfall variation in time and space in this framework, because it is not well enough constrained for the past 780 ka to be included confidently (section 2), but we return to the impact of this choice in the discussion (section 5.2).

To extend our regional or catchment-scale analysis into segment scale, we extracted river long profiles to identify knickpoints. The term ‘knickpoint’ is used to refer to points where channel

gradient changes abruptly along a river long profile (Whittaker and Walker, 2015). Here, we also defined 'segment' as knickpoint-bounded river reaches within which erosion rate does not change abruptly. To this end, we focused on knickpoints that separate zones with different erosion rates. For rivers crossing normal faults, this type of knickpoint (tectonic knickpoint) is often triggered by tectonic perturbations, such as an increase in fault uplift rate (e.g., Whittaker and Walker, 2015; Gallen and Wegmann, 2017). The tectonic knickpoint initiated at the base level will sweep through the entire river network at predictable rates (Berlin and Anderson, 2007; Whittaker and Boulton, 2012). In this case, these genetically related knickpoints (i.e., from the same base level change) often appear in both trunk and tributaries at a similar elevation. Together, these knickpoints characterize an upstream propagating incisional wave that rejuvenates the downstream catchment portion, enabling it to adjust to new boundary conditions (i.e., the increased slip rate) (Kirby and Whipple, 2012). Evidence that landscapes on the southern margin of the Gulf of Corinth to a marked extent reflect this process include 1) Upper Pleistocene Gilbert delta stratigraphy that is now partly incised upstream of active faults reflecting knickpoint migration in the last ca. 780 ka (Hemelsdael and Ford, 2016; Zondervan et al., 2020b); 2) Concentration of knickpoints upstream of active structures in the southern margin of the Gulf of Corinth, which are not replicated on the northern margin, despite similar lithotypes (Watkins, 2019); and 3) a broad increase in sediment supply to the Gulf of Corinth in the last ca. 780 ka (McNeill et al., 2019).

However, identifying such tectonic knickpoints from river profiles can be complex as other factors such as erodibility variations and river reorganizations can also generate knickpoints (Boulton et al., 2014; Scheingross et al., 2020). In general, lithological knickpoints by erodibility variations do not characterize an incisional wave; rather, they are often locally developed near lithological boundaries or faults (e.g., Gabet, 2020; Dulanya et al., 2022). From this, three criteria were commonly used to identify the trunk knickpoint origin: 1) the existence of tributary knickpoints at a similar elevation (e.g., Miller et al., 2013); 2) a downstream increase in k_{sn} (Beeson and McCoy, 2022); and 3) its co-incidence with lithological boundaries and faults (e.g., Wobus et al., 2006). Here we manually identified those prominent points that mark a significant gradient change in the river profiles and $z-\chi$ plots (e.g., Whittaker and Walker, 2015;

Gallen and Wegmann, 2017). Note, we mostly focused our knickpoint analysis on the topography below the Gilbert-type deltas that are known to have been uplifted since the ca.780 ka fault migration (e.g. Zondervan et al., 2020b). We used the height of their topsets (~950 m) as a maximum estimation to broadly delimit the post-ca.780 ka topography (Figure 2.5) as these deltas are exceptionally well preserved and serve as an excellent reference frame (Ford et al., 2013; Fernández-Blanco et al., 2020; Zondervan et al., 2020b). Here, we first classified those knickpoints that met at least two criteria as potential tectonic knickpoints (Table 2.1, see supplementary material). If a fault increases its slip rate, this base level change will initiate tectonic knickpoints in multiple rivers crossing that fault. Thus, we checked our initial knickpoint identification by examining whether similar knickpoints can be found in adjacent rivers that cross the same fault. This enabled us to verify the potential tectonic knickpoints we classified before and to identify three more tectonic knickpoints that met only one criterion-a downstream increase in k_{sn} (in the AS, SK, and SE, Table 2.1). In principle, knickpoint height, measured from where they were generated (e.g., the basin-bounding fault), is a function of initiation time and uplift rate. Thus, we expect tectonic knickpoints from the same stage would have their heights depend mainly on uplift rate (e.g. Miller et al., 2012). Based on this expectation, we tentatively correlated knickpoints along strike by following the uplift rate trend that peaks near the rift centre and decays towards east and west (Figure 2.2b). In these rivers, 'knick1', 'knick2' and 'knick3' refer to the lower, middle, and upper sets of tectonic knickpoints, respectively (cf. Figure 2.5f), and 'L' refers to lithological knickpoints. Finally, a ± 30 m error bar is used to reflect the uncertainty in knickpoint elevation and height. Table 2.1 summarized the key information for knickpoints identified in the post-ca. 780 ka river long profiles (i.e., below 950 m) of the 16 trunk rivers.

To conduct a segment-scale analysis, we exploited the expectation that tectonic knickpoints would migrate as an incisional wave. We first used the elevation contours reached by these tectonic knickpoints to subdivide trunk channels and catchments. This is to delineate the trunk segments and catchment portions that have adjusted to the new boundary conditions. For each trunk segment, we extracted k_{sn} for the segment by each individual lithology (lithology-grouped k_{sn}); likewise, we extracted lithology-grouped slope from the corresponding

catchment portion. Finally, we evaluated the response of segment-scale metrics to lithology and tectonics by examining the variations in lithology-grouped k_{sn} and slope within and between segments and catchment portions (Figure 2.6).

2.3.3 Topographic constraints on lithological erodibility

To estimate lithological erodibility from topography, we focused on these knickpoint-bounded segments within which erosion rate is unlikely to change abruptly. Making the simplifying assumption that erosion rate is relatively uniform within these segments, the erodibility differences between different lithologies (e.g., conglomerates vs. limestone) can be derived from intra-segment variations in lithology-grouped k_{sn} (e.g., Gailleton et al., 2021b; Takahashi et al., 2022)

$$K_{cong}/K_{lime} = (k_{sn-lime}/k_{sn-cong})^n, \quad (2.5)$$

In this study, we use limestone to assess the relative erodibility of other lithologies because it is 1) widely exposed; and 2) documented to be the most resistant to mechanical erosion in previous studies, relative to other lithologies in the area (Zondervan et al., 2020b). For segments with length-scales > 5 km running perpendicular to a normal fault, we note that uplift can decrease upstream due to footwall back-tilting; the impact of this is discussed in section 5.1 and 5.2.

A key parameter in Equation 2.5 is n , the slope exponent in the stream power law (Equation 2.1). Implicit in $n=1$ for a classic version of stream power law (Tucker and Whipple, 2002) is that erosion rate is directly proportional to the energy being expended by the flowing water on that bed (Hancock et al., 1998; Whipple and Tucker, 1999). For instance, Ferrier et al. (2013) have explicitly constrained a linear relationship between million-year-scale incision rate and unit stream power in a Hawaiian island. This has also been demonstrated in areas dominated by normal faulting, such as the Central Apennines, where the unit stream power of rivers is clearly correlated with fault-driven footwall uplift (e.g., Whittaker et al., 2007). Nevertheless, recent authors have argued for a wider range of n values. By fitting a power-law regression line through a k_{sn} - E plot ($k_{sn} = (E/K)^{1/n}$), Lague (2014) found the best-fit n ranges from ~ 1 to 4 for 10 rivers where a steady state is argued, with $n \sim 2$ being the most common. Harel et

al. (2016) also obtained a mean value of n at 2.6 ± 0.4 using a globally compiled dataset. Despite the wide range of reported n , the assumption of $n=1$ has also been justified in many settings (e.g., Lavé and Avouac, 2001; Roberts et al., 2012; Ferrier et al., 2013; D’Arcy and Whittaker, 2014; Smith et al., 2024). In our study area, Gallen and Fernández-Blanco (2021) in a DEM-based study suggested that a best-fit n may be up to 7 for four eastern catchments by inverting river long profiles and marine terraces. However, their best-fit n is much higher compared to other studies using the stream power law; they recognized their high n might result from their best-fit model that is biased towards the profile below the knickzone. In contrast, Zondervan et al. (2020b) have explicitly shown a field derived discharge-based unit stream power model ($n=1$) effectively constrained the downstream distribution of incision in the VO catchment (Figure 2.1), which they successfully compared to independent constraints on footwall uplift. Pechlivanidou et al. (2019) have also employed $n=1$ in their pyBadlands model that successfully reproduces the sediment volume archived in the Corinth basin when spatial variable erodibilities were used.

Because we still do not fully understand the extent to which fluvial processes can be scaled up in time and space to generate river profiles (Wapenhans et al., 2021) and because existing field-based results in the VO catchment are consistent with $n = 1$, we use a stream power erosion model where $n=1$ in the first instance. This allows the relative erodibility to be inversely proportional to limestone-normalized k_{sn} (Equation 2.5), and we therefore calculated the reciprocal of limestone-normalized k_{sn} in each studied segment to constrain the plausible range of the relative erodibilities for different lithologies within this model (Figure 2.7a). However, we further test the choice of $n=1$ and $\theta_{ref} = 0.5$ (or $m=0.5$) using a knickpoint propagation model (below). We return to the potential impact of $n \neq 1$ in section 5.3. Notably, the general tectonic history can still be recovered from landscapes even if the n value assumed differs from the actual value due to the trade-off between the assumed n and estimated K (Goren et al., 2014).

To test whether the assumption of $n=1$ and $\theta_{ref} = 0.5$ is reasonable, we evaluated the extent to which knickpoints modelled under this assumption reproduce the spatial distributions of

observed knickpoints in tributaries. Equation 2.1a can be rewritten to model the knickpoint propagation rate ($\frac{dx}{dt}$) at any location (e.g., [Berlin and Anderson, 2007](#)):

$$\frac{dx}{dt} = K(x)A(x)^m S^{n-1}, \quad (2.6)$$

By integrating knickpoint propagation rate (i.e., $\frac{dx}{K(x)A(x)^m S^{n-1}}$) along the channel length, the time for a knickpoint to travel from base level to its current location (knickpoint age, τ) can be defined as:

$$\tau(x) = \int_{x_b}^x \frac{dx}{K(x)A(x)^m S^{n-1}}, \quad (2.7)$$

Equation 2.7 shows that if erodibility, drainage area, and slope are known, then knickpoint age can be determined from its position, and vice versa. In particular, knickpoints from the same base level event (i.e., genetically associated), being in trunk or tributaries, should have the same age. These enable us to use the position of a given trunk knickpoint to estimate its age, and then use this age to model the positions of its genetically associated knickpoints in tributaries. Here we performed this modelling under $m=0.5$ and $n=1$ for those trunk knickpoints whose genetically associated knickpoints in tributaries can be observed ([Table 2.1](#), [see supplementary material](#)). Note the knickpoint age derived so far are in a relative sense, as only relative erodibilities constrained from previous analysis were used in this modelling.

To convert these relative values to absolute ages, we need an ‘anchor’ age to first estimate absolute erodibilities. In our study area, the offshore basin-wide unconformity provides a first order estimate on knickpoint initiation time (ca. 780 ka, [Gawthorpe et al., 2022](#)). In addition, the prominent knickpoint in the VO catchment ([Figure 2.1](#)) has been estimated to initiate at 700 ± 200 ka from palynological data ([Ford et al., 2013](#)) and at ca. 700 ka from fault throw ([Zondervan et al., 2020b](#)). Here we chose this well-constrained knickpoint to constrain bedrock erodibility and used $\sim 700 \pm 100$ ka to account for the age uncertainties above. Erodibility values were adjusted to enable the modelled knickpoint, after $\sim 700 \pm 100$ ka of migration (Equation 2.7 with $m=0.5$ and $n=1$), matches the position of the observed knickpoint. In this process, we adjusted erodibility under two scenarios: the first assumed uniform

erodibility regardless of lithological variations (here termed uniform erodibility), while the second used relative erodibilities to allow erodibility to vary according to lithology (here termed lithology-dependent erodibilities). We then used the lithology-dependent erodibilities to estimate the age of tectonic knickpoints in other catchments, given their positions were known (Table 2.1, Equation 2.7 with $m=0.5$ and $n=1$).

To illustrate the effectiveness of lithology-dependent erodibilities, we focused on the optimized channel concavity (θ_{opt}), which is expected to fall within the range of 0.35-0.6 if variations in K and U are properly accounted for (Whipple et al., 2000). Here, we used “mnoptimvar” function in TopoToolbox (Schwanghart and Scherler, 2014) to obtain θ_{opt} . This function segments χ -distance profiles (z - χ plots) into bins with equal χ and searches for the θ_{opt} that minimizes elevation differences in these bins (Schwanghart and Scherler, 2014). However, we implemented our analysis on χ_K -distance profiles (z - χ_K plots), where χ_K

$(\int_{xb}^x \left(\frac{A_0}{K_x^{\frac{1}{m}} A_x} \right)^{\theta_{opt}} dx)$ integrates not only drainage area but also erodibility to incorporate the impact of potential spatial variations in erodibility. To this end, the drainage area input argument in the “mnoptimvar” function was adjusted to incorporate erodibility ($K_x^{\frac{1}{m}} A_x$, with $m=0.5$). Depending on whether a uniform erodibility or lithology-dependent erodibilities was used to calculate χ_K , the z - χ_K plots were therefore designated as uncalibrated or lithology-calibrated river profiles. As erodibility was derived from post-ca.780 ka river profiles, we restricted this concavity analysis to 11 rivers that develop the knickpoint interpreted to reflect the ca.780 ka tectonic event (‘knick2’) and to the part of their river profiles that is below the elevation of that knickpoint.

2.4 Results

2.4.1 Topographic metrics across the south coast of the Gulf

Figure 2.3 shows the relationship between topography and lithology. In general, the mean k_{sn} of pre-rift limestone (lime), phyllite and flysch are statistically higher (at the 95% confidence level) than that of conglomerates (cong.), sandstone to siltstone (sand-silt) and poorly consolidated sediment (p-con sed.) according to an unpaired t test (Figure 2.3b). Because k_{sn}

distributions may deviate from normality, median values were also compared as a more robust measure of central tendency. The median k_{sn} values follow the same overall pattern, decreasing from phyllite > flysch > lime. > cong. > sand-silt > p-con sed., albeit with marked overlaps. In particular, this pattern can also be observed in hillslope angle (Figure 2.3d).

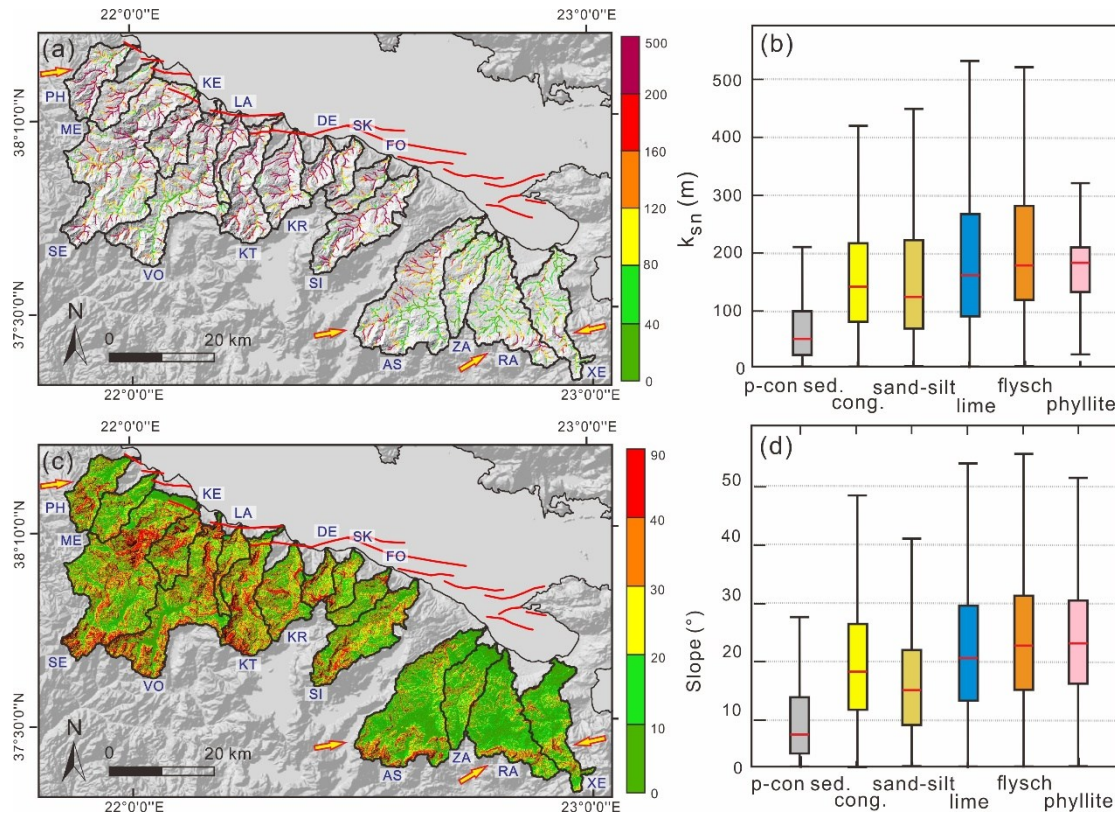


Figure 2.3. The relationship between topography and lithology. Left panels are maps of (a) k_{sn} and (c) hillslope angle. Right panels show box plots of (b) k_{sn} and (d) hillslope angle for the lithologies in the study area, with colours and lithology abbreviations in accordance with Figure 2.2.

A possible interpretation for the systematic variations in median k_{sn} and hillslope angle is that different lithologies have variable erodibility, with steeper channels and hillslopes corresponding to less erodible rocks (e.g., Yanites et al., 2017). However, the marked overlaps also highlights the possibility that such variations could simply result from the spatial distribution of these lithologies, with bedrock lithologies such as phyllite, flysch and limestone preferentially distributed in high uplift rate zones (e.g., Roda-Boluda and Whittaker, 2018). To test these two competing hypotheses, we focused on the N-S variations (from catchment

outlet to headwater) in the k_{sn} and hillslope angle map (Figure 2.3a, c). Along this N-S direction, bedrock usually transitions from soft syn-rift rocks (e.g., p-con sed. and cong.) to harder pre-rift rocks (e.g., limestone) (Figure 2.2a) while erosion rates would decay systematically due to transient dynamics (e.g., Reinhardt et al., 2007) and the back-tilting effect (Zondervan et al., 2020b). Figure 2.3a and c show that despite decreasing erosion rates, a general southward increase trend in k_{sn} and slope can be observed in catchments where such lithological transitions exist (e.g., PH, XE, RA). Moreover, we find these southward increases are relatively abrupt and coincide with lithological transition boundaries (e.g., from sand-silt to limestone in catchments XE, RA). Thus, we demonstrate that despite the potentially depressing impact of spatial variable erosion rates, lithology exerts a discernible control on topography in the study area.

To explore the tectonic control on topography, we plot catchment-averaged k_{sn} and slope along rift margin strike (Figure 2.4). Notably, the 16 studied catchments along the rift margin (approximately W-E direction) not only encounter different uplift rates (Figure 2.2b), but also have variable proportions of different lithologies (Figure 2.2a). Figure 2.4a shows that k_{sn} decays from ~220 m in the rift centre (e.g., catchment DE, KR, KT) to ~80 m in the east (e.g., catchment XE, RA, ZA, AS). However, we do not observe an obvious decrease from centre to west (Figure 2.4a). Similarly, we find catchment-averaged slope only shows an eastward decreasing pattern (from ~20° to ~9°), with median slope for four western catchments (i.e., KE, SE, ME, PH) all being around 18° (Figure 2.4b). Here we interpret the clear eastward decrease in both k_{sn} and slope to reflect the control of normal faulting (e.g., Fernández-Blanco et al., 2020). Furthermore, we hypothesise that the lack of westward-decreasing pattern can at least partly be attributed to the higher proportions of pre-rift bedrock lithologies in the western catchments compared to their central counterparts.

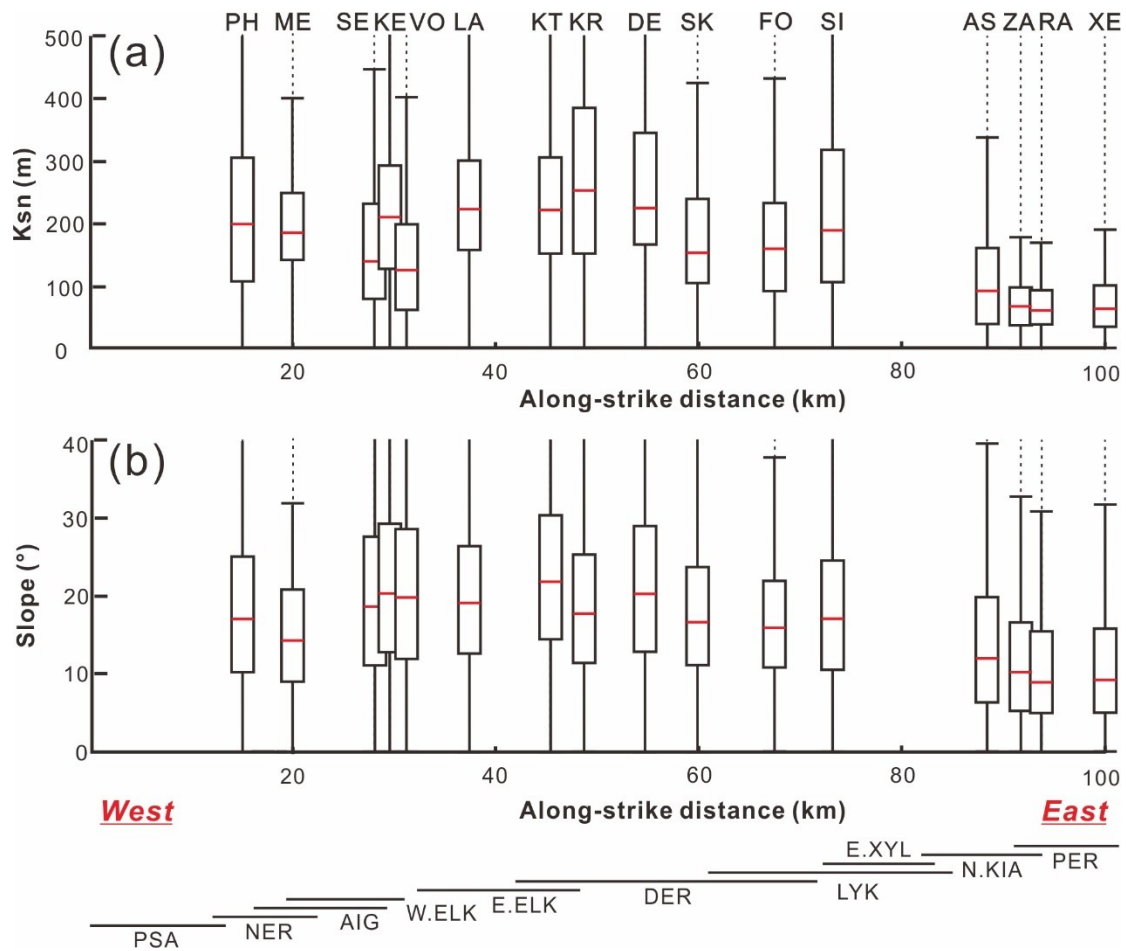


Figure 2.4. Box-and-whisker plots showing along-strike variations (*cf.* Figure 2.2) in (a) channel steepness and (b) hillslope angle, along with the distribution of the faults.

These results show that the topography in the Gulf of Corinth records both tectonic and lithological signals. In particular, these signals are likely interfering with each other in these averaged metrics, which obscures some expected patterns in the regional or catchment-scale analysis.

2.4.2 Post-ca.780 ka river profiles, knickpoints and topography

Figure 2.5a-e show the long profiles of rivers crossing active fault segments. Of the 16 rivers analysed in this study, most channels display a convex shape with multiple knickpoints upstream of the faults. In contrast, channels towards both the eastern and western ends of the rift have long profiles that come closer to more 'typical' concave profiles (RA, ZA, ME, PH, Figure 2.5a, e). These convex channels have previously been interpreted to reflect transient responses to active faulting (Demoulin et al., 2015; Fernández-Blanco et al., 2019), but

lithological difference likely contributes to some convexities in these lithologically heterogeneous catchments (e.g., Gabet, 2020).

To divide river profiles into segments, we integrated three commonly used criteria to distinguish tectonic and lithological knickpoints in profiles below 950 m (Table 2.1; Figure 2.5). Along the N.KIA fault and PER fault, two rivers (i.e., the RA and ZA, Figure 2.5a) have mostly concave profiles, while the other two rivers, XE and AS, each develop a tectonic knickpoint that marks a downstream k_{sn} increase (the yellow diamond at ca. 460 m and the red triangle at ca. 240 m in Figure 2.5a, Table 2.1). For three rivers upstream of the LYK fault and E. XYL fault (i.e., the SI, FO, SK), a set of tectonic knickpoints has been identified for their heights systematically increasing from ca. 450 m to 590 m along fault strike (red triangles in Figure 2.5b, Table 2.1). These systematic variations in elevation support the identification of the lower knickpoint in the SK as tectonic in origin (Figure 2.5b). In contrast, the other two knickpoints that either coincide with a fault or show no downstream increase in k_{sn} are classified as lithological type in the FO and SK (grey circles in Figure 2.5b, Table 2.1). Along the DER fault, the three rivers (i.e., the DE, KR and KT) have four tectonic knickpoints that meet at least two criteria (Figure 2.5c, Table 2.1). Except for the lowest tectonic knickpoint at ca. 370 m in the KT (the blue square in Figure 2.5c), the other three knickpoints have their elevations increases from ca. 410 m (DE) to ca. 700 m (KR and KT) (red triangles in Figure 2.5c). In addition, two knickpoints, including the one in the DE (ca. 680 m) and the one in the KT (ca. 185 m) near the DER fault, are identified as lithological type as they do not meet any criteria (grey circles in Figure 2.5c, Table 2.1). Along the W. ELK fault and E. ELK fault, four rivers (i.e., the LA, VO, KE, SE, Figure 2.5d) have one or two tectonic knickpoints that meet at least two criteria (Table 2.1). In particular, a set of knickpoint with their elevations systematically decreasing from ca. 610 m to ca. 460 m can be identified (red triangles in Figure 2.5d). Apart from this knickpoint set, the VO has another lower tectonic knickpoint at ca. 415 m and the KE has an upper knickpoint at ca. 700 m (the blue square and the yellow diamond in Figure 2.5d). Two rivers crossing the NER fault and AIG fault (i.e., the ME, PH, Figure 2.5e) only have tectonic knickpoints that are close to or above 1000 m, suggesting these knickpoints were likely initiated before ca. 780 ka. In addition, two lithological knickpoints that do not meet any

criterion exist in their profiles (grey circles in Figure 2.5e, Table 2.1). Figure 2.5f plots knickpoint height, measured from where they were generated (e.g., the basin-bounding fault) against their along-strike distance, measured as the projected position of channel outlets on the line in Figure 2.2a. Notably, the knickpoints with systematic variations in elevation also have heights that correlate with uplift rate (Figure 2.5f). Thus, we tentatively correlated this knickpoint set (termed ‘knick2’) with the post-ca.780 ka tectonic event; this interpretation is examined later in section 4.3. The knickpoints below and above the ‘knick2’ were correlated and labelled as ‘knick1’ and ‘knick3’, respectively (Figure 2.5f). In our analysis, obvious knickpoints below ‘knick2’ were not identified in some profiles, but we do not rule out the possibility that ‘knick1’ might develop as minor features in these profiles (Figure 2.5).

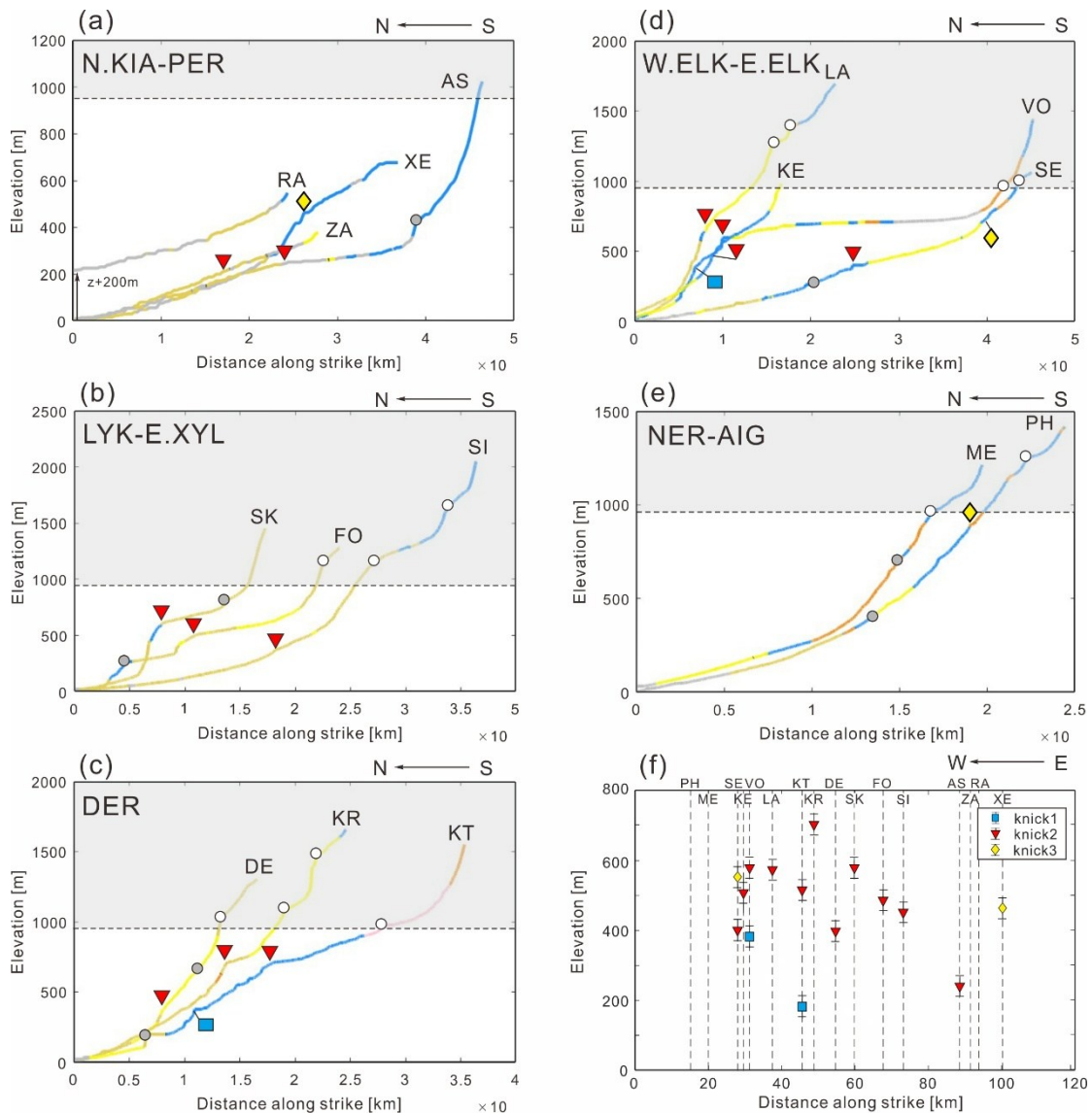


Figure 2.5. (a-e) Long profiles of trunk rivers crossing faults, coloured by lithology (cf. Figures 2.2 and 2.3). (f) Along-strike variation in knickpoint height. The character corresponds to catchments shown in Figure 2.1. In profiles below 950 m, the lower, middle, and upper sets of tectonic knickpoints are referred to as ‘knick1’ (blue squares), ‘knick2’ (red triangles) and ‘knick3’ (yellow diamonds), respectively, and prominent lithological knickpoints are shown as grey circles; in profiles above 950 m (grey bars), knickpoints are marked as white circles; more details are provided in section 3.2 and supplementary material. Fault abbreviations: AIG-Aigion fault, DER-Deveni fault, E. ELK-East Eliki fault, LYK-Lykoporia fault, N. KIA-North Kiato fault, PER-Perachora fault, NER-Neo Erineos fault, W.ELK-West Eliki fault, W.XYL-West Xylokastro fault.

To evaluate the response of segment-scale topography to tectonics and lithology, Figure 2.6 shows the lithology-grouped k_{sn} and slope for the lowest trunk segment in 11 catchments with the ‘knick2’ set. In most segments, the median k_{sn} and slope are highest for the limestone and lowest for the p-con sed., with the sand-silt and the cong. exhibiting modest k_{sn} and slope (Figure 2.6). Notably, we find the median k_{sn} variations between different lithologies (e.g., limestone vs. cong.) are often markedly larger at the segment scale than their regional counterparts (cf. Figures 2.6a and 2.3b). This pattern can also be observed in slope, albeit with a smaller magnitude (cf. Figures 2.6b and 2.3d). In addition, we note lithology-grouped metrics do not show the along strike pattern found in uplift rate (Figures 2.2c and 2.6). This is interpreted to reflect local erodibility variations that introduce relatively high or low values, disrupting the potential along strike pattern.

Our results show that both tectonic and lithological knickpoints develop in post-ca.780 ka river long profiles, and three sets of tectonic knickpoints can be correlated. In the knickpoint-bounded segments, we find topography is more sensitive to lithology, as illustrated by the larger k_{sn} and slope variations between different lithologies, than in catchment-scale topography. However, lithology-grouped metrics at the segment scale might be more likely to be biased by local features, highlighting the need to consider multiple segments to obtain regionally representative values.

Table 2.1. Data associated with the knickpoints identified in the 16 trunk rivers.

River	Along-strike distance (km)	downstream k_{sn} increase	tributary knickpoints	co-incidence with lithological boundaries and faults	type	elevation (m)	height (m)	age (ka)
XE	100	×	×	×	L	275	275	919±184
		√	×	√	knick3**	463	463	1823±365
RA	93.6	√	×	×	L*	238	238	1036±207
ZA	91.4	×	√	×	L*	75	75	287±57
		√	×	×	L*	214	214	878±176
AS	88.3	√	×	×	knick2**	241	241	571±114
		×	×	×	L	313	313	1552±310
		√	×	×	L*	428	428	1874±375
		×	×	√	L*	580	580	3389±678
SI	73	×	√	√	knick2**	451	451	695±139
FO	67.5	√	×	×	L*	328	300	316±63
		√	×	√	knick2**	514	486	723±145
SK	59.7	√	×	×	knick2*	588	579	597±119
		×	×	√	L*	826	817	1302±260
DE	54.6	×	×	×	L	165	165	323±65
		√	√	×	knick2**	409	409	660±132
		×	×	×	L	679	679	1157±231
KR	48.6	√	√	√	knick2***	702	702	616±123
KT	45.4	×	×	×	L	184	0	294±59
		√	√	√	knick1***	367	183	331±66
		√	√	×	knick2**	699	515	1167±223
LA	37.3	×	×	√	L*	150	113	341±68
		√	√	×	knick2**	610	573	562±112
VO	31.1	√	×	√	knick1**	415	382	356±71
		√	√	×	knick2**	611	578	623±125
KE	29.4	×	×	√	L*	155	93	146±29
		×	×	√	L*	228	166	301±60

SE	27.9	√	×	√	knick2**	569	507	956±191
		×	×	√	L*	714	652	1940±388
		×	×	√	L*	292	232	587±117
		√	×	√	knick2**	460	400	1018±204
		√	√	×	knick3**	613	553	1661±332
ME	19.7	×	×	×	L	334	270	1369±274
		×	×	×	L	689	625	-
PH	15	×	×	×	L	224	167	297±59
		×	×	×	L	436	379	-
		√	√	×	knick3**	864	807	-

Note '*' is used to indicate the number of criteria that a knickpoint met. 'knick1', 'knick2' and 'knick3' refer to the lower, middle, and upper sets of tectonic knickpoints, respectively (c.f., [Figure 2.5f](#)), and 'L' refers to lithological knickpoints (see section 4.2). Knickpoint age is calculated with a ±20% error bar, if preferred lithology-dependent erodibilities are constrained (see section 4.3).

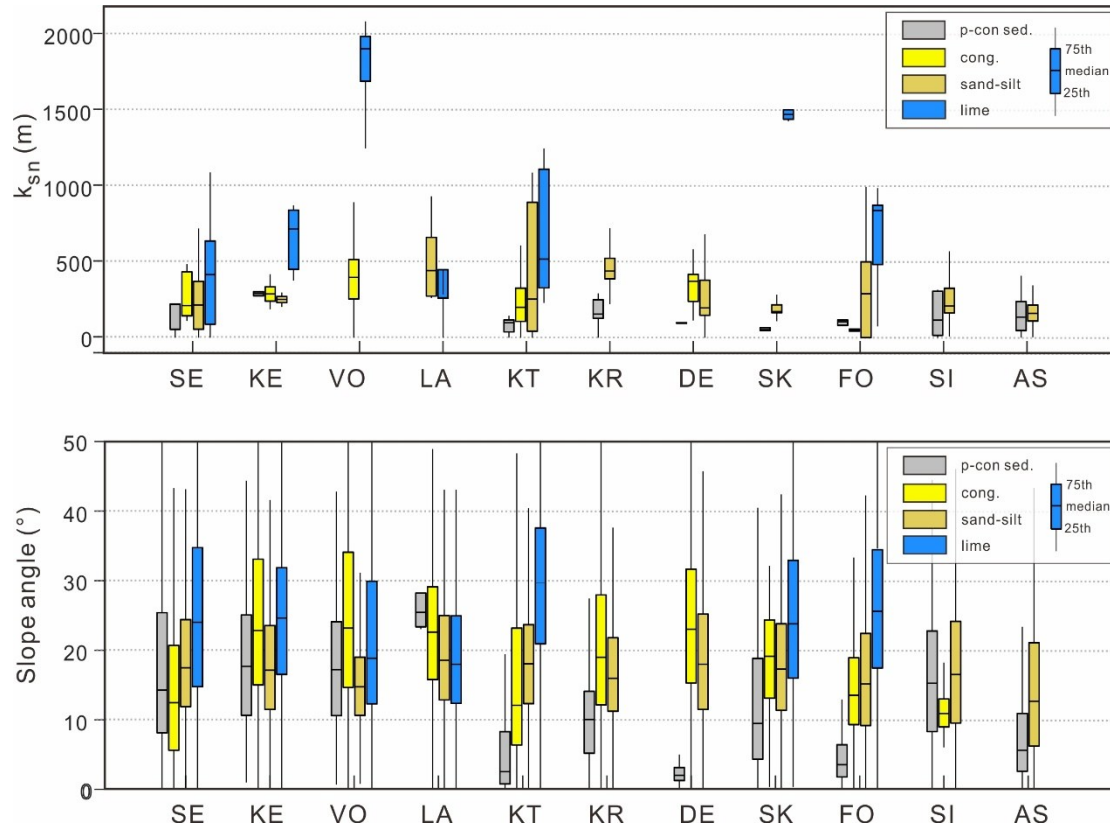


Figure 2.6. Box-and-whisker plots show lithology-grouped k_{sn} (a) and slope (b) for the lowest trunk segment in catchments with 'knick2' knickpoints. In the same segment, variations between different lithologies are used to evaluate the impact of lithology; for the same lithology, variations across different segments are used to evaluate the impact of tectonics (cf. Figure 2.3 and Figure 2.4). Note the lowest segment for VO and KR is the segment below the 'knick1' knickpoint.

2.4.3 Lithological erodibility and calibrated metrics

Making the simplifying assumption that uplift/erosion rate is relatively uniform within the segments, intra-segment k_{sn} variations were used to constrain the range of relative erodibility for different lithologies (Figure 2.7a). The quartiles indicate that the k_{sn} for limestone is generally 3-8 times higher than the k_{sn} for p-con sed., 1-4 times higher than cong. and 2-3 times higher than the k_{sn} for sand-silt. Nonetheless, the data show that there are substantial variations in relative erodibilities in different segments (Figure 2.7a). To derive reference relative erodibilities for the study area, we proceeded with a first-order estimate ($K_{lime} : K_{cong.} : K_{sand-silt} : K_{p-con sed.} = 1 : 2 : 3 : 4$, Figure 2.7a). These relative erodibilities are then used in a knickpoint propagation model where $m=0.5$ and $n=1$ are assumed, to model knickpoint migration in such lithologically heterogeneous catchments (see section 3.3). Figure 2.7b maps

the predicted locations of tributary knickpoints that are genetically related to a given trunk knickpoint in the VO catchment (see more knickpoint propagation in supplementary material). The close match of the modelled and observed tributary knickpoints testifies to the effectiveness of the knickpoint propagation model (Figure 2.7b, Miller et al., 2013). Moreover, the propagation model and the prominent, $\sim 700 \pm 100$ ka knickpoint in the VO catchment (Zondervan et al., 2020) enable conversion of relative erodibilities to absolute values. Using this approach with the assumption that erodibility varies according to lithology (i.e., relative erodibility), then an erodibility of $0.6\text{--}0.8 \times 10^{-6} \text{ yr}^{-1}$ is obtained for limestone. If we use the upper end-member of $0.8 \times 10^{-6} \text{ yr}^{-1}$ as a maximum, the erodibility for cong., sand-silt, and p-con sed. would then be 1.6, 2.4, $3.2 \times 10^{-6} \text{ yr}^{-1}$ respectively. Alternatively, we also obtain an erodibility of $0.8\text{--}1.0 \times 10^{-6} \text{ yr}^{-1}$ by using a uniform erodibility for different lithologies. In general, erodibility values obtained in both scenarios are geologically plausible compared to previous studies in areas of active faulting (e.g., Whittaker and Boulton, 2012). The lithology-dependent erodibilities correspond well with the three erodibility classes (1.5 , 2.5 , and $4 \times 10^{-6} \text{ yr}^{-1}$) used by Pechlivanidou et al. (2019). Additionally, our lithology-calibrated erodibility values are comparable to, though generally lower than, the values estimated by Gallen and Fernández-Blanco (2021) and Zondervan et al. (2020b).

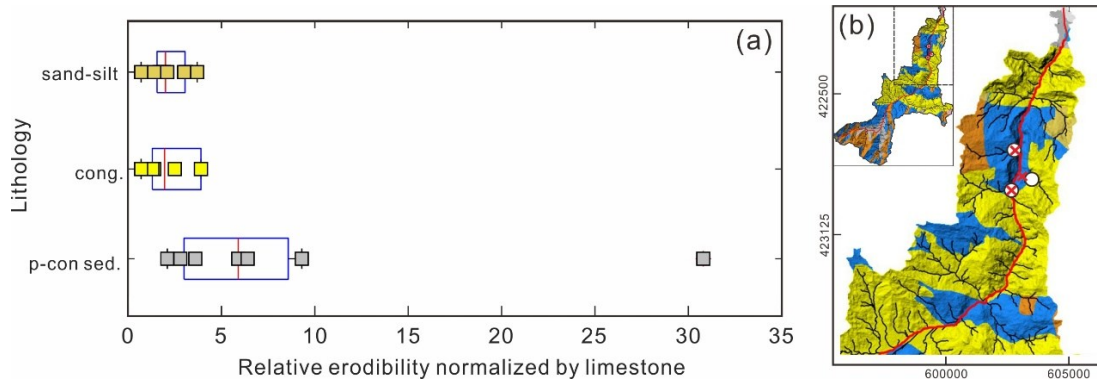


Figure 2.7. (a) Boxplot of the range of relative erodibility (i.e., the reciprocal of limestone-normalized k_{sn}) for different lithologies; (b) Comparison between observed (open circles) and modelled (red crosses) positions of migrating knickpoints; more in supplementary material.

To illustrate the differences between a uniform erodibility and lithology-dependent erodibilities, we compared optimized θ (θ_{opt}) obtained from 11 studied post-780 ka river profiles under these two different scenarios (Figure 2.8). If a uniform erodibility was used, almost half of the studied river profiles (4/11) have their θ_{opt} below 0.35 (Figure 2.8). In

contrast, almost all (10/11) rivers have their θ_{opt} within the range of 0.35-0.6 using the lithology-dependent erodibilities (Figure 2.8). This is consistent with Pechlivanidou et al. (2019) (see their supplementary Table S1), supporting the use of 0.5 as a reference θ .

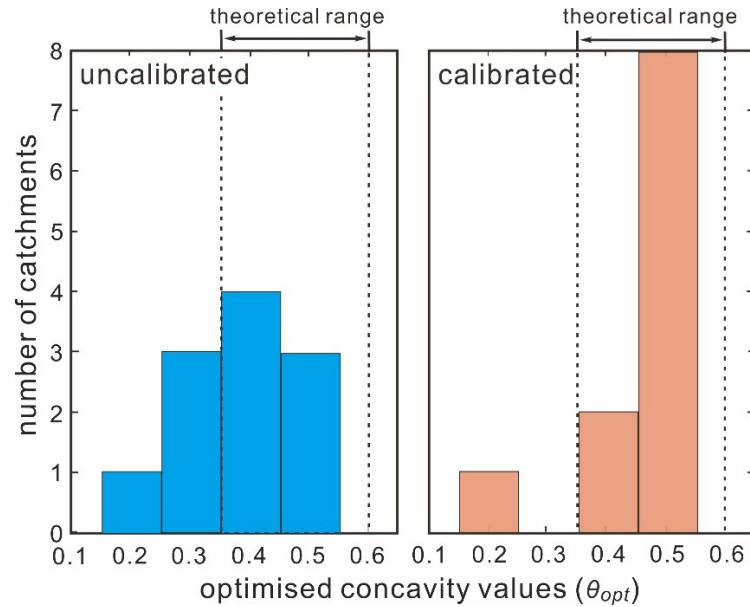


Figure 2.8. Optimized ϑ (ϑ_{opt}) for post-ca.780 ka river profiles in both a uniform erodibility (blue) and lithology-dependent erodibilities (orange) scenarios.

Figure 2.9A plots knickpoint height against their age, estimated from lithology-dependent erodibilities and the knickpoint propagation model (Table 2.1). Three distinct clusters of knickpoints are revealed, supporting our previous height-based interpretation (Figure 2.5f). We interpret this to show that three episodes of uplift rate increase, with estimated timing of ca. 300-400 ka, ca. 550-950 ka and ca. 1800 ka, have occurred in the study area (Figure 2.9A). Moreover, the ratio of height to age represents a knickpoint vertical migration rate (Figure 2.9A), equivalent to uplift rate in the case of $n=1$ (Goren et al., 2014). In general, our results show that vertical migration rates of knickpoints in the ‘knick2’ and ‘knick1’ sets fall within the range of 0.35-1.33 mm/yr, and two knickpoints in the ‘knick3’ set have lower rates (0.20-0.35 mm/yr) (Figure 2.9A). Figure 2.9B shows that along the W.ELK and E.ELK, the age of knickpoints in the ‘knick2’ set decreases from ca. 1020 ± 200 ka in the SE catchment to ca. 560 ± 110 ka in the LA catchment. For rivers further east, knickpoints initiated from four faults exhibit similar ages (ca. 650 ka), except the one in the KT catchment (ca. 1170 ± 220 ka). Nevertheless, the knickpoint age appears to peak in the FO catchment, where the DER and

LYK faults overlap, and decrease gradually towards both east and west (Figure 2.9B). Finally, two ‘knick1’ knickpoints from the DER and LYK faults have similar ages, with the knickpoint in the KR catchment (ca. 330 ± 70 ka) slightly older than the one in the VO catchment (ca. 360 ± 70 ka) (Figure 2.9B).

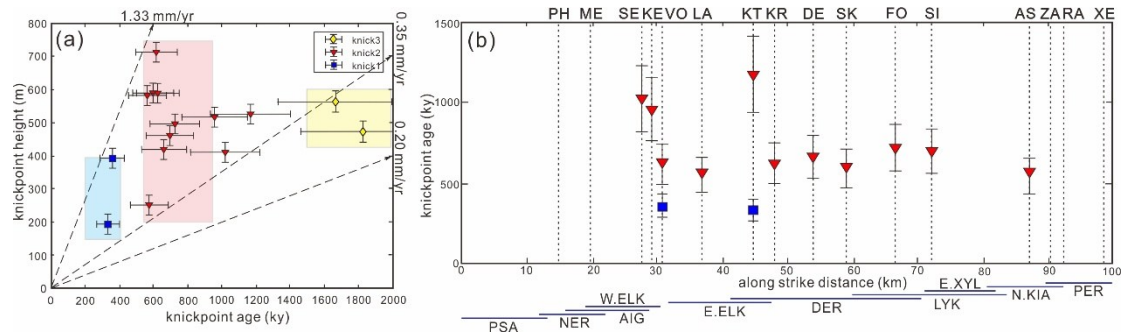


Figure 2.9. (a) Knickpoint height plotted against knickpoint initiation age for all identified tectonic knickpoints; (b) Along-strike age variations for knickpoints in both ‘knick1’ and ‘knick2’ sets, with distribution of the faults.

2.5 Discussion

Our study demonstrates that tectonic and lithological controls on tectonically active areas might be unravelled through a simple topographic analysis using readily available data. The results show that: 1) landscapes are responding to both tectonics and lithology, with segment-scale metrics being more sensitive to lithological differences than in their catchment-scale counterparts; 2) relative rock erodibility estimates ($K_{\text{lime}} : K_{\text{cong.}} : K_{\text{sand-silt.}} : K_{\text{p-con sed.}} = 1 : 2 : 3 : 4$) are geologically plausible, but display substantial variability in different segments; and 3) lithology-calibrated metrics such as knickpoint age and vertical migration rate fall within a reasonable range, highlighting the opportunity to refine tectonic signals. It is noted that tectonic knickpoints, which separate segments with contrasting erosion rate, are fundamental to our results. Thus, we first revisit the identification of tectonic knickpoints and then discuss the above three points in more detail.

2.5.1 Identification of tectonic knickpoints

In this study, three commonly used criteria were integrated to evaluate the origin of knickpoints. An important finding is that knickpoints, depending on the criteria used, could be identified as having a tectonic or non-tectonic origin (Table 2.1). This highlights the

uncertainties in identifying tectonic knickpoints (c.f., [Gabet et al., 2020](#); [Beeson and McCoy, 2022](#)) and we take this opportunity to comment on this issue. First, the lack of a downstream k_{sn} increase may not be conclusive evidence for excluding a tectonic origin. For example, the knickpoint in the SI catchment, despite lacking a downstream increase in k_{sn} averaged over a segment, is identified as tectonic type in our study ([Figure 2.5b](#), [Table 2.1](#)). This interpretation is supported by its association with several knickpoints at a similar elevation in tributaries and adjacent rivers, as well as its location relative to mapped faults or lithological boundaries ([supplementary Figure 2.S6](#)). In fact, this knickpoint does correlate with a k_{sn} increase, but at a local scale (~ 5 km) ([supplementary Figure 2.S6](#)); however, the k_{sn} used here and in many other studies to determine knickpoint origins is often averaged over a whole segment (e.g., [Whittaker and Walker, 2015](#); [Gallen and Wegmann et al., 2017](#)). In other words, whether k_{sn} increases downstream of a knickpoint could vary depending on the approach used to obtain or average k_{sn} ([Smith et al., 2022](#)). In addition, tectonic knickpoints in our study are not always accompanied with tributary knickpoints at a similar elevation. This could be due simply to the lack of tributaries in some catchments (e.g., [DE and SK](#)) or/and to the heterogeneous lithology that can smear out knickpoints ([Yanites et al., 2022](#)). On the other hand, migrating knickpoints with genetically related tributary knickpoints could also result from river reorganization (e.g., [Giachetta and Willet, 2018](#)). In our study area, several river reorganization events have occurred (e.g., in [DE](#), [SI](#), [AS](#), [Gawthorpe et al., 2018](#); [Fernández-Blanco et al., 2020](#)), but we favour fault acceleration as the ultimate driver for knickpoint generation. This is because 1) these reorganization events have been suggested to be induced by fault back-tilting at ca. 780 ka ([Gawthorpe et al., 2018](#); [Rohais et al., 2007](#); [Zelilidis, 2000](#)), and 2) correlatable knickpoints across adjacent rivers testify to a regional event rather than local river reorganization ([Figure 2.5f](#)). Finally, some tectonic knickpoints could appear near lithological contacts (e.g., [SK and KT](#), [Figure 2.5](#)), as knickpoint propagation might be impeded by hard bedrock lithologies ([Miller et al., 2013](#); [Beeson and McCoy, 2022](#)).

By integrating multiple criteria, we identified 16 unambiguous tectonic knickpoints, which is fewer than previous studies in this area ([Demoulin et al., 2015](#); [Fernández-Blanco et al., 2019](#)). This is partly because only prominent knickpoints below 950 m - the height of Gilbert delta's

topsets (Fernández-Blanco et al., 2019) - are identified, as we aim to focus on post-ca.780 ka long profiles. Another reason is that some knickpoints previously identified as having a tectonic origin could be attributed to a lithological origin when multiple criteria are considered (c.f., Fernández-Blanco et al., 2019). Notably, footwall back-tilting can change the convexity of river long profiles and can lead to the development of internal drained basins in the catchment headwaters (e.g., Whittaker et al., 2007; Attal et al., 2008) that might be expected to be located near the fulcrum position, signifying the transition in the deformation field from uplift to subsidence. However, we do not expect such effect to be significant for the knickpoints in our study because 1) the fulcrum where uplift rate approaches zero is located far upstream (ca. 25km, supplementary Figure 2.S1), and 2) the magnitude of the back-tilting is $< 0.1 \text{ mmyr}^{-1}\text{km}^{-1}$.

2.5.2 Landscape response to tectonics and lithology

The results show that both the steepness and hillslope angle vary as a function of bedrock lithology (Figure 2.3). This is well illustrated by the generally high k_{sn} and hillslope angle in catchment headwaters where harder bedrocks are exposed. In particular, this pattern cannot be explained by tectonics as catchment headwaters are expected to exhibit lowest uplift rates due to transient dynamics and back-tilting. However, we also demonstrate that heterogeneous lithology alone cannot explain both the transient state and along-strike variations in k_{sn} and hillslope angle. Indeed, tectonic knickpoints in convex profiles clearly testify to fluvial responses to uplift rate changes (Figure 2.5). In addition, both k_{sn} and slope show a clear eastward decrease (Figure 2.3) that correlates with along strike variations in uplift rate (Figure 2.2b). Consequently, our results explicitly demonstrate that landscapes in the Gulf of Corinth are responding to both tectonics and lithology.

A key observation is that segment-scale metrics are clearly more sensitive to lithological differences than their catchment-scale counterparts (c.f. Figures 2.3 and 2.6). In our case, erosion rate is spatially variable due to transient dynamics or/and the back-tilting effect. Notably, the former is most obvious across a tectonic knickpoint (e.g. Miller et al., 2013; Roda-Boluda and Whittaker, 2018) and the latter - back-tilting - depends largely on the midpoint positions of studied river reaches across the fault strike assuming the linear decreasing rate is

applicable to other faults in our study area (Zondervan et al., 2020b). Thus, erosion rate variations are more pronounced at the regional scale, where tectonic knickpoints exist and studied reaches are longer. This explanation is straightforward but provides some implications for other studies in tectonically active areas. For example, many studies evaluated the impact of lithology on topography through analysing the correlation between lithology and topographic metrics averaged over a region (e.g. Miller et al., 2012; Racano et al., 2023). Here, we caution this practice as we show minor variations in median k_{sn} (e.g., limestone vs. sand-silt) at the regional scale can represent an average ~3 times differences at the segment scale (cf. Figures 2.3b and 6a). Here, we propose this discrepancy reflects the extent to which lithological signals have been modified by spatially variable uplift rates (Yanites et al., 2017).

Finally, the potential impact of climate variations needs to be acknowledged. While hard to constrain explicitly over the last ca. 780 ka, spatial variations in catchment averaged rainfall and broad glacial-interglacial precipitation rates do not appear extreme in the Gulf of Corinth (e.g. Watkins et al., 2019). However, it is clear that some variation in climate occurred and this would be expected to influence knickpoint propagation (e.g. Jansen et al., 2011). Nevertheless, we expect such temporal variations would have uniform effects across all studied catchments and therefore would not change our knickpoint age as this is implicitly included in K . We note that modern observations indicate annual precipitation is ~2 times higher at higher elevations relative to lower elevations due to orographic effects, but further spatial constraints in paleoclimate are lacking. Even if we assume this orographic pattern has sustained over the past ca. 780 ka, then k_{sn} in catchment headwaters should be ~1.4 times lower ($m=0.5$) than in catchment outlets. However, we observe the opposite pattern (e.g., in the PH), which is interpreted to be a response to lithological differences (Figure 2.3a). Thus, we consider that spatial variations in climate is not a first-order control in our case.

2.5.3 Are estimated relative erodibilities effective to first order?

Considering that lithological differences are more evident in segments, we used segment-scale steepness differences to estimate relative erodibilities ($K_{lime} : K_{cong.} : K_{sand-silt.} : K_{p-con sed.} = 1 : 2 : 3 : 4$) (Figure 2.7). This is comparable with Zondervan et al. (2020b) who found cong. is three times more erodible than limestone. In the Eastern Jura Mountains, Switzerland, Yanites

et al. (2017) also obtained a two-fold difference between limestone and shale. However, such erodibility contrasts, for example, between limestone and p-con sed., are much smaller than orders of magnitude differences found in some studies (Stock and Montgomery, 1999; Sklar and Dietrich, 2001). Furthermore, the limestone-normalized k_{sn} can vary substantially for different segments (Figure 2.7). These might be attributed to some unaccounted factors in our study, such as local fault damage (e.g. Kirkpatrick et al., 2021), the effect of back-tilting (e.g. Whittaker et al., 2008) and lithofacies variations in the mapped lithological units (e.g. Yanites et al., 2017; Kent et al., 2021). For example, fault-related fracture can significantly reduce the erodibility of a particular lithology (e.g., the limestone in LA, Figure 2.6). This may partly explain the relatively small lithological contrast in erodibility. In addition, an upstream decay in uplift/erosion rate due to back-tilting would cause the downstream reach to exhibit higher steepness than its upstream counterpart. This indicates that the steepness-based estimation on relative erodibility for downstream lithologies (e.g., p-con sed.) is likely depressed, with the extent depending on the reach length and decaying rate (Equation 2.5). The variable extent of depressing and lithofacies variations may contribute some of the scatter (Figure 2.7).

These findings therefore raise the question: to what extent do our obtained relative erodibilities account for lithological differences over a regional scale? To illustrate the effectiveness of these relative erodibilities, we considered how lithology-calibrated long profiles help clarify some seemingly conflicting observations. In some cases, the optimized concavity (θ_{opt}) is used as reference concavity (θ_{ref}) (Gailleton et al., 2021a). Notably, we found almost half of the studied river profiles (4/11) have their θ_{opt} values below 0.35 (Figure 2.8). This seems to contradict with the theoretical and field studies that support $\theta_{ref} \sim 0.5$ (e.g., Whipple et al., 2000; Pechlivandous et al., 2019). However, this might be attributed to the impacts of non-uniform climate, lithology and uplift on θ_{ref} values (e.g., Roe et al., 2002; Duvall et al., 2004; Leonard et al., 2023). If this is the case, then θ_{opt} should be within the theoretical range as long as the influence of non-uniform climate and/or lithology is properly corrected. This is well illustrated by Leonard et al. (2023) who found ~70% of catchments fall within the typical 0.4 to 0.6 range by accounting for the effect of spatially variable precipitation. In our

study area, climate is relatively uniform in the post-ca.780 ka topography and spatially variable lithology might be the main reason for the deviation in the θ_{opt} . Indeed, we found almost all (10/11) rivers have the θ_{opt} value within 0.35-0.6 after calibration (Figure 2.8). This analysis thus supports that our relative erodibilities account for the effect of spatially variable lithology. More broadly, our study agrees with Leonard et al. (2023) that a reference concavity near 0.5, rather than an optimized value from an uncalibrated long profile, is preferred for those areas with spatially variable climate or lithology (c.f., Gailleton et al., 2021a).

Moreover, we assessed the extent to which lithology-calibrated metrics conform to existing constraints on tectonics. In general, previous stratigraphy-based studies demonstrate the basin has mainly experienced two tectonic events from ca. 780 ka to present (Bell et al., 2008; Nixon et al., 2016; Ford et al., 2013; Ford et al., 2017; Gawthorpe et al., 2022). The first tectonic event has been estimated to occur at ca. 700 ± 200 ka for the ELK fault from palynological data (Malartre et al., 2004; Ford et al., 2017), ~ 600 ka for the N.KIA and PER faults based on fluvial topography inversion (Gallen and Fernández-Blanco, 2021), 780 ka for two cored boreholes in the central Corinth basin as the unconformity between seismic unit 1 and 2 is interpreted to correspond to the Brunhes-Matuyama boundary (McNeil et al., 2019) and ~ 620 ka for the same basin-wide unconformity according to seismic character and clinoforms (Nixon et al., 2016). Notably, the initiation age of knickpoints in the ‘knick2’ set, though spatially variable (ca. 550-950 ka, Figure 2.9a, Table 2.1), fall within the broad range suggested by these studies. In addition, the second tectonic event has been estimated to occur at ca. 400 ka for the PSA fault (Ford et al., 2017) and ca. 340 ka for the Corinth basin (horizon H4 in Nixon et al., 2016). This corresponds well with the two knickpoints in the ‘knick1’ set estimated at ca. 330 ± 70 ka and 360 ± 70 ka (Figure 2.9a, Table 2.1). Another important metric is time-averaged knickpoint vertical migration rate (Figure 2.9a), equivalent to uplift rate in the case of $n=1$ (Goren et al., 2014). Our results show the time-averaged vertical migration rate of knickpoints in the ‘knick1’ and ‘knick2’ sets falls within the range of 0.35-1.33 mm/yr (Figure 2.9a). This is comparable to independently constrained uplift rates of 1-1.3 mm/yr over the late Quaternary (e.g., De Martini et al., 2004, Armijo et al., 1996; Pirazzoli et al., 2004; Figure 2.2b). The significantly different lower bounds may be partly attributed to increasing

uplift rate through time (Pirazzoli et al., 2004; Ford et al., 2013) and the shorter averaging timescale for late Quaternary rates (e.g., <350ka, Armijo et al., 1996; De Martini et al., 2004). Likewise, the time-averaged vertical migration rate of the 'knick3' knickpoint in the KE catchment puts an upper bound on uplift rate of the western Pyrgaki-Mamoussia fault (~0.33 mm/yr, Figure 2.9a). This is consistent with an uplift rate of ~0.32-0.21 mm/yr estimated from the subsidence rate of the Selinous delta in the immediate hangingwall of the western Pyrgaki-Mamoussia fault (Figure 2.1; Barret et al., 2019). To first order, our geomorphic results agree with previous studies on tectonic evolution, supporting the effectiveness of our relative erodibilities.

While our results do show that the stream power law with $n=1$ successfully reproduces observed knickpoint positions (supplementary material) and produces reasonable geomorphic constraints on tectonic events over the past ca. 780 ka, it is important to consider the potential impact of this simple choice of n . In the case of $n>1$, relative erodibilities estimated from k_{sn} variations would be much more scattered (Equation 2.5). This highlights the uncertainties in deriving relative erodibilities entirely from k_{sn} variations with an assumed n (e.g., Shi and Shao, 2014), but does not necessarily disprove our efforts to use such k_{sn} variations to constrain a plausible range for relative erodibilities. Furthermore, $n>1$ enables knickpoints to migrate faster in reaches where river slope is high (Equation 2.6). In this case, rivers experiencing higher uplift rate (i.e., steeper slopes) would respond faster to a fault slip rate increase (e.g., Whittaker and Boulton, 2012). In other words, our model with $n=1$ might overestimate the knickpoint age for rivers at fault centres where fault slip rates are expected to be greater. Nevertheless, this does not invalidate our attempts to use this simple model to interpret fault growth and linkage history as we explore below (Section 5.4); instead, this provides an alternative mechanism for knickpoints near the fault centre to propagate further upstream relative to those near the fault tip. In addition, the dependence of response time on uplift rate through slope indicates the spatiotemporal variations in uplift rate would be reduced in the model with $n=1$ (given the knickpoint age in high uplift zones is comparably overestimated). However, our results provide reasonable estimates on the magnitude of

temporal variations in uplift rate (from ~ 0.3 mm/yr to 1.3 mm/yr), supporting the use of $n=1$ (e.g., Goren et al., 2014).

2.5.4 Implications for tectonic evolution in the Gulf of Corinth

In comparison with previous constraints, our geomorphic results provide comparable, but more extensive estimates on the timing and magnitude of tectonic events. This is particularly important for those areas without well-constrained geomorphic markers or chronostratigraphic frameworks. In addition, these extensive estimates provide an opportunity to refine the tectonic model of the Gulf of Corinth. For example, the large age variations for knickpoints in the 'knick2' set might reflect diachronous faulting in the Gulf of Corinth (Figure 2.9b, e.g., Demoulin et al., 2015; Ford et al., 2017). To test this hypothesis, we interpret our geomorphic results in conjunction with previous studies (e.g., Nixon et al., 2016; Gawthorpe et al., 2022; Figure 2.10).

Figure 2.9b shows an eastward-younging trend for knickpoints controlled by the ELK fault. This could be interpreted to reflect the propagation process of the ELK fault, as proposed by Hemelsdael and Ford (2014). If this is the case, then our study would suggest the ELK fault initiated at ca. 1020 ± 200 ka in the west and later in the east (ca. 560 ± 110 ka). Taken at face value, this corresponds well to the 'Early breaching' event in Hemelsdael and Ford (2014). Furthermore, we note this matches with thickness changes which shows the depocenter controlled by the N-dipping ELK fault started from ca. 530 ka (Figures 2.10a, b). Thus, we interpret that the eastward propagation of the ELK fault sequentially initiated a series of knickpoints. Another trend is that the knickpoints appear to have initiated slightly earlier near the FO catchment and spread towards both east and west (Figure 2.9b). Notably, this topographic change is also accompanied by the development of an offshore depocenter with the thickest part near the FO catchment (Figure 2.10b). Here we interpret that these trends reflect the interaction of the DER and LKY faults, whereby the earliest knickpoint and thickest strata correspond to the breached relay ramp. Likewise, we interpret the ca. 340 ka event as a result of the final linkage of the ELK and DER-LYK faults (Figure 2.9b, *sensu* final breaching in Hemelsdael and Ford, 2014). This tectonic event might explain why the H4 horizon (ca. 340ka) is unconformable in the western Gulf (Nixon et al., 2016).

Surprisingly, only two younger unambiguous tectonic knickpoints are identified in our study despite the considerable, basin-wide slip rate increase at ca. 340 ka ([Figure 2.10d](#); [Nixon et al., 2016](#)). In particular, we note the two knickpoints developed in catchments controlled by onshore faults; by contrast, offshore fault-controlled catchment KR, DE and SK, though at a comparable distance to the breached relay ramp, do not develop an unambiguous knickpoint ([Figure 2.5](#)). In an onshore fault-controlled catchment, the change in base level is a combination of an increase in uplift rate and an increase in subsidence rate. If faults are offshore, however, the absolute uplift rate increase can be markedly reduced by the back-tilting effect, and the change in sea level rather than subsidence rate influences the base level change. In our case, the base level change from a reduced uplift rate increase might be further offset by a contemporary sea-level rise (>60m, from MIS10 to MIS9, [Hemelsdael and Ford, 2014](#)). This provides a possible explanation for why knickpoints are preferentially initiated in catchments controlled by onshore faults. A similar observation is also found in south-central Crete, Greece where catchments controlled by offshore faults developed fewer knickpoints compared to nearby onshore fault-controlled catchments ([Gallen and Wegmann, 2017](#)). There, the authors suggest this difference might reflect the less effective erosion efficiency of the oversteepened channels in some offshore fault-controlled catchments ([Gallen and Wegmann, 2017](#)). Further studies are still needed to elucidate the different geomorphological and stratigraphic responses to the same tectonic event.

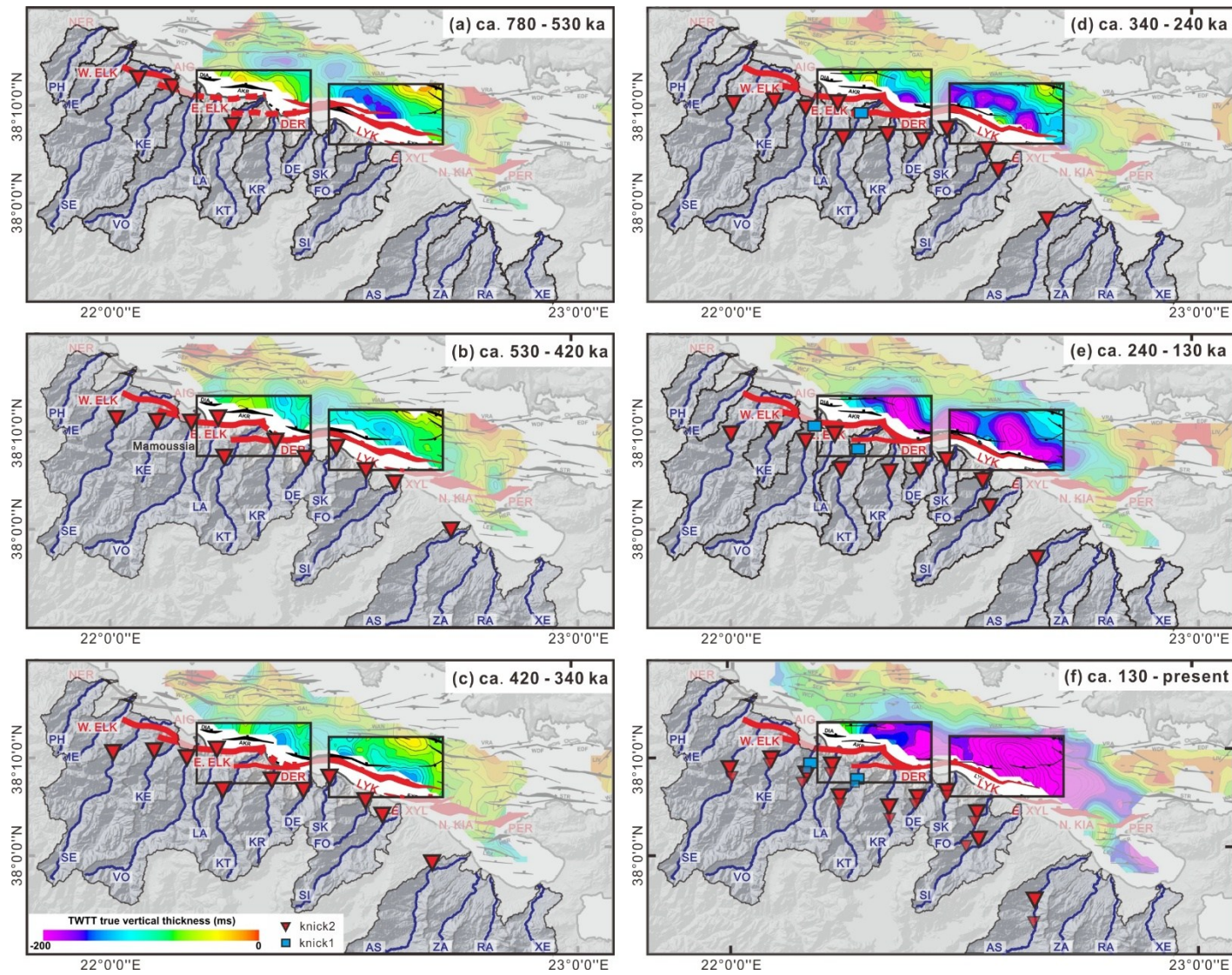


Figure 2.10. Possible fault growth and linkage history reconstructed from knickpoint initiation ages and stratigraphic thickness variations from ca. 780 ka to present (modified from [Nixon et al., 2016](#) and [Gawthorpe et al., 2022](#)). Opaque red triangles and blue squares in (a-f) are the modelled knickpoint positions at the start of each time slice, and their transparent counterparts in (f) mark the current knickpoint positions. Note the areas where faults are interpreted to grow and link are highlighted by the two rectangles. Key faults considered here are marked in red, with their yet-to-be-activated segments shown as dashed lines.

2.5.5 Summary of response to faulting in lithologically heterogeneous landscapes

To sum up our main findings, we use a schematic diagram to illustrate the key topographic changes in a lithologically heterogeneous landscape as faults grow and link ([Figure 2.11](#)). In areas near the fault centre, for instance where the fault nucleated, or at a simple breached relay ramp, two catchments with differing bedrock erodibility detect the signal of uplift rate change at approximately the same time (t_1 , [Figures 2.11a and b](#)). The more competent lithology (blue colours, [Figure 2.11b](#)) will impede rapid knickpoint propagation upstream in plan view, developing steep, angle of repose hillslopes and a substantial high channel steepness knickzone downstream of the knickpoint. However, the knickpoint in the catchment underlain by the softer lithology (i.e., relatively higher erodibility, orange colour, [Figure 2.11b](#)) would propagate further upstream in plan view in the same time period, with downstream reaches and their corresponding hillslopes being relatively less steep (cf. [Figures 2.11a, b](#)). This timescale of knickpoint propagation will be affected by the distributions of bedrock lithology; for instance in [Figure 2.11a](#), the higher erodibility unit (orange) that is not yet incised will in future lead to an increase in knickpoint propagation rates and a composite distribution of channel steepness and hillslopes that will embed both a tectonic and lithological signal. This behaviour could be contrasted with tip propagation of a normal fault as we have detected, for instance, for the ELK fault. Here, a catchment near the tip would detect the tectonic signal later as the fault propagates laterally (t_2 , [Figure 2.11c](#)). In the lower uplift rate near the tip depresses the channel steepness (and potentially hillslope angle if the local threshold has not been reached) downstream of the knickpoint (cf. [Figure 2.11c, b](#)).

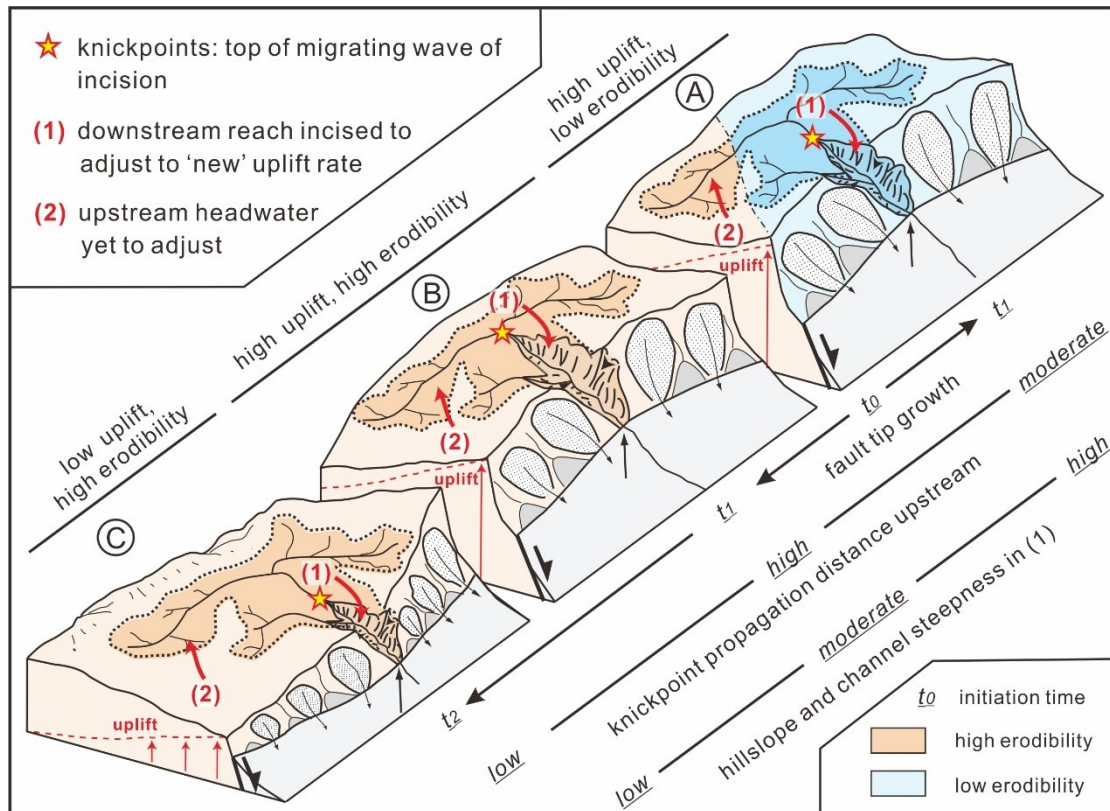


Figure 2.11. Schematic diagram showing contrasting landscape responses to faults growth and linkage under high uplift and low erodibility (a), high uplift and high erodibility (b) and low uplift and high erodibility (c). Note the three catchments that differ in uplift rate and erodibility have similar drainage area.

2.6 Conclusions

This study exploits several geomorphic techniques to unravel tectonic and lithological effects on transient catchments in tectonically active and lithologically heterogeneous areas. The results show that spatial variations in lithology can be accounted for to first order by lithology-dependent erodibilities through a simple topographic analysis. In particular, lithology-calibrated topographic metrics yield key constraints on fault growth and linkage processes. Here we used 16 catchments along the southern margin of the Gulf of Corinth as an example, but this case study illustrates the potential to utilize readily available data to unravel complex tectonic histories from tectonically active but lithologically heterogeneous landscapes. Indeed, most of our key findings listed below may be instructive for other similar areas.

1. Topographic metrics and river long profiles show that the landscapes in our study area are responding to both lithology and tectonics. Moreover, segment-scale metrics are more

sensitive to lithological differences than their catchment-scale counterparts. This is because uplift rate variations are more pronounced at catchment scale, which may suppress landscape sensitivity to lithology.

2. In the knickpoint-bounded segments, steepness variations between different lithologies are used to constrain their relative erodibilities ($K_{\text{lime}} : K_{\text{cong.}} : K_{\text{sand-silt.}} : K_{\text{p-con sed.}} = 1 : 2 : 3 : 4$), which are further converted into actual lithology-dependent erodibilities by modelling a well-constrained, $\sim 700 \pm 100$ ka knickpoint in the Vouraikos (VO) catchment. To first order, the lithology-dependent erodibilities accounts for lithological differences in topography over regional scale. This finding is supported by the observation that if the relative erodibilities were used to calibrate river long profiles in χ distance, we obtain long profile concavities that fall within the theoretical range.

3. Lithology-calibrated metrics provide geomorphic constraints on the timing and the magnitude of tectonic events in the Gulf of Corinth. In general, our results align with previous studies that the basin has experienced two tectonic events from ca. 780 ka to present, with a long-term uplift rate of < 1.3 mm/yr. In particular, along-strike variations in knickpoint age might reflect the eastward growth of the Eliki (ELK) fault and the linkage of the Derveni (DER) and Lykoporia (LKY) faults.

4. Integrating multiple criteria is crucial for accurate identification of tectonic knickpoints. Three commonly used criteria—the existence of associated knickpoints in tributaries, a downstream k_{sn} increase for the same lithology, and its co-incidence with lithological boundaries and faults—cannot in isolation be used as conclusive evidence to identify tectonic knickpoints.

2.7 Acknowledgements

ZZ was supported by an Imperial College Janet Watson Studentship. We are very grateful for the insightful comments from two anonymous reviewers whose suggestions helped to improve this manuscript.

2.8 Supplementary material

2.8.1 Table 2.S1. Compiled uplift rates in the south coast of the Gulf of Corinth.

Locality	Along strike distance (km)	Uplift rate (mm/yr)		References
		Late Quaternary	Holocene	
1	1.7	0.7-0.8		Houghton et al., 2003
2	14.1		1.6-1.9	Palyvos et al., 2008
3	16.2		0.4-0.9	Palyvos et al., 2005; Zondervan, 2017
4	20.4	1.00-1.20		De Martini et al., 2004
5	28	1.25		De Martini et al., 2004
6	33.4		0.9-1.5	Pirazzoli et al., 1994; Stewart and Vita-Finzi, 1996
7	37.2		1.4-2.0	McNeil and Collier, 2004
8	40.3		0.8-2.1	Pirazzoli et al., 1994; Stewart and Vita-Finzi, 1996
9	42.5	0.9-1.1		McNeil and Collier, 2004; De Martini et al., 2004; McNeil et al., 2005
10	48		1.3-2.2	Stewart and Vita-Finzi, 1996
11	50.5		2.9-3.5	Pirazzoli et al., 2004
12	66.1	1.6		Flotte et al., 2001
13	73.7	1.3		Armijo et al., 1996
14	92.8	0.44		Collier et al., 1992
15	94.6	0.5		Leeder et al., 2003; Pechlivanidou et al., 2019
16	96.8	0.2-0.35		Leeder et al., 2003; Pechlivanidou et al., 2019

Note that the localities, as shown in Figure 2.2b, are numbered from west to east.

2.8.2 Figures 2.S1-2.S17

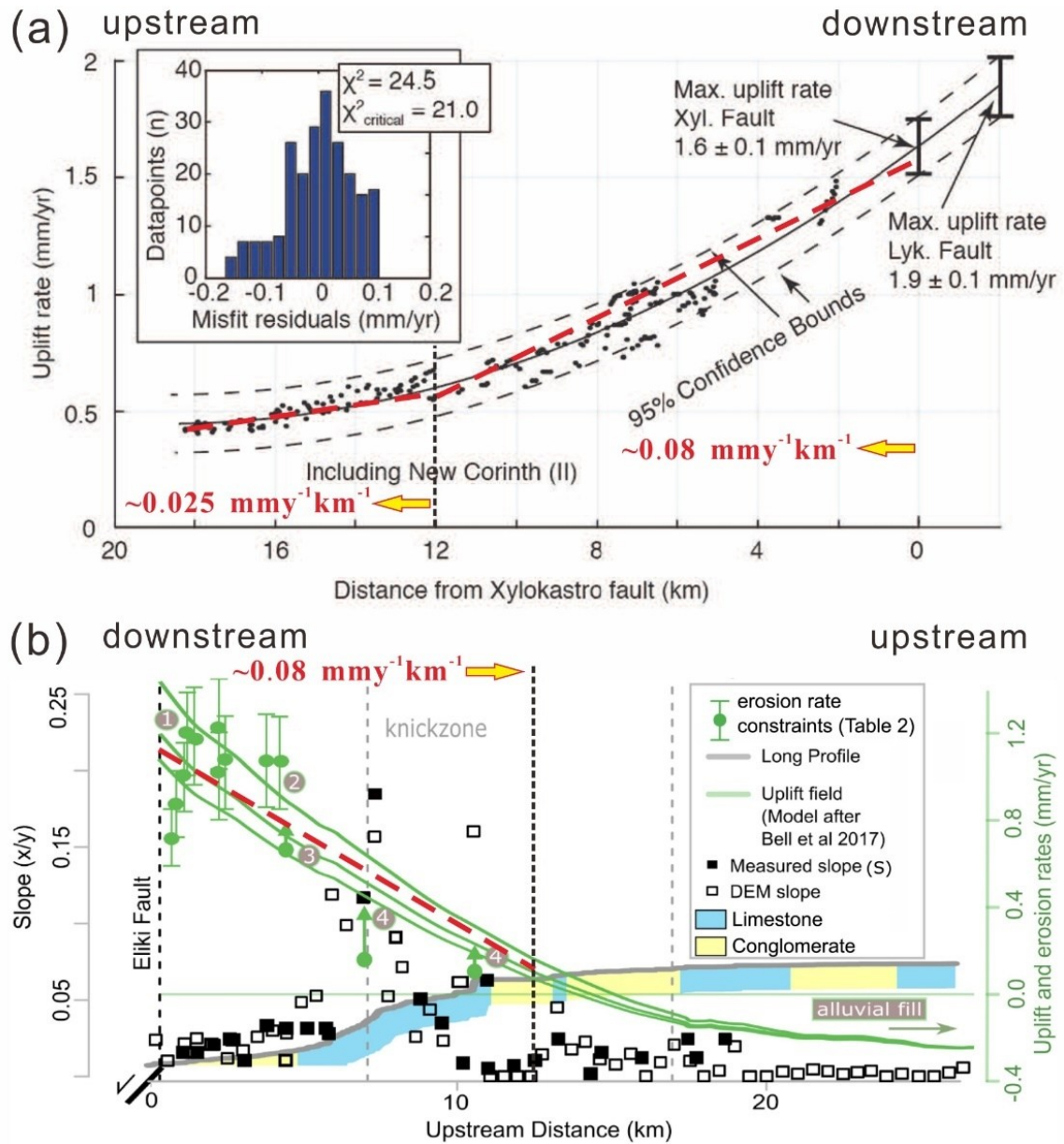


Figure 2.S1. Uplift rate variations across fault strike for the Sithas/SI (a) and the Vouraikos/VO (b) (modified from [de Gelder et al., 2019](#) and [Zondervan et al., 2020b](#)). Note the linear decay model with a rate of $\sim 0.08 \text{ mmyr}^{-1}\text{km}^{-1}$ (dashed red lines) is mostly effective within the range of $\sim 12 \text{ km}$ upstream of the fault, beyond which the decay rate decreases significantly. In the SI, the fulcrum position is likely located $>30 \text{ km}$ upstream of the fault using the decay rate of $\sim 0.025 \text{ mmyr}^{-1}\text{km}^{-1}$; In the VO, the potential fulcrum position is likely located $>25 \text{ km}$ upstream of the fault where alluvial fill appears (see their Figure 2 in [Zondervan et al., 2020b](#)).

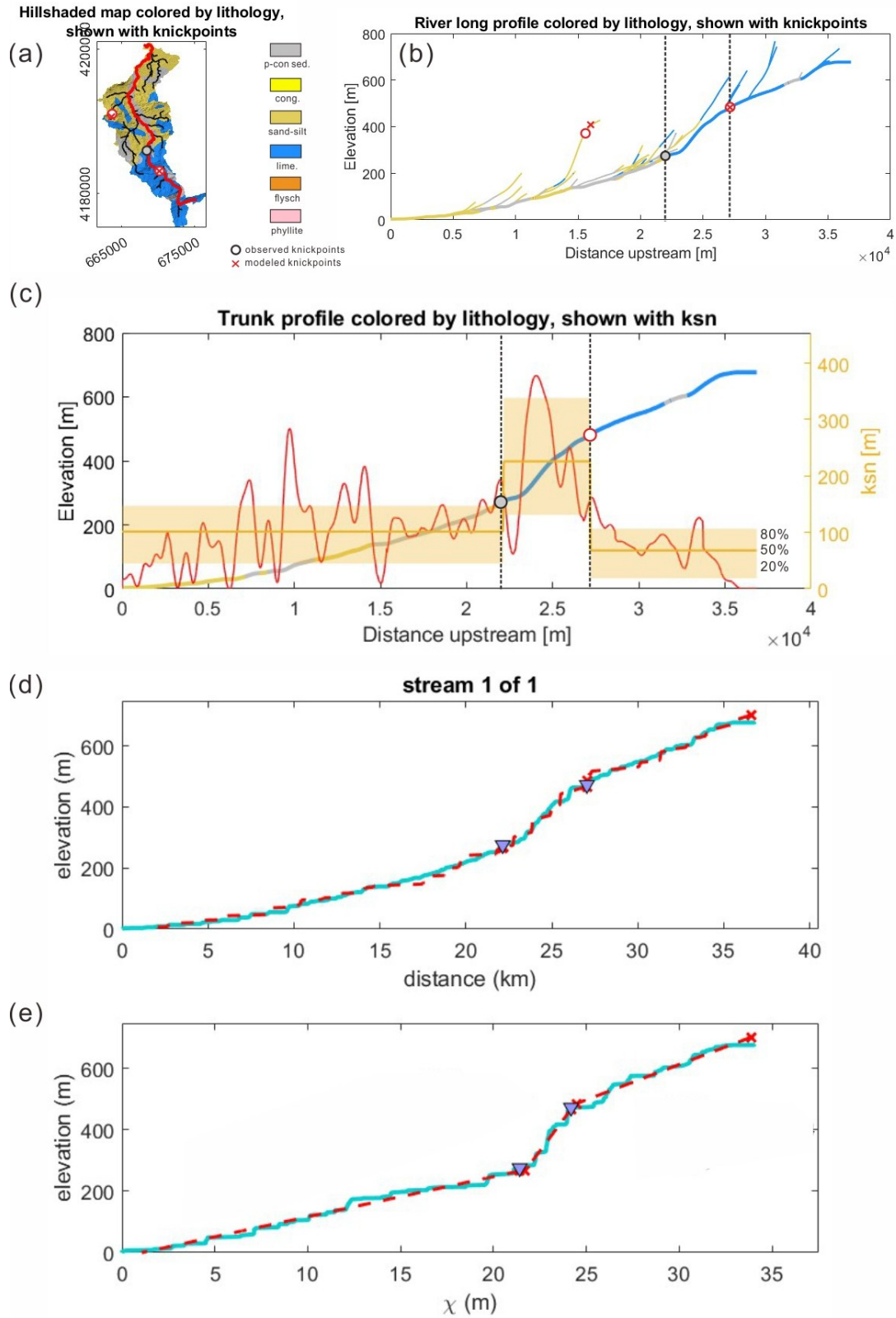


Figure 2.S2. Knickpoints in the XE catchment, shown on the hillshaded map (a), and river profile (b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

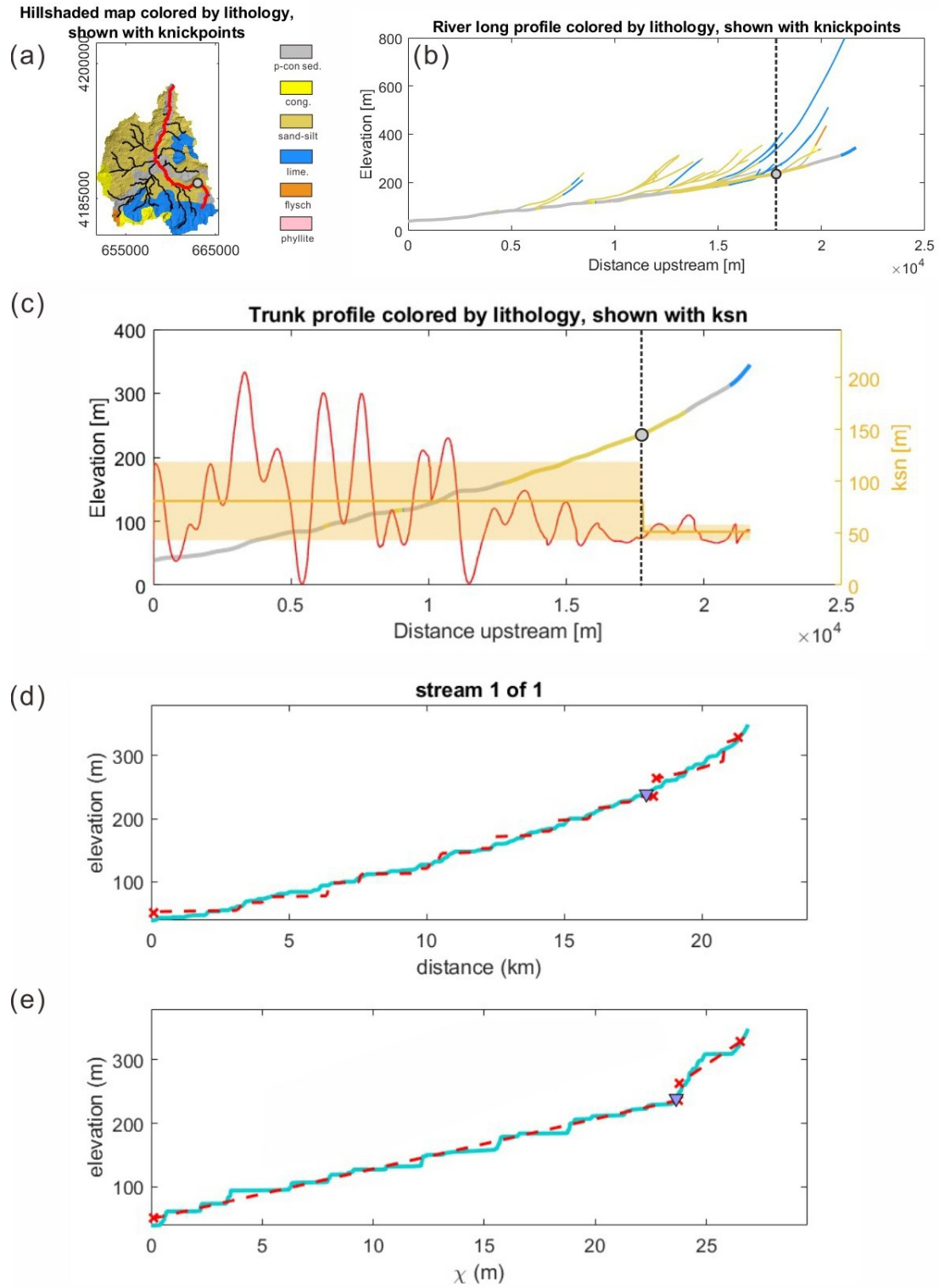


Figure 2.S3. Knickpoints in the RA catchment, shown on the hillshaded map (a), and river profile (b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same color are from the same stage.

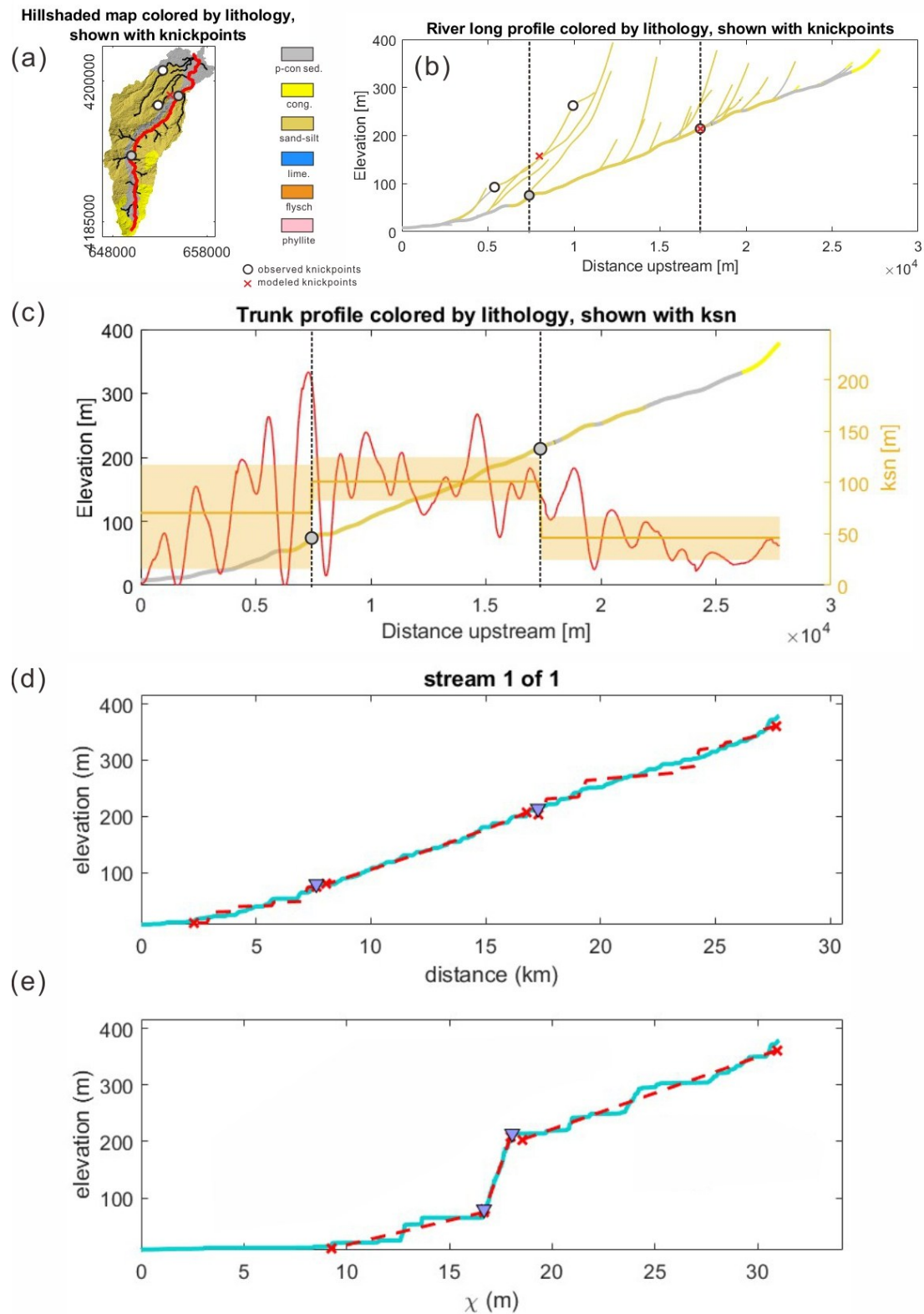


Figure 2.54. Knickpoints in the ZA catchment, shown on the hillshaded map (a), and river profile (b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

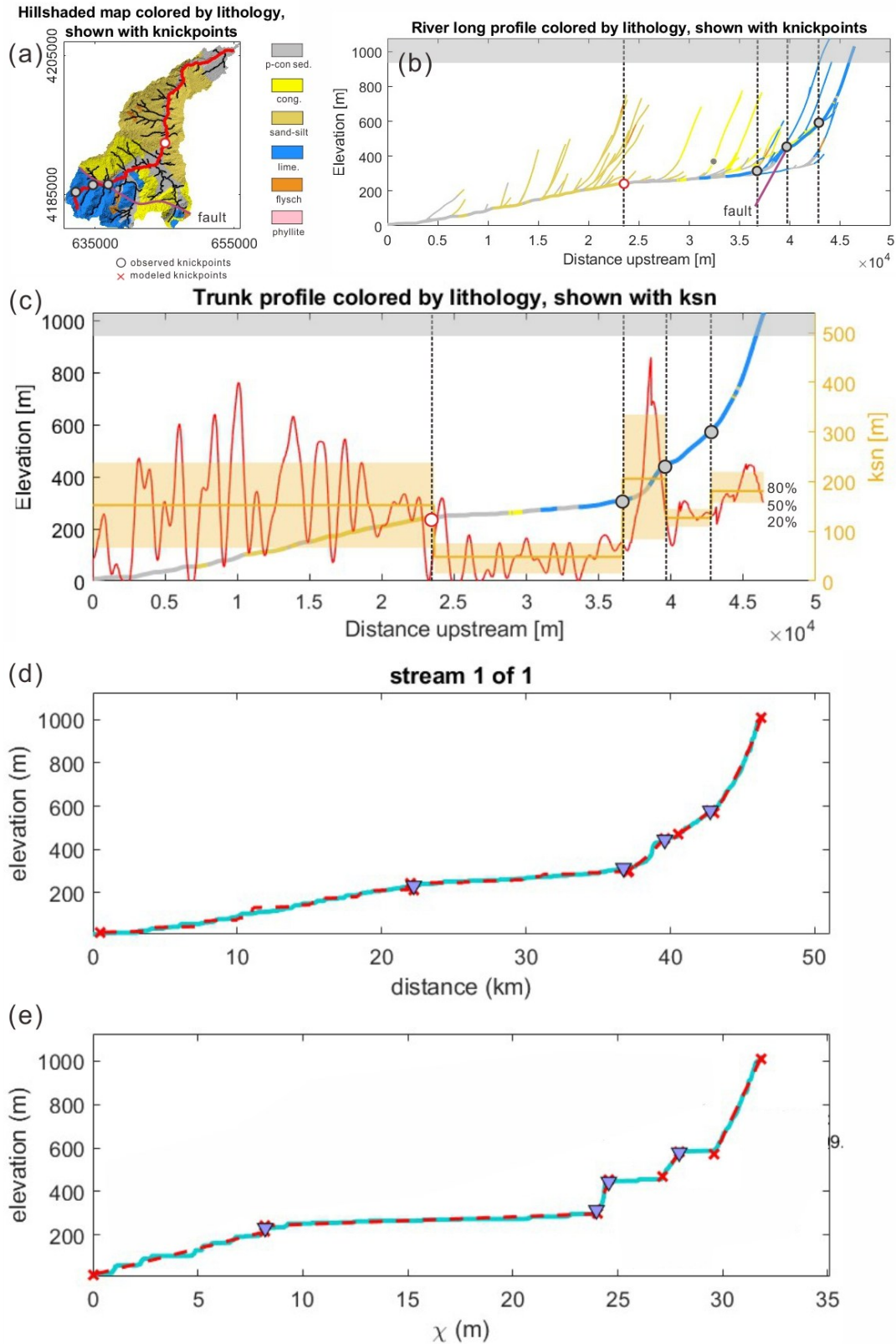


Figure 2.S5. Knickpoints in the AS catchment, shown on the hillshaded map (a), and river profile (b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

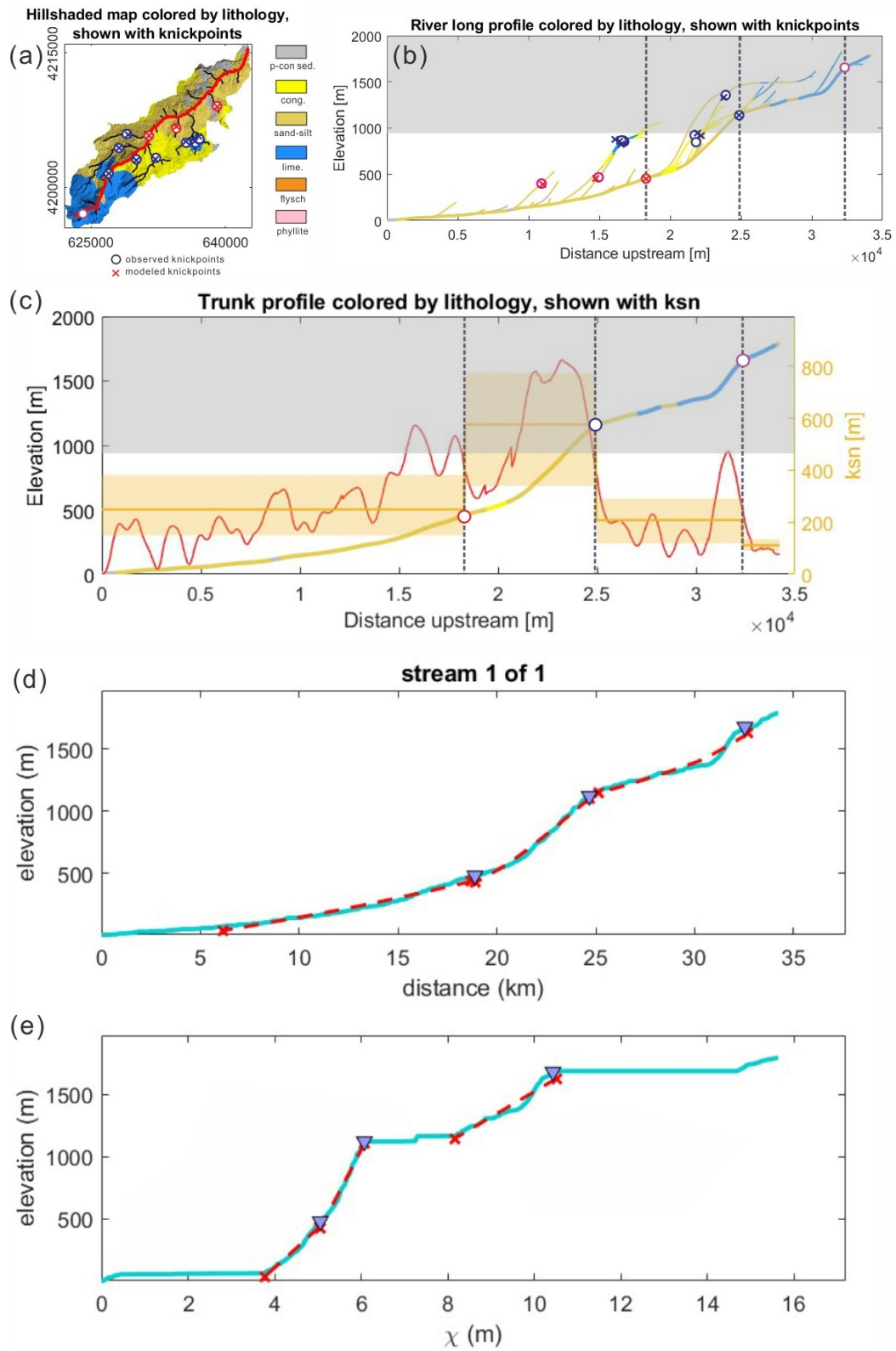


Figure 2.S6. Knickpoints in the SI catchment, shown on the hillshaded map (a), and river profile (b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

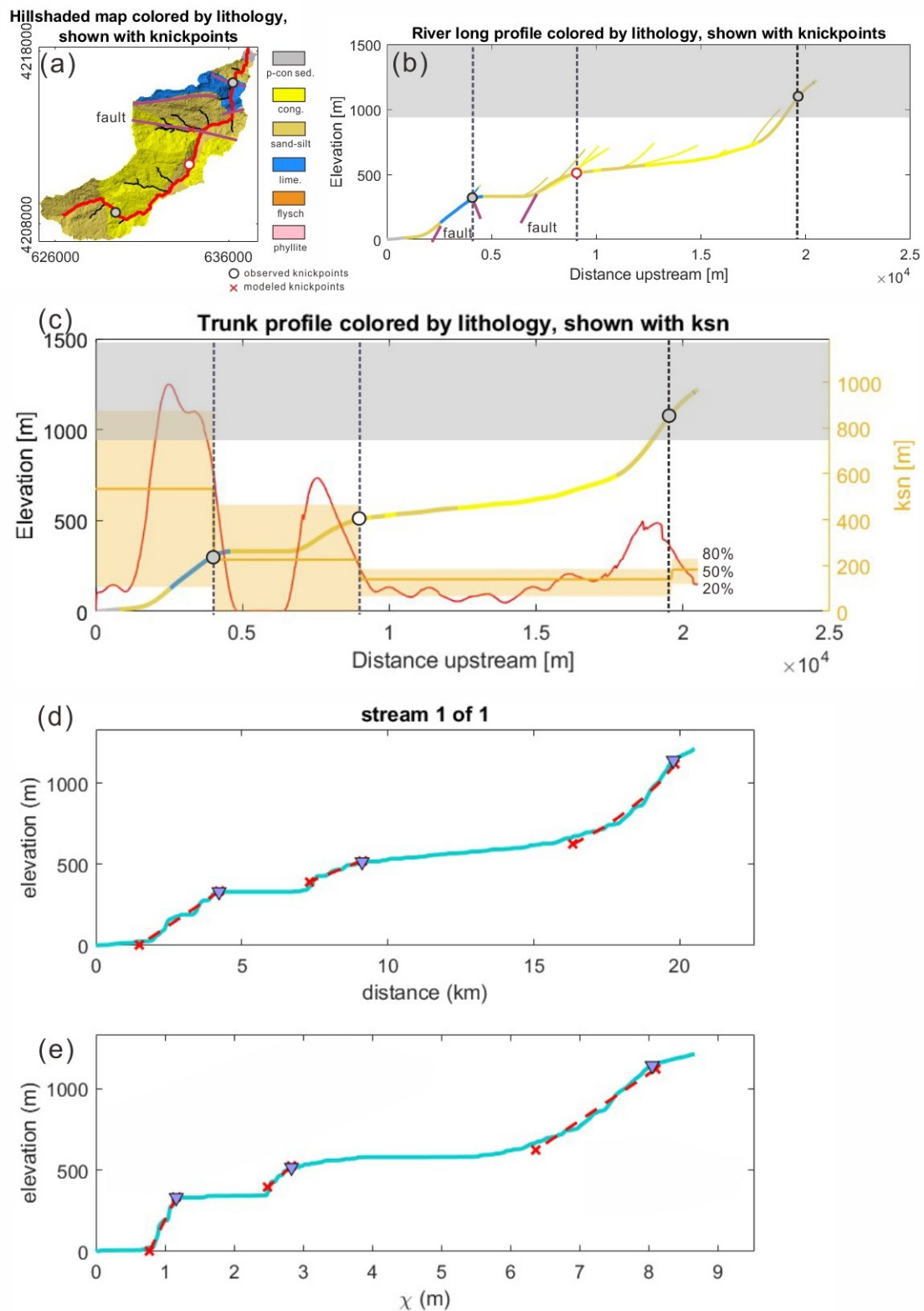


Figure 2.S7. Knickpoints in the FO catchment, shown on the hillshaded map (a), and river profile (b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

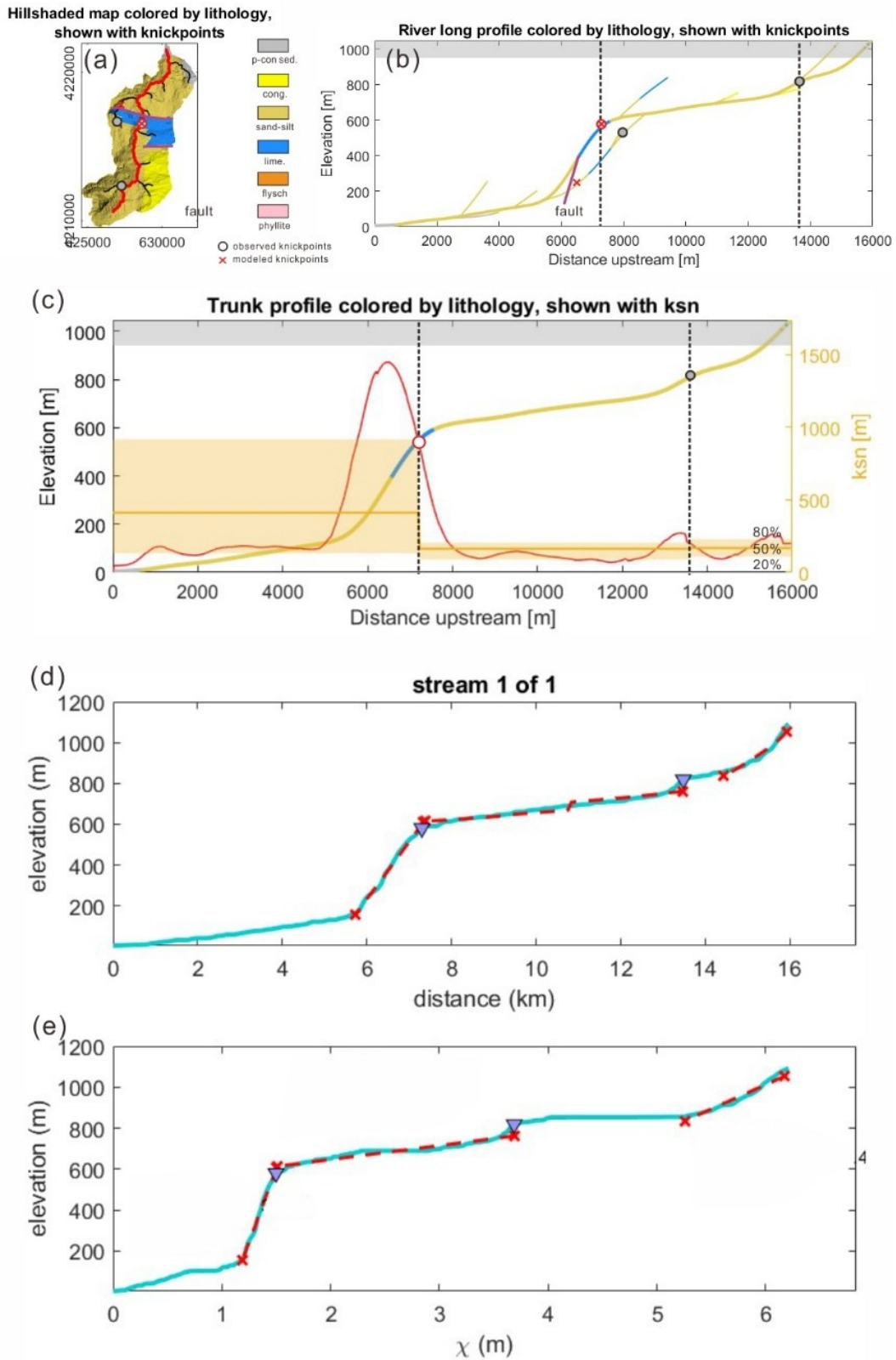


Figure 2.S8. Knickpoints in the SK catchment, shown on the hillshaded map (a), and river profile (b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

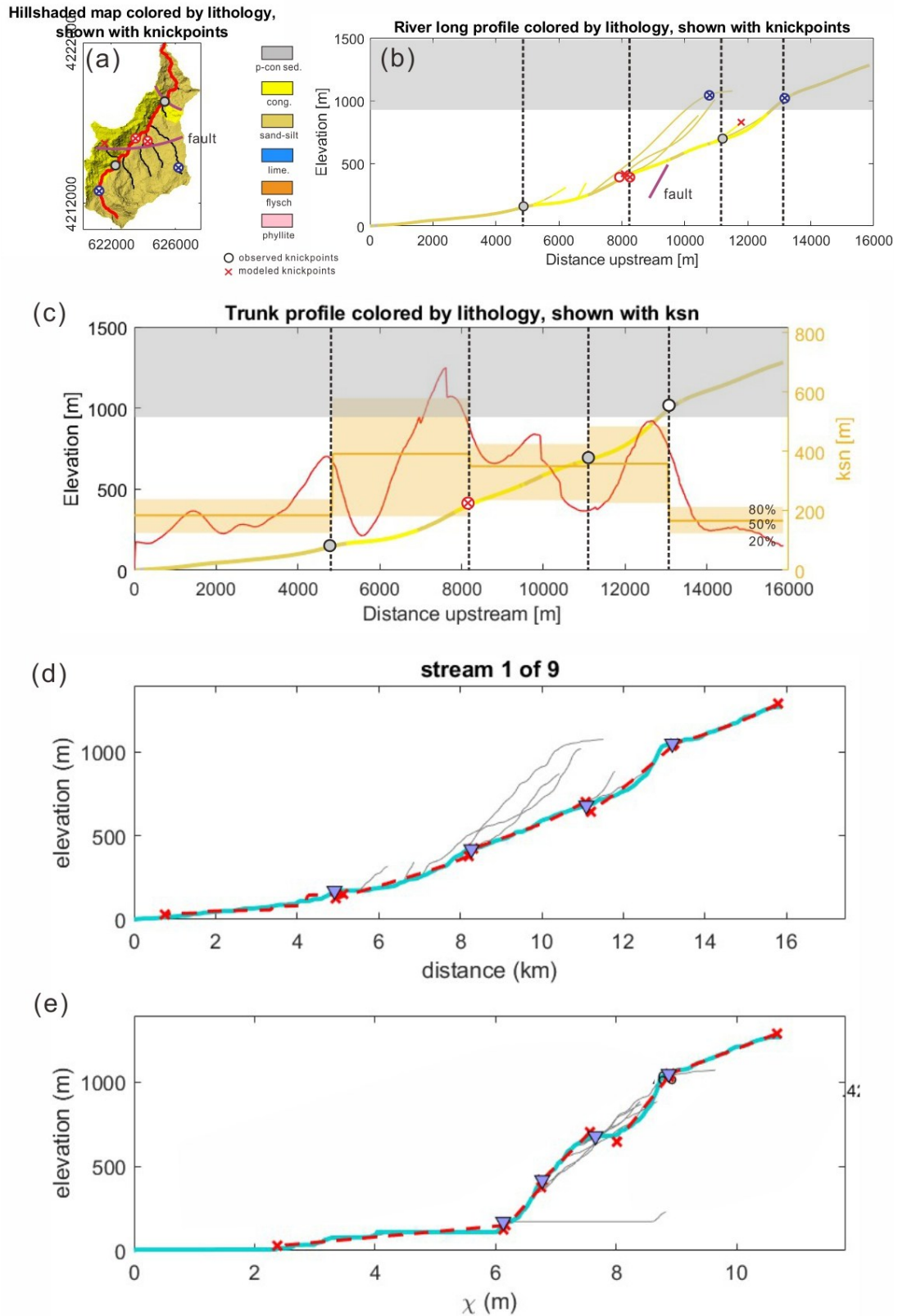


Figure 2.S9. Knickpoints in the DE catchment, shown on the hillshaded map (a), and river profile (b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

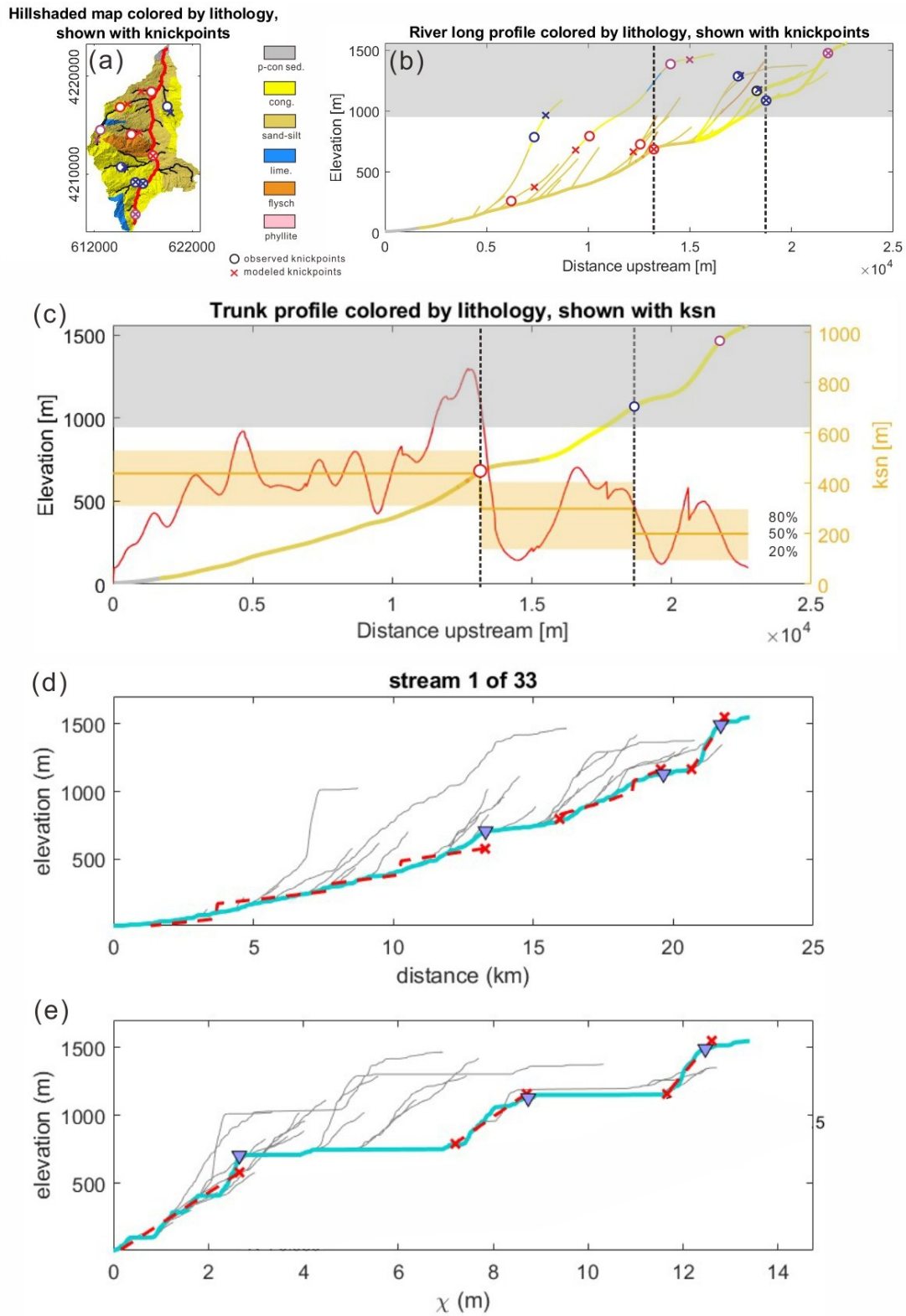


Figure 2.S10. Knickpoints in the SK catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

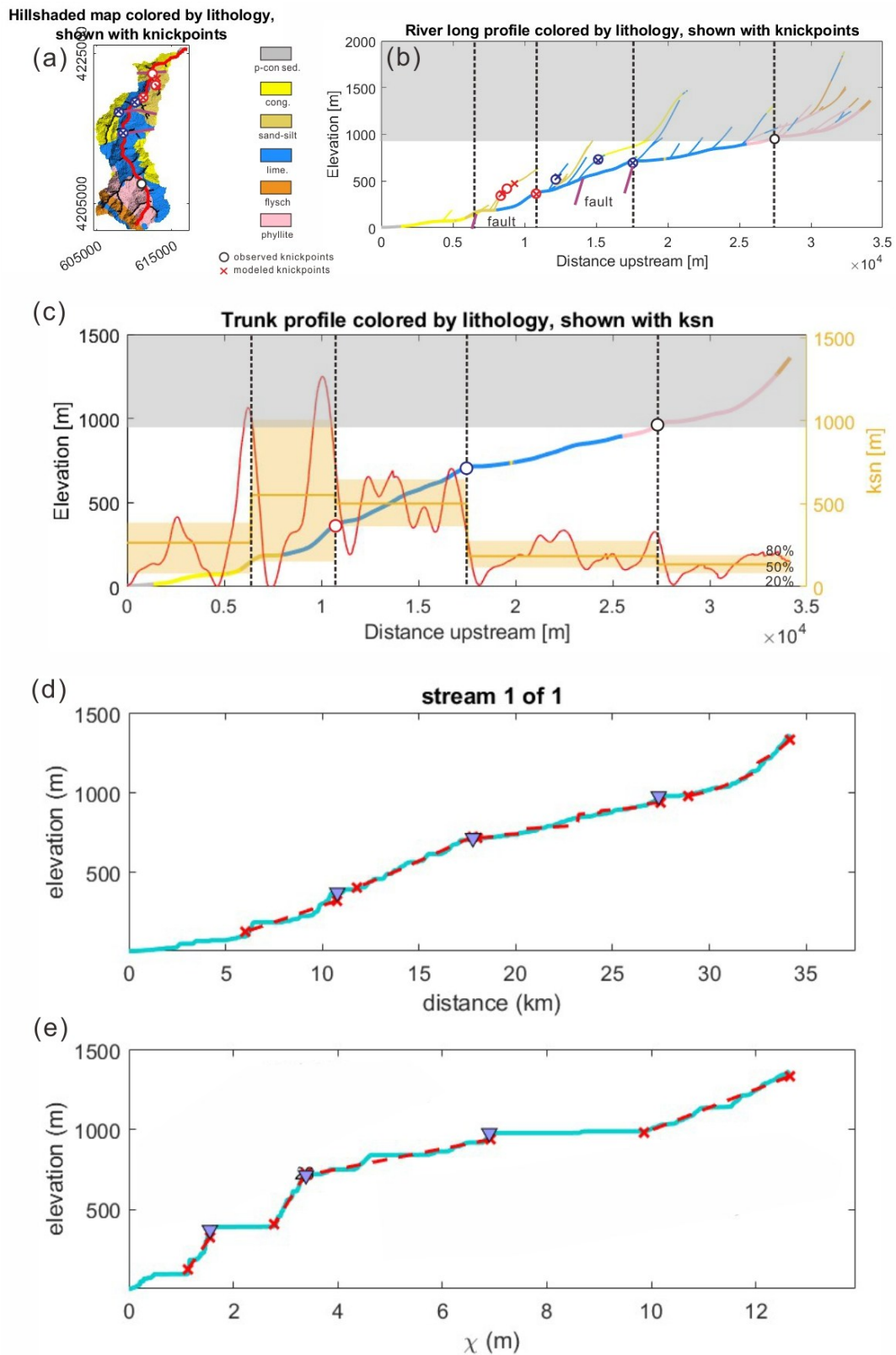


Figure 2.S11. Knickpoints in the KR catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

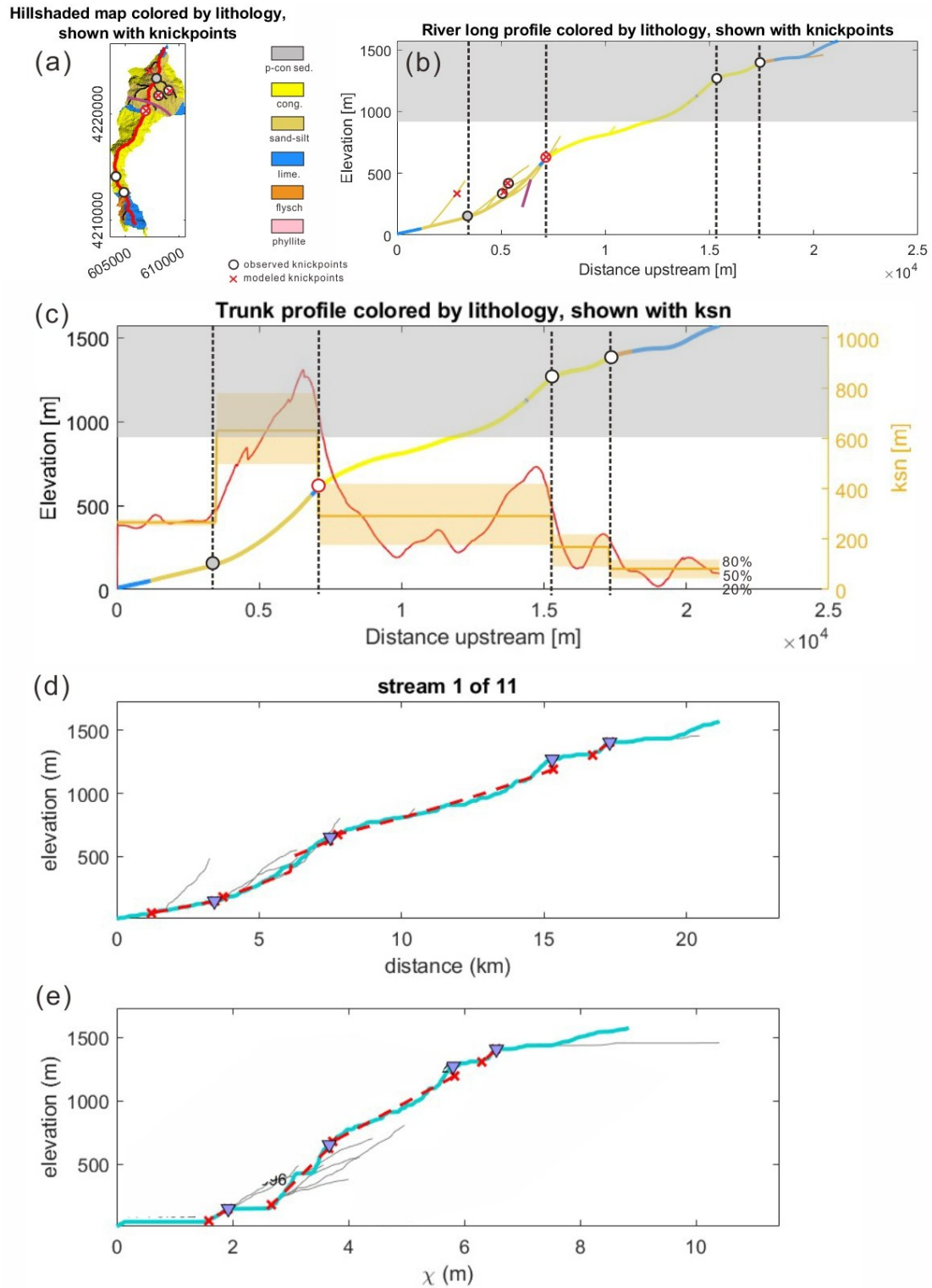


Figure 2.S12. Knickpoints in the LA catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

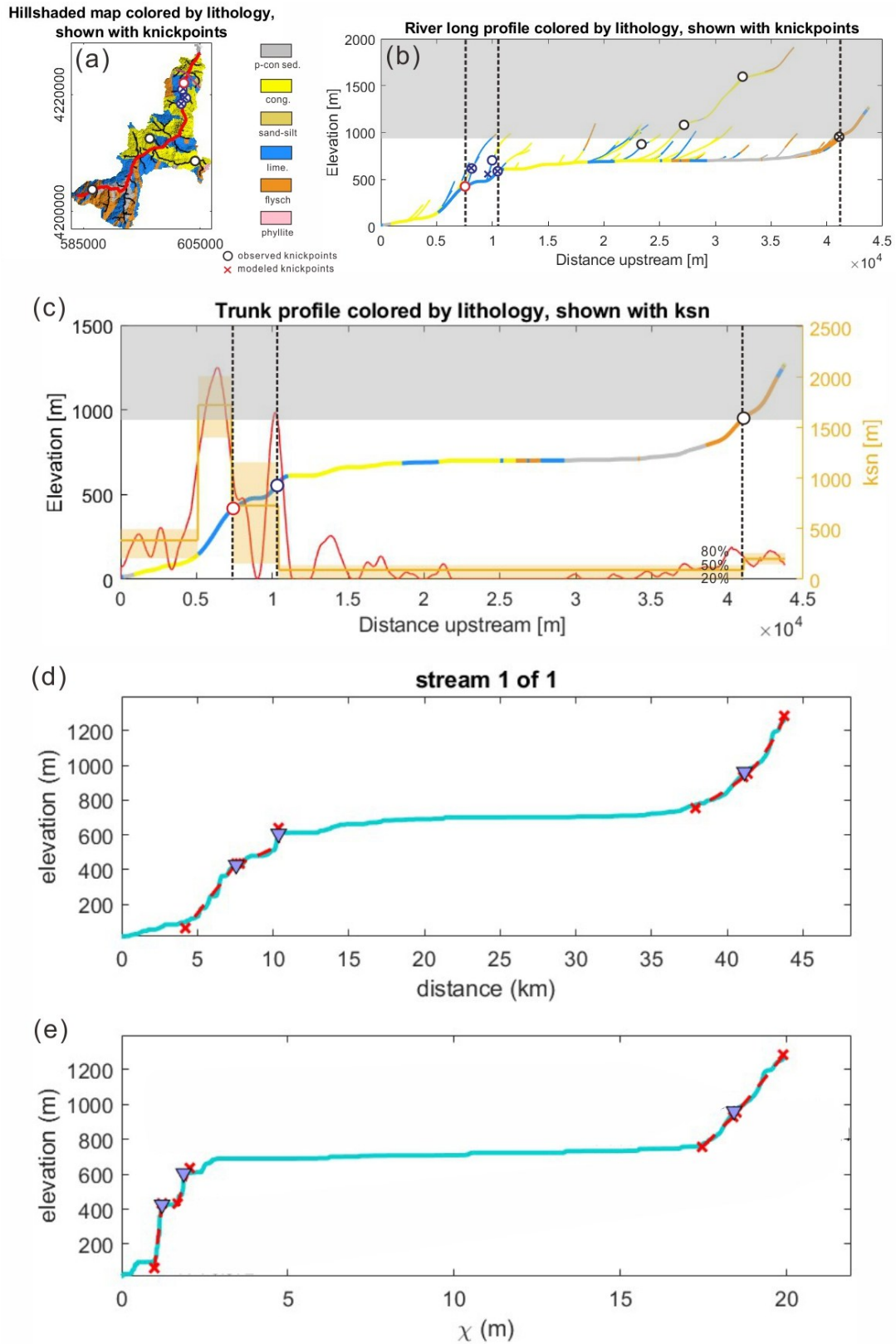


Figure 2.S13. Knickpoints in the VO catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

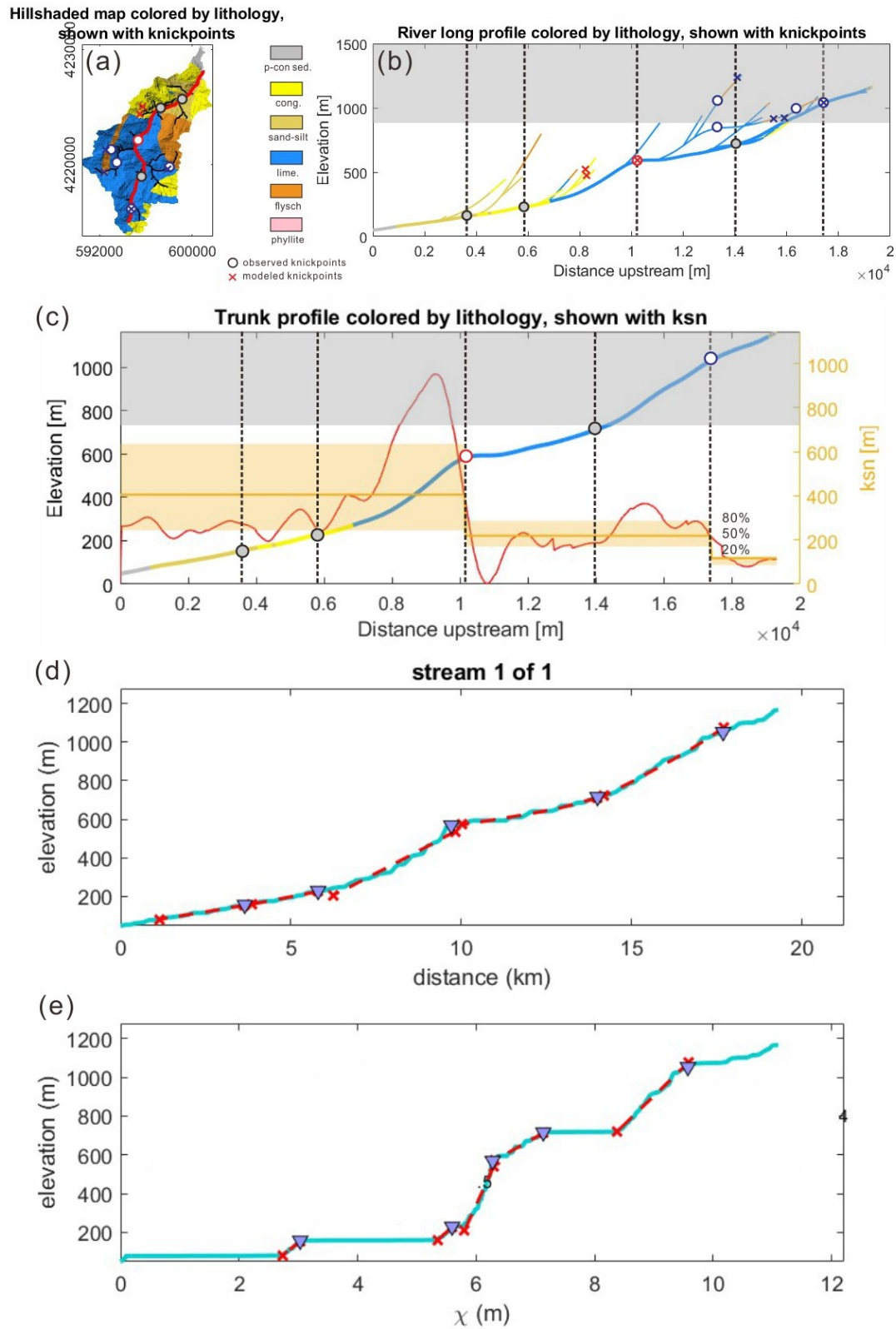


Figure 2.S14. Knickpoints in the KE catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

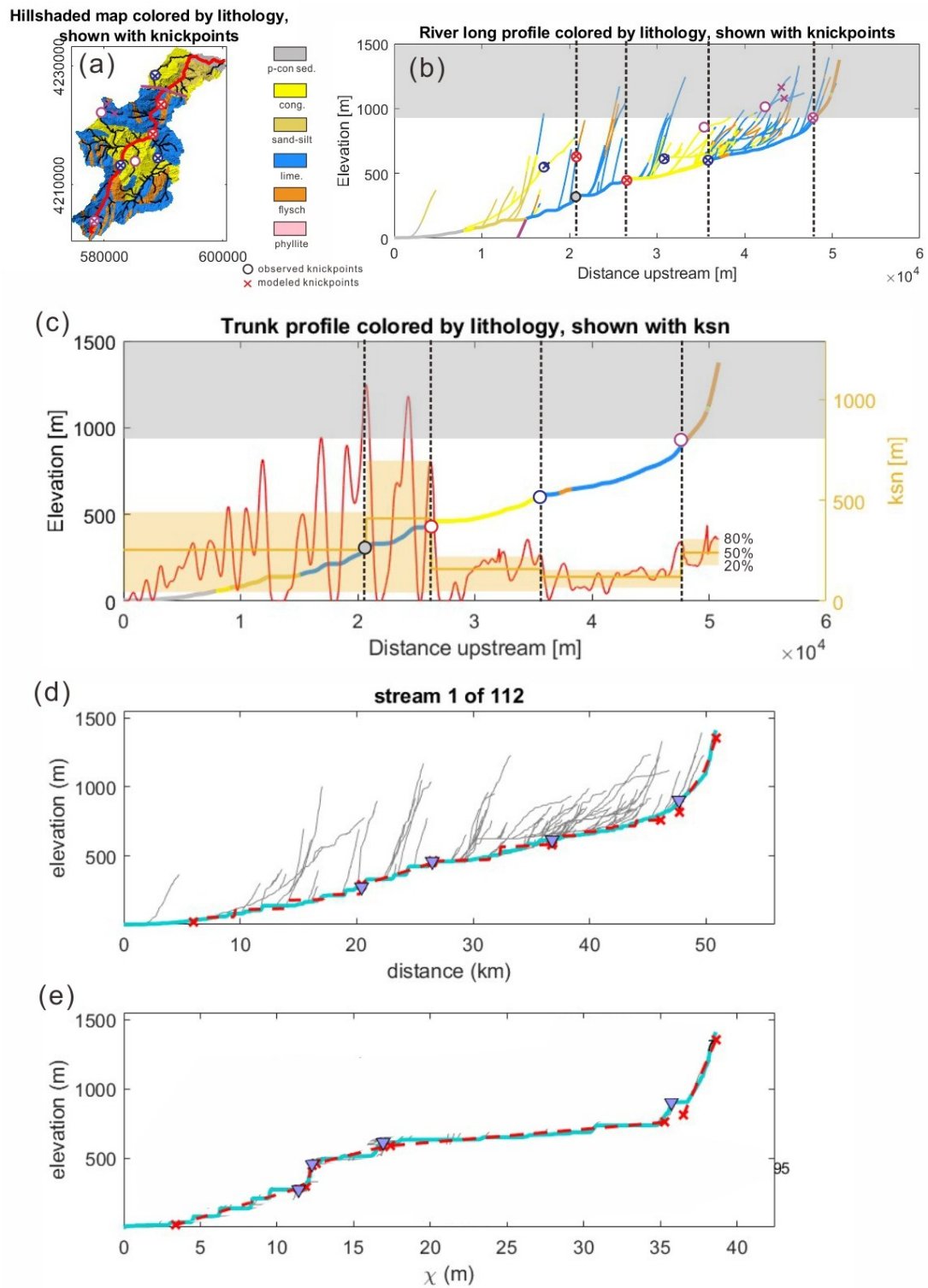


Figure 2.S15. Knickpoints in the SE catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

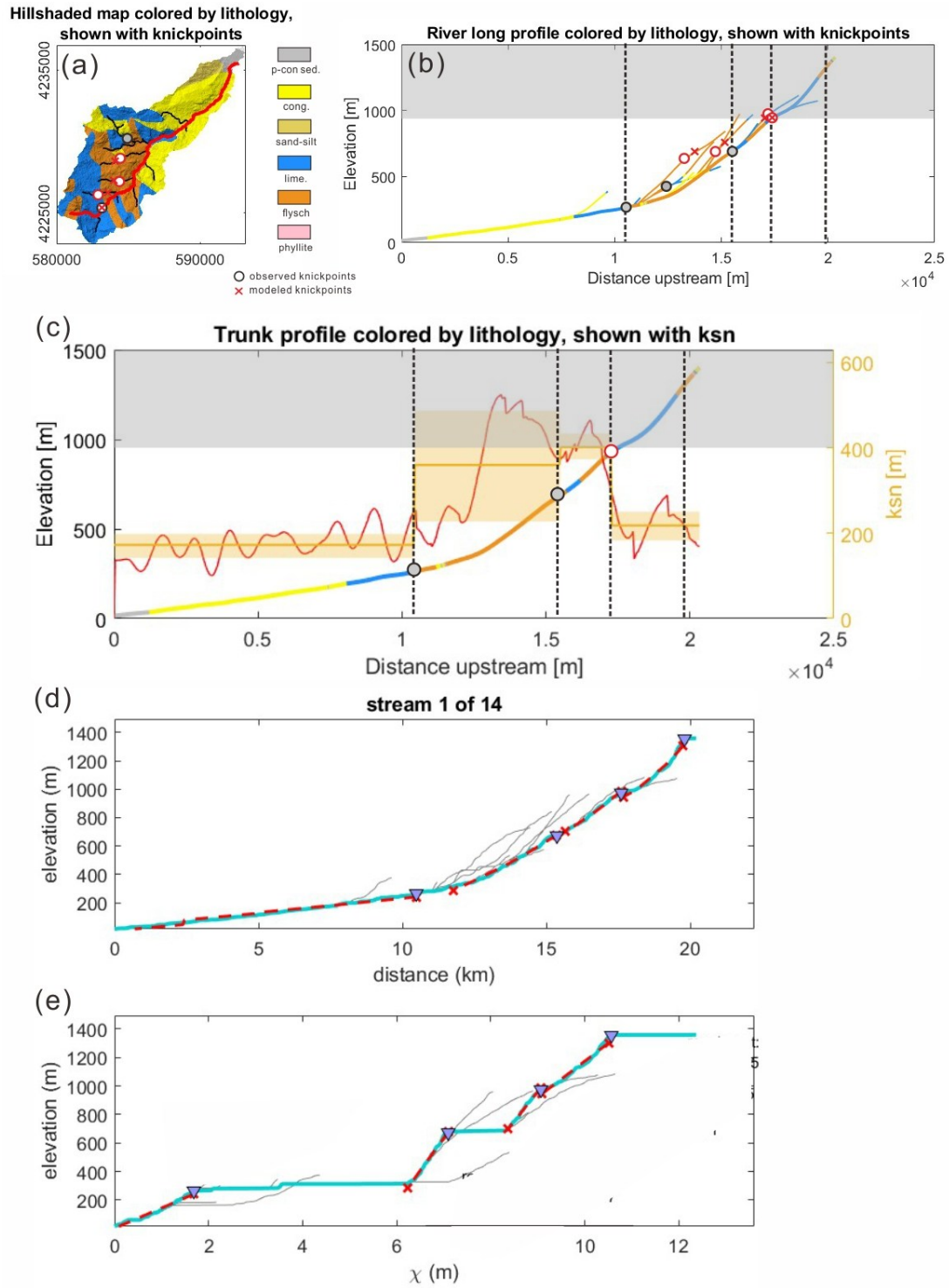


Figure 2.S16. Knickpoints in the ME catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

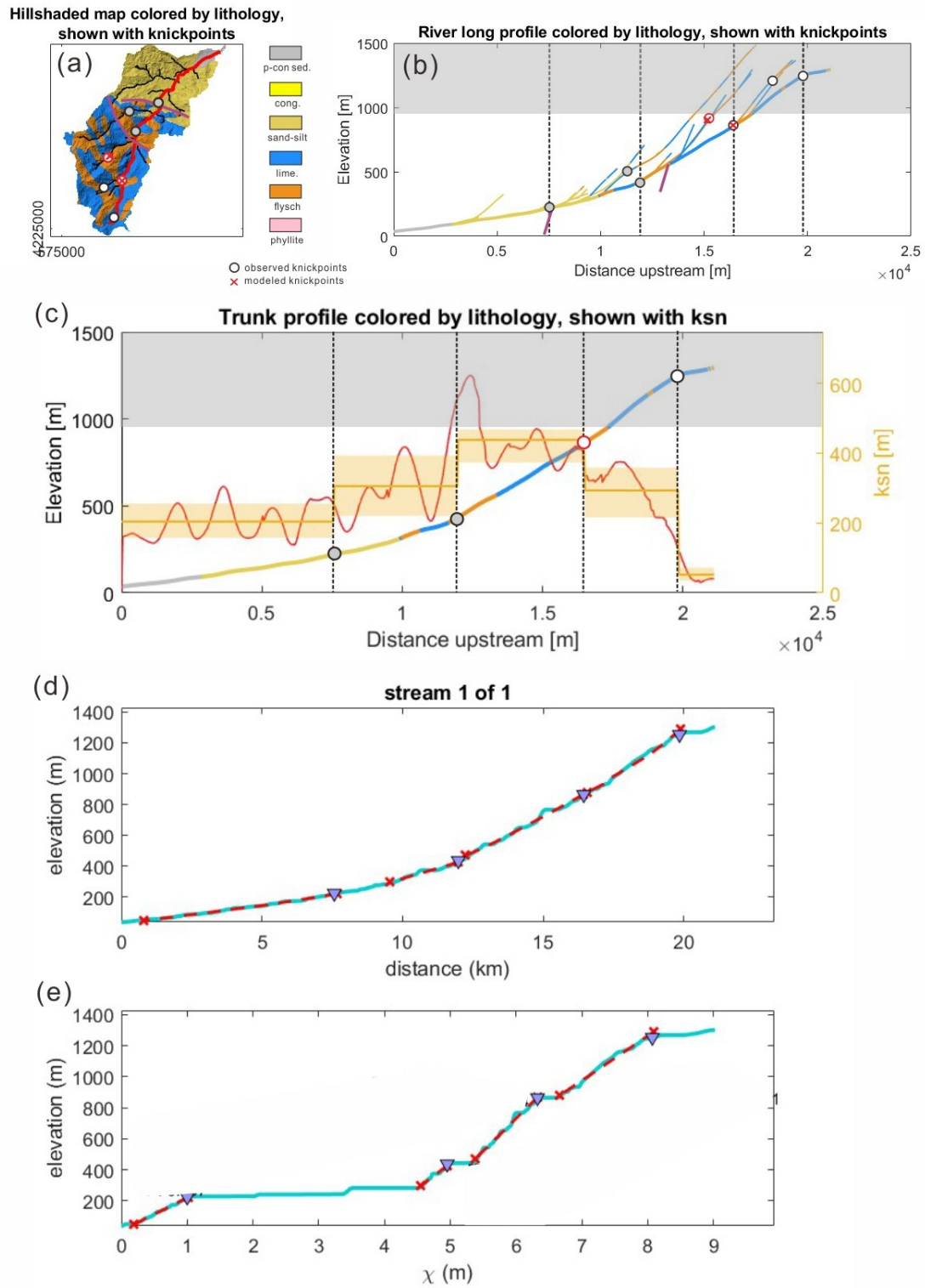


Figure 2.S17. Knickpoints in the PH catchment, shown on the hillshaded map (a), and river profile(b), are identified from (c) the k_{sn} map, (d) river profile and (e) the χ profile. Note the circles (observed) and crosses (modelled) with the same colour are from the same stage.

3 DECODING TECTONIC AND LITHOLOGICAL CONTROLS ON SEDIMENT EXPORTED FROM FOOTWALL CATCHMENTS: IMPLICATIONS FOR HANGINGWALL STRATIGRAPHY ¹

Abstract

Hillslope and river channel processes combine to erode topography and release sediments in response to external forcing, such as tectonics. Any forcing that drives these processes to shape topography likely also controls the resulting sediment characteristics; however, the extent to which the forcing translates into exported sediment grain size remains elusive, especially where tectonic and lithological forcings coexist. Here, we focus on two catchments with these characteristics in the Gulf of Corinth, Greece, to probe how tectonics and lithology compete to control exported sediment grain size via hillslope and channel processes. Topographic analysis reveals that both tectonic and lithological forcings can drive hillslope and channel processes to shape topography. In the Sithas catchment, normal faulting >600 ka initiates knickpoints that steepen hillslopes and trigger concentrated mass wasting downstream of knickpoints. In the Phoenix catchment, which is perturbed only by active faulting for <100 ka, the landslide-prone bedrock triggers pervasive mass wasting and steepens hillslopes. Combined with the observation that mass wasting produces coarser grains, our data therefore show that both tectonic and lithological forcings can be expressed as coarsened sediment on hillslopes. If hillslope sediment is friable, however, coarsened sediment from hillslopes can be rapidly abraded upon entering river channels and lose its coarse signal before reaching catchment outlets (e.g. tectonically induced coarsening in the Sithas catchment). In contrast, coarsened sediment from hillslopes can preserve its coarse signal to catchment outlets if its lithology is resistant to abrasion (e.g. lithologically induced coarsening in the Phoenix catchment). En route to catchment outlets, sediments coarser than the threshold for entrainment are filtered out; however, this selective transport attenuates rather than obliterates the coarse signal as coarse sediment input itself can increase the entrainment threshold by influencing channel steepness. Finally, we explore how continuous unroofing and knickpoint propagation may cause grain size variations in exported sediments and impact hangingwall stratigraphy. Consequently, our study illustrates lithology can outweigh tectonics in determining the characteristics of sediment exported from footwall catchments and paves the way for more comprehensive stratigraphic interpretation.

¹ A version of this chapter has been published as the following paper:

Zhou, Z., Whittaker, A.C., Bell, R.E., Hampson, G.J., Chandresh, R., Watkins, S.E., Zondervan, J.R., 2026. Decoding tectonic and lithological controls on sediment exported from footwall catchments: Implications for hangingwall stratigraphy. *Geological Society of America Bulletin*. doi: [10.1130/B38399.1](https://doi.org/10.1130/B38399.1)

3.1 Introduction

Hillslope and river channel processes combine to erode topography and release sediment in response to external forcing (e.g. Egholm et al., 2013; Riebe et al., 2015; Sklar et al., 2024). Any forcing that shapes topography, such as tectonics, likely also controls sediment grain size (e.g. D’Arcy et al., 2017; Neely and DiBiase, 2020; Whittaker et al., 2010). The anticipated link between the forcing and its resultant grain size, however, is mediated by hillslope and channel processes (e.g. Armitage et al., 2013; Castelltort and Van Den Driessche, 2003; Jerolmack and Paola, 2010). These geomorphic processes, including non-linear landsliding and selective transport, can both establish and disrupt this anticipated relationship (c.f. Whittaker et al., 2010; Litty et al., 2017). The extent to which this link is established or disrupted complicates stratigraphic interpretations. If it does not hold, the use of preserved sediment grain size to reconstruct the external forcing is called into question (e.g. Foreman et al., 2012; Peizhen et al., 2001; Schlunegger and Norton, 2015). Conversely, if this link is maintained, a decrease in exported sediment grain size could explain an upward-fining stratigraphy that might otherwise be attributed to increased accommodation (e.g. Allen et al., 2015; Armitage et al., 2011). It is thus of fundamental importance to assess how effectively external forcing is translated into exported sediment grain size.

One approach to probe this question is to track sediment generation and transport directly in upland catchments via a process-based analysis (Allen et al., 2015; Sklar et al., 2017). This is well-illustrated in some studies that examine the causal relation between sediment grain size and normal faulting (Attal et al., 2015; Whittaker et al., 2010). The causal chain begins with the initiation of migrating knickpoints that rejuvenate their downstream landscapes in response to changes in uplift rates or base-level falls (e.g. Reinhardt et al., 2007; Whittaker et al., 2007). In the rejuvenated landscapes, steepened hillslopes then shorten the weathering time for hillslope sediments (Attal et al., 2015; Marshall and Sklar, 2012) and shift the dominant sediment generation process from weathering to mass-wasting (e.g. Whittaker et al., 2010), thereby contributing coarser sediments to channels (Sklar et al., 2017). Moreover, steepened channels facilitate the export of coarser sediments to catchment outlets due to their increased ability to entrain coarser grains and decreased potential to deposit (e.g. Attal et al., 2015; Bekaddour et al., 2013). Yet, the correlation between uplift rate and exported sediment grain size is typically exemplified with respect to catchments underlain by a homogeneous and hard lithology (Attal et al., 2015; Whittaker et al., 2010). In recent years, bedrock lithology has also been demonstrated to drive landscape evolution (e.g. Carr et al., 2023; DiBiase et al., 2018; Zondervan et al., 2020) and control sediment grain size (e.g. Attal and Lavé, 2006; Dingle et al., 2017; Roda-Boluda et al., 2018). The response of exported sediment grain size to increases in uplift rates

(tectonic perturbations) is likely to be modulated by the impact of lithology, but this hypothesis has been poorly explored to date.

In the Gulf of Corinth, Greece, [Watkins et al. \(2020\)](#) compared sediment grain size at multiple river outlets with the uplift rates experienced by each river and constituent lithological proportions in its catchment ($N = 47$). Their results indicated that exported sediment grain size exhibits a stronger relationship with the proportion of hard lithology than with uplift rate, highlighting that lithology might outweigh tectonics in controlling sediment grain size. Unfortunately, the authors had few direct constraints on sediment sources and were unable to explicitly examine the mechanisms by which the correlations occur. In contrast, [Attal et al. \(2006\)](#) sampled both hillslope and channel sediments in the Marsyandi river (Nepal) and found lithology-dependent abrasion accounts for downstream variations in lithological proportions; however, their work did not constrain tectonic impacts on topographic erosion and resultant sediment grain size. Thus, field areas with well-constrained tectonic and lithological frameworks as well as systematic sediment grain size sampling are still needed.

In this study, we set out to investigate how tectonics and lithology compete to control exported sediment grain size over 100-ka timescales in the Gulf of Corinth, where both normal faulting and lithological heterogeneity are well constrained ([Gawthorpe et al., 2018](#); [Watkins et al., 2020](#); [Zhou et al., 2024](#)). For example, tectonic knickpoints and the impact of lithologically heterogeneous bedrock have been examined to reconstruct fault growth and linkage; however, their full impact on hillslope and channel dynamics is yet to be illustrated ([Zhou et al., 2024](#)). Here, we first combine topographic analysis with mass-wasting mapping to assess the relative impact of tectonics and lithology in driving geomorphic processes. In this context, we track grain size variations from hillslopes to tributaries and catchment outlets to evaluate the extent to which the dominant forcing propagates across hillslope and channel processes and into exported sediment grain size. The results are extrapolated to provide implications for hangingwall stratigraphy from a long-term perspective.

3.2 Study area

The Sithas and Phoenix rivers, located in the Gulf of Corinth, Greece, are bounded by active normal faults ([Figure 3.1a](#); [Bell et al., 2009](#); [Gawthorpe et al., 2018](#); [Zhou et al., 2024](#)). Fault linkage and the activation of new faults have increased slip rate and, consequently, footwall uplift rate in the region ([Ford et al., 2017](#); [Gawthorpe et al., 2018](#); [McNeill et al., 2007](#); [Nixon et al., 2016](#); [Nixon et al., 2024](#)). For example, the Neo Erineos fault segments, cutting across the Phoenix river was likely activated within the last 100 ka, with long-term footwall uplift rates of 1–2 mm/yr ([Palyvos et al., 2010](#); [McNeill et al., 2005](#); [Table 3.1](#)). The offshore East Xylokaastro fault bounding the Sithas river, however, reached its uplift rates of 1.3–1.6 mm/yr following its linkage with adjacent faults around 600 ka ago ([Armijo et](#)

al., 1996; de Gelder et al., 2019; Zhou et al., 2024; Table 3.1). Normal faulting has generated both microseismic swarms (e.g. Kaviris et al., 2021) and frequent $M > 6$ earthquakes, including a recent event near the Phoenix river in 1995 (e.g. Bernard et al., 1997; Nixon et al., 2024). In the short term, these earthquakes could trigger mass wasting events on hillslopes and increase sediment concentration in rivers (e.g. Yanites et al., 2010). In the long term, repeated earthquakes across multiple cycles can increase uplift rates, developing upstream-migrating knickpoints in footwall rivers (Demoulin et al., 2015; Fernández-Blanco et al., 2019; Zhou et al., 2024). Importantly, knickpoints from the same tectonic perturbation characterise an upstream-migrating incisional wave (Whittaker, 2012), downstream of which mass wasting is enhanced in the rejuvenated landscape (Bigi et al., 2006; Gallen et al., 2011). In addition, back-tilting of fault blocks causes footwall uplift and hangingwall subsidence rates to systematically decline with distance from the fault, as evidenced by marine terraces (e.g. Armijo et al., 1996), landscape characteristics (e.g. Bell et al., 2018; Fernández-Blanco et al., 2020; Pechlivanidou et al., 2019), and seismic sections (e.g. de Gelder et al., 2019). In particular, a linear decay model with a rate of $\sim 0.08 \text{ mmyr}^{-1}\text{km}^{-1}$ has been found to broadly reproduce the strike-perpendicular footwall uplift rate variations, supporting its use for other active faults in the Corinth rift (Zhou et al., 2024). As a result, the Phoenix and Sithas rivers on the footwall experience increasing uplift rates from upstream to downstream.

Both rivers incise through heterogeneous bedrock, from pre-rift basement upstream to syn-rift strata downstream (Figure 3.1b, Bornovas and Rondogianni-Tsiambaou, 1983). The pre-rift basement comprises mainly limestone, sandy turbidite flysch, and phyllites-quartzites (Gawthorpe et al., 2018; Papanikolaou and Lekkas, 2008; Rohais and Moretti, 2017). The syn-rift stratigraphy includes proximal alluvial and deltaic conglomerates, distal sandstone and siltstone, as well as more recent poorly consolidated sediment (Figure 3.1b, Rohais et al., 2007; Ford et al., 2013). These geological units, distinguished by age and lithology, represent first-order bedrock groupings with distinct rock strengths (Figure 3.1b, e.g. Zhou et al., 2024); however, some variations are expected in these broad groupings. For example, in areas where sandstone and siltstone dominate, occasional conglomerate beds deposited by gravity flows may occur (e.g. Muravchik et al., 2019; Gawthorpe et al., 2018). Moreover, the limestone and flysch of the Pindos unit in western Corinth (e.g. the Phoenix catchment) are thinner-bedded and locally more deformed than their counterparts of the Gavrovo-Tripolis unit in eastern Corinth (e.g. the Sithas catchment) (Wicker et al., 2024). This could cause the Pindos limestone and flysch to exhibit lower rock strength and be more prone to mass wasting than their Gavrovo-Tripolis counterparts (e.g. Gallen et al., 2015). This is especially true given that the Pindos limestone contains interbedded chert layers (Skourlis and Doutsos, 2003) that could further reduce aggregate rock strength. In areas surrounding the Phoenix catchment, Kavoura and Sabatakakis (2020), based on

an inventory of 209 landslide events recorded between 1913 and 2015, found that slope and lithology outweigh precipitation and elevation in governing landslide susceptibility. Their work suggests that tectonics, through its control on slope (Zhou et al., 2024), and lithology are fundamental drivers of mass wasting, although their influence on individual mass-wasting events is likely mediated by local factors such as pore fluid pressure, vegetation cover and type, and soil development and structure.

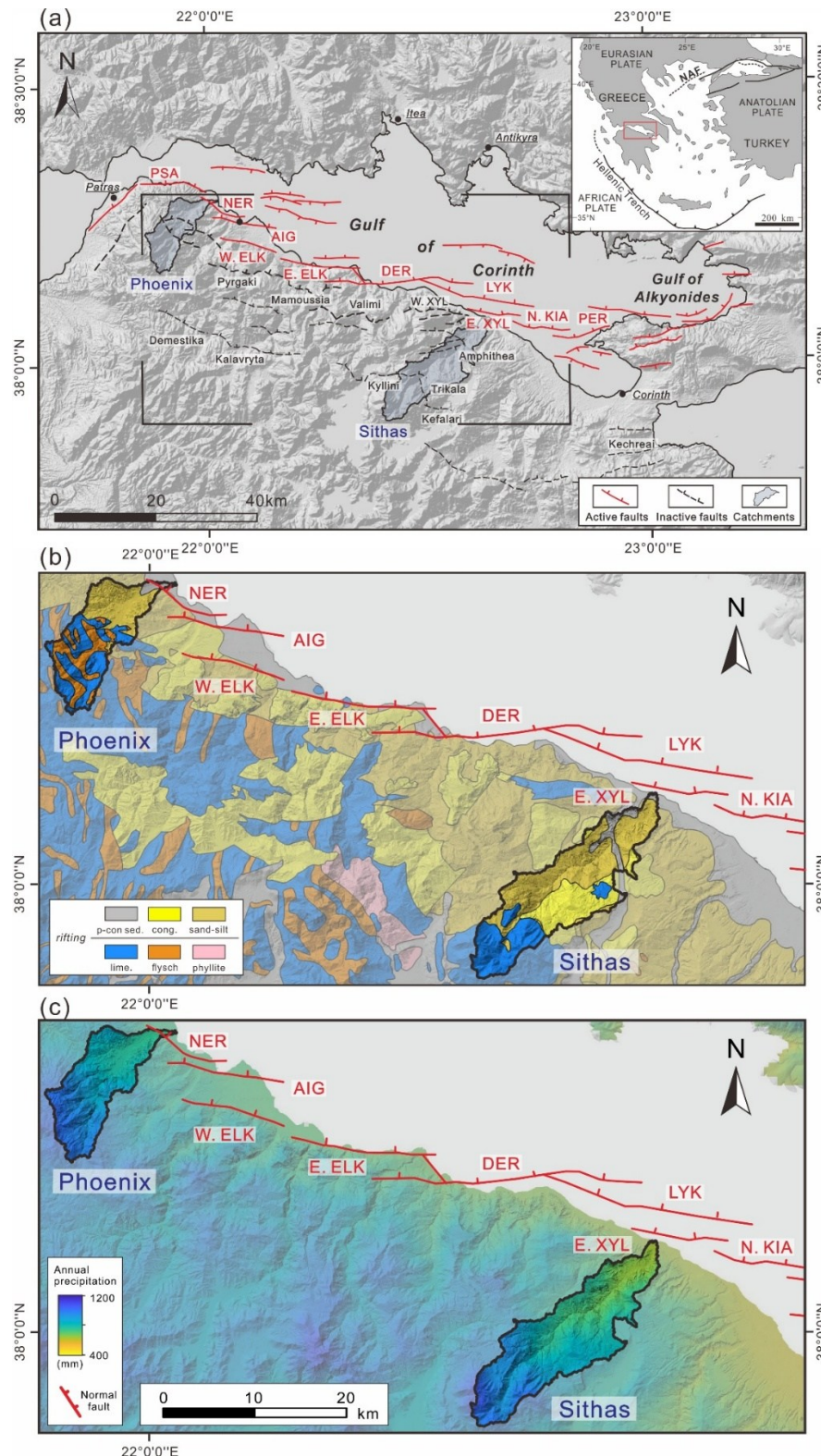


Figure 3.1. Gulf of Corinth in maps of hillshaded relief (a), lithology (b), and precipitation (c), with active faults and the two study catchments highlighted. Active fault abbreviations (red): AIG-Aigion fault, DER-Deveni fault, E. ELK-East Eliki fault, E. XYL-East Xylokastro fault, LYK-Lykoporia fault, N. KIA-North Kiato fault, PER-Perachora fault, PSA-Psathopyrgos fault, NER-Neo Erineos fault, W. ELK-West Eliki fault, E. XYL-East Xylokastro fault (after [Nixon et al., 2024](#)). Geological unit abbreviations: p-con sed. – poorly consolidated sediment; cong. – conglomerate; sand.-silt. – sandstone-siltstone; lime. – limestone.

Table 3.1. Basic characteristics of study catchments.

Catchment	Area	Climate	Tectonic perturbation		Heterogenous bedrock	
Sithas	153 km ²	~600-1000 mm/yr, minor orographic effect	1-2 mm/yr	>600 ka, significant knickpoints	lime. and flysch upstream, cong. and sand.-silt. downstream	lime. and flysch locally fractured
Phoenix	88 km ²			<100 ka, no associated knickpoints		lime. and flysch mostly massive

In the Gulf of Corinth, where contributing rivers range in drainage area from ~10 to ~1000 km² ([Watkins et al., 2020](#)), the Sithas and Phoenix rivers are comparable in both drainage areas (88 km² vs. 153 km²) and precipitation patterns ([Figure 3.1c](#); [Table 3.1](#)). Moreover, both rivers traverse clear and marked upstream-to-downstream gradients in tectonics and lithology ([Table 3.1](#)): the upstream zones are characterised by low uplift rates and hard bedrock, with trunk channels strongly coupled to hillslopes; by contrast, the downstream zones exhibit high uplift rates and soft bedrock, with sediment input dominantly from tributaries ([Figure 3.2](#)). Thus, both rivers serve as a natural laboratory for examining how the two forcings compete to control sediment grain size. In particular, the two rivers also differ in some tectonic and lithological respects ([Table 3.1](#)). For example, knickpoints developed in the Sithas River where tectonic perturbations occurred prior to 600 ka ([Zhou et al., 2024](#)), but are not present in the Phoenix River, perturbed within the last 100 ka ([Whittaker et al., 2024](#); [McNeill et al., 2005](#); [Table 3.1](#)). In addition, the hard limestone in the Phoenix river is interlayered with chert and is to first-order more deformed than massive limestone exposed in the Sithas river ([Wicker et al., 2024](#); [Table 3.1](#)). The boundary between upstream and downstream zones corresponds to upstream migrating knickpoints in the Sithas river ([Zhou et al., 2024](#)), but is anchored to an inactive fault in the Phoenix river ([Figure 3.2](#)). This well-constrained contrast allows us to assess the conditions under which tectonic or lithological forcing is propagated into sediment grain size.

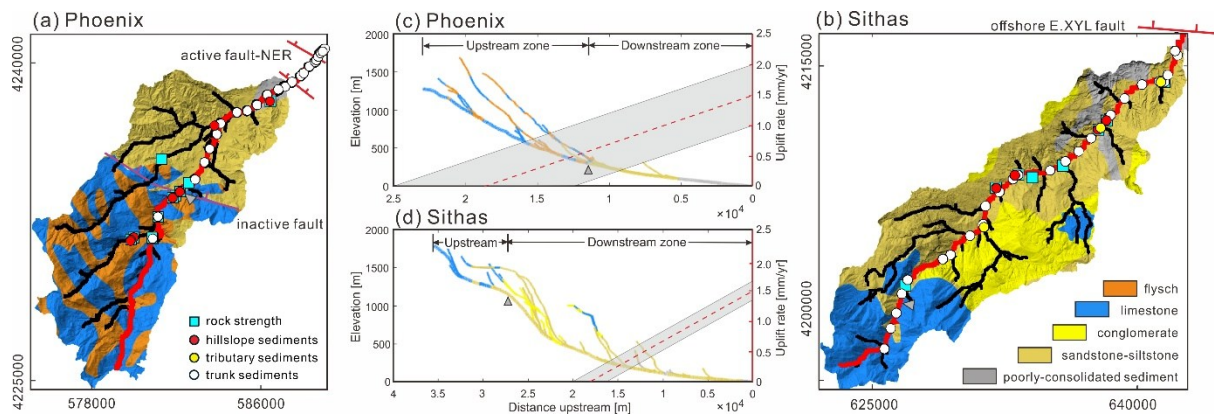


Figure 3.2. Main characteristics of two study catchments in maps (a, b) and river profiles (c, d), shown with trunk channels (red), key tributaries (black) and field localities marked by white, yellow, and red dots (sediment grain size measurement in trunk, tributary, and on hillslopes, respectively), as well as cyan squares (rock strength measurement). First-order variations in uplift rates are estimated using the linear back-tilting model derived from well-constrained faults in the Corinth rift (de Gelder et al., 2019; Zondervan et al., 2020). The triangles mark the boundary between the upstream and downstream zones in the trunk channel.

3.3 Methods

Here, we first measure rock strength to characterise the contrast between geological units. These lithological characteristics are then combined with tectonic constraints to assess their relative influence on geomorphic processes, as indicated by topographic metrics. In this context, the extent to which tectonic or lithological forcing is translated further into exported sediment grain size is examined by tracking grain size variations from hillslopes to tributaries and, ultimately, to catchment outlets.

3.3.1 Rock strength

To assess the impact of lithological contrasts, we adopted geological units defined in Bornovas and Rondogianni-Tsiambaou (1983) as a first-order grouping (Figure 3.1b). Four main geological units that differ in age and lithology exist in our study area: pre-rift flysch (*flysch*), pre-rift limestone (*lime.*), syn-rift conglomerate (*cong.*), and syn-rift sandstone to siltstone (*sand.-silt.*) (Figure 3.2). The four geological units are expected have distinct rock strengths, as age and lithology are main controls on rock strength (e.g. Campforts et al., 2020; Haag and Schoenbohm, 2025), a view further supported by statistically distinct topographic metrics among the units (Zhou et al., 2024). Nevertheless, variations in lithotype and the degree of fracturing may exist within each geological unit. For instance, the *flysch* is sandstone-dominated but also includes shale; we use the term ‘*flysch*’ to label this geological unit in line with previous studies (e.g. Papanikolaou and Lekkas, 2008; Skourlis and Doutsos, 2003).

To further characterise the rock strength range and variability within the four geological units, we measured rock strength at 26 localities using a Schmidt hammer (Figure 3.S1; Table 3.S1). The Schmidt

hammer is a portable device that has been widely used to estimate the uniaxial compressive strength of rocks (Bursztyn et al., 2015; Kent et al., 2021) and to characterise the degree of fracturing and weathering (e.g. Selby, 1982; Goudie, 2006; Borrelli et al., 2014). For each locality, we recorded rock strength measurements and their orientation (horizontal or inclined), along with a description of the bedrock lithology and its corresponding geological unit (Figure 3.3; Table 3.S1). The heterogeneity within geological units, such as lithotype variations, is thus captured in our approach. Furthermore, 22 rock strength measurements were taken on the bedrock surface to account for outcrop-scale heterogeneity (Bernard et al., 2019), including variations related to variable pebble size and composition in conglomerates. In addition, 693 measurements taken from adjacent river catchments where identical geological units are well-exposed were also compiled (Table 3.S1; Zondervan, 2017). This compilation is supported by similar median values and quartiles obtained from the two data sets for the same geological units. In total, 1596 measurements taken across the four main geological units were used for further analysis (Table 3.S1). The Schmidt hammer rebound values were corrected for hammer orientation using the instrument's conversion scale to obtain horizontal equivalent values. For each group of 22 readings, we then excluded the largest and smallest values to calculate the mean (Kent et al, 2020). In addition, we converted the Schmidt hammer rebound values (R) to uniaxial compressive strength (UCS), using the formula derived by Katz et al. (2000) for a range of limestones, sandstone, and soft chalk with R values similar to those found in the study area (Equation 3.1). This formula provides results that are comparable to those obtained by other well-known relations, such as the one by Yagiz (2009), giving us confidence that our UCS estimates are reasonable.

$$UCS = 2.21e^{0.007R} \quad (3.1)$$

Here, we grouped Schmidt hammer measurements and uniaxial compressive strength by geological units and presented them as box and whisker plots (Figure 3.4a-b) to constrain the rock strength range within each geological unit. In addition, rock strength variability was evaluated with field observations (Figure 3.4c-f; Table 3.S1) to identify the impact of factors such as fracturing.

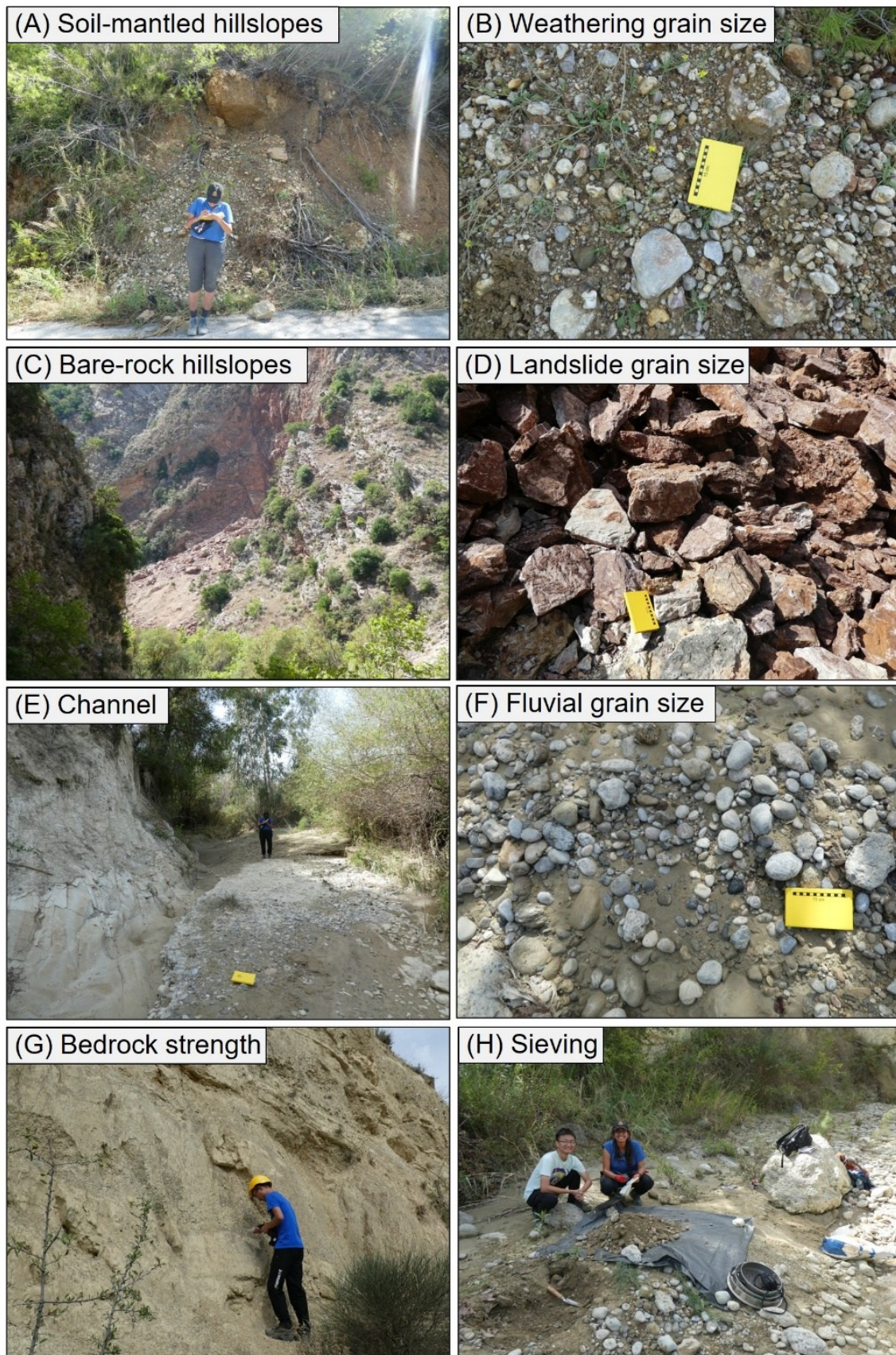


Figure 3.3. Field photographs illustrating hillslope and channel processes (A, C, E), their corresponding sediments (B, D, F), rock strength measurements (G), and sieving (H).

3.3.2 Topographic and mass-wasting analysis

To characterise geomorphic processes on hillslopes and in channels, we performed a topographic analysis. Here, a LiDAR DEM with a 5 m spatial resolution ([Hellenic National Cadastre & Mapping Agency S.A.](#)) was merged and projected in WGS 1984 UTM Zone 34N; we then extracted drainage networks and river long profiles using a threshold of 40000 pixels (i.e. 1 km²) ([Figure 3.2](#)). In this study, we focused on two important and commonly used topographic metrics: channel steepness and hillslope gradient. Their use is based on the concept that channels and hillslopes should adjust to achieve equilibrium with their boundary conditions, including tectonics, climate, and lithology. The first metric channel steepness is a variant of channel slope (S) that accounts for water discharge, which can be estimated from mean annual precipitation (P) and drainage area (A) (e.g., [Leonard et al., 2023](#)) ([Equation 3.2](#)).

$$S = k_s(PA)^{-m/n} \quad (3.2)$$

where k_s is the channel steepness, and m/n is the concavity (θ) ([Snyder et al., 2000](#)). Typically, a reference concavity (θ_{ref}) is used to normalise channel steepness (k_{sn}) in order to enable comparison among channels ([Wobus et al., 2006](#)). [Equation 3.2](#) can be integrated along the channel length to give:

$$z = z(x_b) + \left(\frac{k_{sn-p}}{Q_0^\theta} \right) \chi \quad (3.3)$$

$$\chi = \int_{x_b}^x \left(\frac{Q_0}{P_x A_x} \right)^\theta dx \quad (3.4)$$

where x_b is base level, χ is an integral quantity and Q_0 is a reference water discharge, set to 1 km³/yr ([Leonard et al., 2023](#); [Perron and Royden, 2013](#)). Here we chose $\theta_{ref} = 0.5$ based on past work in the region (e.g., [Zhou et al., 2024](#); [Zondervan et al. 2020](#)) and we used annual precipitation ([Figure 3.1c](#); [Hijmans et al., 2005](#)) to calculate water discharge-normalised channel steepness (k_{sn-p}). The k_{sn-p} value can be used as a proxy for the capacity of channels to incise bedrock and transport sediments ([Harries et al., 2021](#); [Leonard et al., 2023](#)). In general, channels with higher steepness for their discharge are able to transport coarser sediments (e.g. [Attal et al., 2015](#)).

Hillslope gradient was derived using the D8 algorithm in TopoToolbox ([Schwanghart and Scherler, 2014](#)) and used as a proxy for hillslope processes (e.g. [Roering et al., 2001](#)). At low hillslope gradients, bedrock weathers into sediments that are subsequently transported to channels via creep-like processes. This creep-like sediment transport can be approximated by a linear diffusional model ([Culling, 1960](#)). In this case, mean hillslope gradients can be used as a measure for relative erosion rates and therefore weathering time on hillslopes ([Attal et al., 2015](#); [Hurst et al., 2012](#)). As hillslope gradients gradually approach the critical threshold value ([Burbank et al., 1996](#)), the dominant hillslope

processes would transition from weathering and creep-like transport to mass-wasting processes, such as landsliding and rockfalls (e.g. Roering et al., 2001; Whittaker et al., 2010).

The DEM-based analysis is further complemented by our mass-wasting mapping on Google Earth Pro. The Google Earth imagery, acquired between 2017 and 2024, has a spatial resolution of ~0.5 m. This allows us to identify recent mass-wasting events (those can be identified from imagery, Borgomeo et al., 2014), using the criteria outlined by Guzzetti et al. (2012): distinct shapes (e.g. hillslope scarps and cone-like deposits), changes in colour, vegetation cover, terrain texture, and topographic settings (e.g. Borgomeo et al., 2014; Roda-Boluda et al., 2018). In this process, a hillslope scar closely associated with a cone-like deposit downslope is identified as a mass-wasting event, and its boundary is delineated. Where mass-wasting events were too pervasive to isolate individually, by necessity, we grouped them as a whole. This simplification is acceptable for our study, as we focus solely on estimating the area of catchment affected by mass wasting processes. Accordingly, we do not analyse frequency-area distributions for individual mass-wasting events, as some studies do in an effort to quantify mass-wasting volume (e.g. Guzzetti et al., 2002; Roda-Boluda et al., 2018). In total, a minimum of 590 active mass-wasting events were identified in the two catchments, and their digitized boundaries are provided in supplementary material.

To evaluate the tectonic and lithological controls on geomorphic processes, we first examined the spatial variations in k_{sn-p} , gradient, and mass-wasting (Figures 3.5 and 3.6). In this broad-scale analysis, we mostly focused on comparing the upstream and downstream zones, where distinct contrasts in tectonics and lithology are evident (Figure 3.2). In addition, we evaluated the tributary-scale geomorphic response to tectonics and lithology to identify the dominant control. Here, tributary mean k_{sn-p} and gradient, with standard variations as error margins, were plotted against its confluence distance from the catchment outlet. In this plot, the size of the tributary symbol was scaled according to the area ratio of that tributary and the whole catchment, with those exceeding 2% identified as key tributaries and highlighted in colour. Finally, we selected three tributaries from each catchment that vary systemically in uplift rates and lithology to interrogate the coupling between hillslopes and channels. Each tributary was divided into 10 evenly spaced reaches, for which k_{sn-p} , mean hillslope gradients and mass-wasting area (see below) within a 1,000 m-wide swath were calculated (Figure 3.7). The 2% threshold, three tributaries, and 10 reaches, above, are empirical values in our study that capture the most significant changes while avoiding unnecessary complexity.

3.3.3 Sediment grain size

To track sediment grain size evolution throughout the catchments, we characterised the grain size distributions of hillslope, tributary, and trunk river sediments. Two approaches were used: Wolman

point-counting and in-situ sieving. The first approach has been widely used to characterise coarse sediment grain size distributions ($> 1\text{mm}$, e.g. Attal and Lavé, 2006; Watkins et al., 2020) and is time-efficient (e.g. Litty et al., 2017; Miller et al., 2014; Wolman, 1954). Thus, we mostly employed Wolman point-counting for our purpose, including 55 channel localities where high-flow bars were sampled and 17 hillslope localities where representative areas (e.g. the central part of a landslide, Attal and Lavé, 2006) were visually selected. The Wolman point count method requires 100 random clasts within the sample areas to be selected, for which the intermediate axis is traditionally measured (Wolman, 1954). However, identifying the intermediate axis of clasts can be subjective and difficult to repeat (Brooke et al., 2018). Thus, we elected to measure the long axis, consistent with recent studies (Brooke et al., 2018; Watkins et al., 2020). This modified Wolman point count was applied to both physical clasts in the field (physical Wolman) and clasts identified in high-resolution, scaled photographs (photo Wolman). This is supported by many studies that demonstrate the two Wolman methods produce statistically indistinguishable distributions (Duller et al., 2010; Miller et al., 2014; Whittaker et al., 2010). Here we present the grain size data as cumulative distribution curves and focus on the 84th percentiles (D_{84}) of the distributions (Figure 3.9). D_{84} value is statistically representative of the coarse fraction and more sensitive to coarse sediment inputs than the D_{50} value (Whittaker et al., 2010). In addition, we assigned an empirical 15% error margin to the D_{84} value, as suggested by other Wolman datasets (Roda-Boluda et al., 2018; Whittaker et al., 2011).

The second approach, in-situ sieving, was employed to constrain the full distribution of grain sizes in 7 trunk-channel localities (Figure 3.3; Figure 3.S1). After choosing a representative area in high-flow bars (e.g. the central part), we first removed the surface layer to minimise the impact of armouring and then dug the subsurface sediments for sieving. The sediments were then sieved into grain-size bins of <10 , 10–20, 20–40 and 40–80 mm, and clasts greater than 80 mm were weighed individually (e.g. Quick et al., 2020; Watkins et al., 2020). In this process, we ensured the largest clast was $<10\%$ by weight of the total amount sieved in order to accurately capture the D_{84} (Bunte and Abt, 2001; Whittaker et al., 2010). Following Watkins et al. (2020), the intermediate diameter of clasts greater than 80 mm was estimated from their weights, assuming a spherical geometry and a density of 2450 kgm^{-3} . These clasts were then appropriately categorized into either 80–160 or 160–320 mm bins, and we used the upper bin limits (e.g. 20 mm for the 10–20 mm bin) to convert the cumulative mass distributions into grain size distributions (Figure 3.9). The sieved D_{84} value was compared with the Wolman D_{84} value to cross-validate the effectiveness of the two methods in capturing sediment grain size (Figure 3.S3) and the raw data is provided in Table 3.S2.

For the 17 sampled hillslope localities, the sediments were categorized by lithology and generation processes (i.e., weathering and mass-wasting processes, the latter including landslides, rockfalls, and

the formation of scree cones). For the same bedrock lithology, the transition from weathering to mass-wasting often depends on tectonic uplift rate (e.g. Larsen and Montgomery, 2012). Thus, this categorization allows the impact of lithology and tectonics on hillslope sediment grain size to be examined. Likewise, channel sediments from key tributaries (Figure 3.6; *t7*, *t9*) were also measured to evaluate directly if the tectonic or lithological controls also apply to tributary sediment grain size (Figure 3.8). The number of tributaries sampled is relatively limited (*t7* and *t9*), mainly due to their difficult accessibility. Hence, we also used sediment from a hillslope locality to provide a first-order estimate of tributary sediment (*t2*). This is supported by the proximity of the hillslope locality to the channel (tens of meters): the limited weathering time might allow the hillslope sediment to retain a size similar to that of the tributary sediment. An alternative strategy was used for other tributaries to examine the tributary sediment grain size - evaluating their impacts on trunk-channel sediments. From the perspective of sediment mixing, tributaries that coarsen the trunk-channel sediments should supply comparably coarse sediments (e.g. Ferguson et al., 2006; Rice, 1998). This observation can then be combined with tributary tectonic and lithological characteristics to indirectly evaluate whether tectonics or lithology is dominating tributary sediment grain size. To this end, we first plotted the grain size variations of trunk river sediments with respect to tributary confluences (Figure 3.10). Similarly, we also plotted hillslope gradients in a 1,000 m-wide swath along the trunk channel to evaluate the impact of hillslope inputs (Figure 3.10). Finally, hillslope and tributary impact on trunk-channel sediments are considered in the context of tectonics and lithology (Figures 3.10 and 3.11).

3.4 Results and analysis

3.4.1 Rock strength and geomorphic processes

Figure 3.4a and 3.4b present the Schmidt hammer rebound values and uniaxial compressive strength for the four key geological units in our study area. To first order, the four geological units exhibit distinct Schmidt hammer rebound values, with the median for *lime.* > *flysch* > *cong.* > *sand-silt* (Figure 3.4a). The uniaxial compressive strength, derived from Schmidt hammer rebound values, follows the same pattern: the median is ~70 MPa in the *lime.*, ~29 MPa in the *flysch*, ~26 MPa in the *cong.*, and ~9 MPa in the *sand-silt* (Figure 3.4b). Notably, the median values reported here are comparable with those in other studies (Goudie, 2006; Zondervan et al., 2020). However, variations in rock strength can also be observed within these geological units (Figure 3.4a). In general, these variations are associated with the degree of fracturing and consolidation, as well as significant changes in lithotypes (Figure 3.4; Table 3.S1). For example, fractured flysch and limestone have significantly lower rock strength compared to their unfractured counterparts (Figure 3.4c). In addition, the fractured shale layer in the *lime.* exhibits extremely low rock strength (~19) (Figure 3.4d). We also find that occasional

conglomerate beds in the *sand.-silt.* exhibit much lower rock strength (~ 12 , Figure 3.4e) than the strongly consolidated Gilbert-delta conglomerate in the *cong.* (~ 55 , Figure 3.4f), illustrating the significant impact of consolidation.

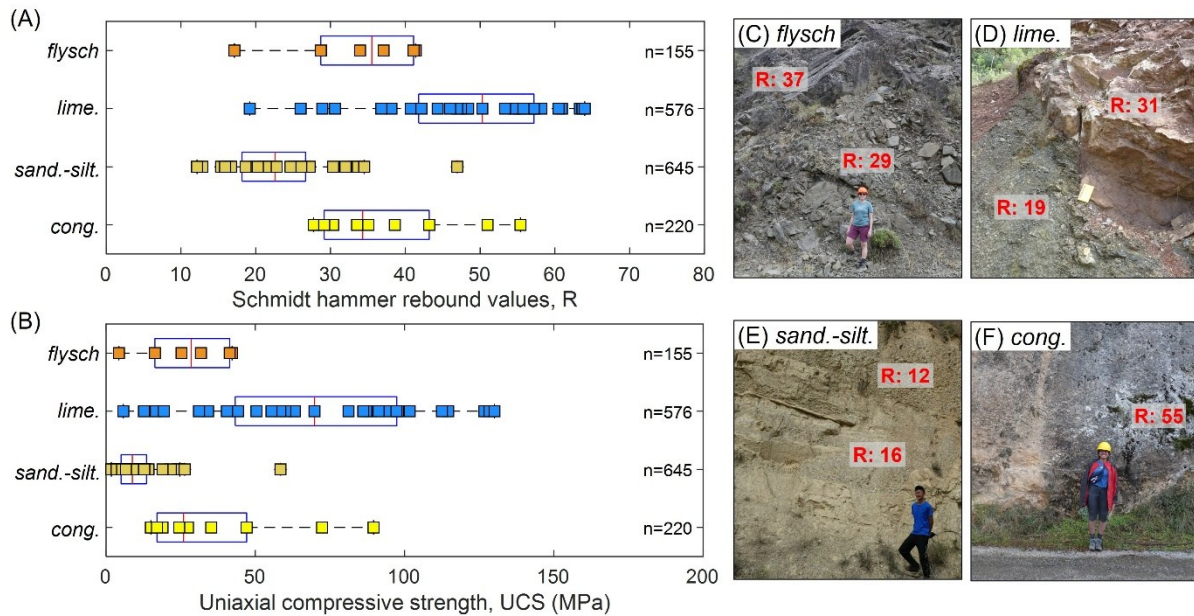


Figure 3.4. Schmidt hammer rebound values (A) and uniaxial compressive strength (B) of the four geological units in the studied catchments, shown with field photos (C, D, E, F). Red text refers to the mean representative Schmidt hammer rebound values taken nearby (Table 3.S1).

To characterise hillslope and channel processes, the distributions of discharge-normalised channel steepness (k_{sn-p}), hillslope gradients, and mass-wasting events were mapped (Figure 3.5 and 3.6). In the Phoenix catchment, k_{sn-p} increases significantly from the downstream to the upstream zone (Figure 3.5a), where bedrock hardens but uplift rates decrease (Figure 3.2). Similarly, hillslope gradients are also higher in the upstream zone, where steep hillslopes (purple) are preferentially observed (Figure 3.5b). These observations thus demonstrate lithological variations dominate hillslopes and channels, with tectonic influence being a secondary factor. In the tributary-scale analysis, the tributary mean k_{sn-p} also increases across the upstream-downstream boundary (Figure 3.5d); however, the mean gradient exhibits an insignificant change (Figure 3.5e), likely due to the wide spatial extent of the tributaries (Figure 3.5a). Indeed, tributaries that span both the upstream and downstream zones can disrupt the potential pattern (e.g. *t5*; Figure 3.5). In addition, mass-wasting events exhibit clear spatial variation, with widespread occurrences in the upstream zone ($\sim 10\%$ by area) and few near the active fault despite the high uplift rate (approaching 0% by area, Figure 3.5c, 3.5f).

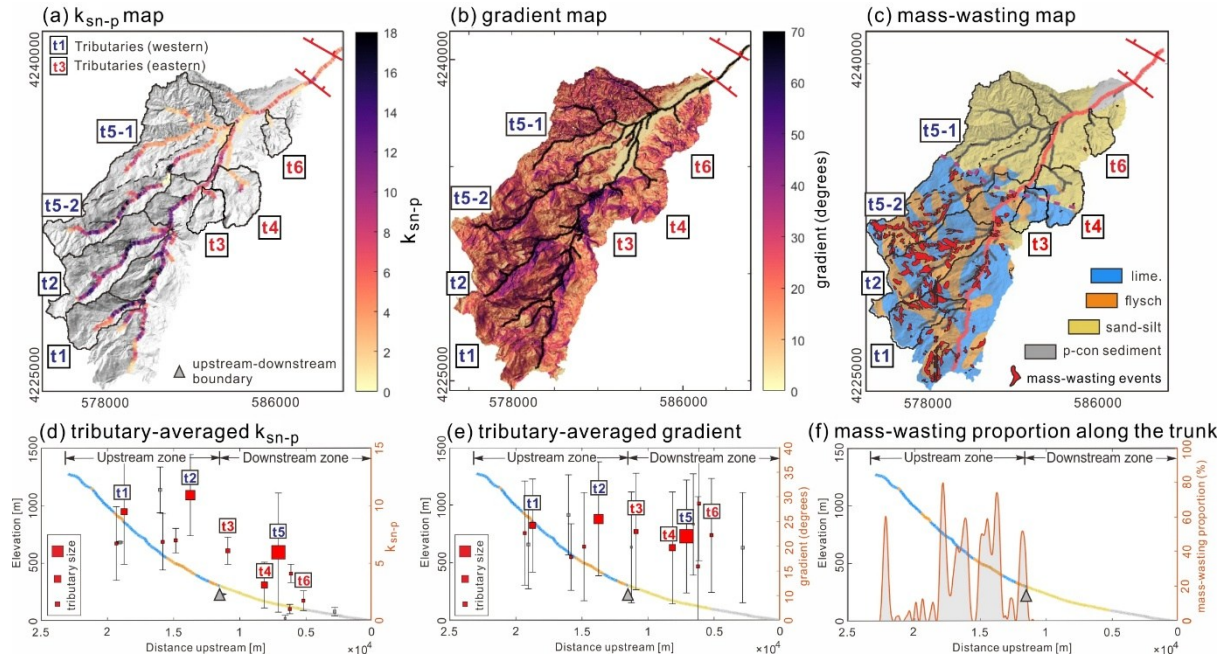


Figure 3.5. The Phoenix river k_{sn-p} map (a), gradient map (b) and mass-wasting map (c). Tributary-averaged k_{sn-p} , gradient (with standard deviations as error margins; d, e), and mass-wasting density (f), along the trunk channel, shown as a profile coloured by lithology in the background. Here, mass-wasting density is defined as the proportion of the area affected by mass wasting.

In the Sithas catchment, first-order variations in k_{sn-p} and hillslope gradients are associated with the main tectonic knickpoints (Figure 3.6). The knickpoints from the same tectonic perturbation separate the steepened downstream reaches from the gentler upstream reaches, illustrating the impact of the transient erosional dynamics documented by Zhou et al. (2024). In comparison with tectonic control, lithological control on k_{sn-p} and hillslope gradients is less evident but still exists (Figure 3.6). For instance, in the tributary *t1*, the *lime.* has steeper hillslopes and higher k_{sn-p} than conglomerate (Figure 3.6a, 3.6b). In short, both tectonics and lithology are important in controlling hillslopes and channels in the Sithas catchment. In the tributary-scale analysis, k_{sn-p} and hillslope gradients exhibit large variations in their mean values, as both are likely to exhibit bimodal distributions due to the impact of tectonic knickpoints (Figure 3.6d and 3.6e). Despite the large variations, tributary mean slope systematically decreases towards the knickpoint, which we interpret as a tectonic signal. In contrast, the tributary-averaged k_{sn-p} increases towards the knickpoint, likely reflecting the lithological transition from the *sand-silt* to the *cong.* geological unit (Zhou et al., 2024). Thus, tributary-scale hillslopes and channels are interpreted to be dominated by tectonics and lithology, respectively. Finally, mass-wasting events in the Sithas catchment are pronounced in the *sand-silt*-dominated downstream zone, especially the immediate footwall area where hillslopes are steepened (Figure 3.6c, 3.6f; Figure 3.S2a, 3.S2b). In addition, mass wasting in the *lime.*-dominated upstream zone is mostly concentrated on steep hillslopes along the trunk channel (Figure 3.6c, 3.6f).

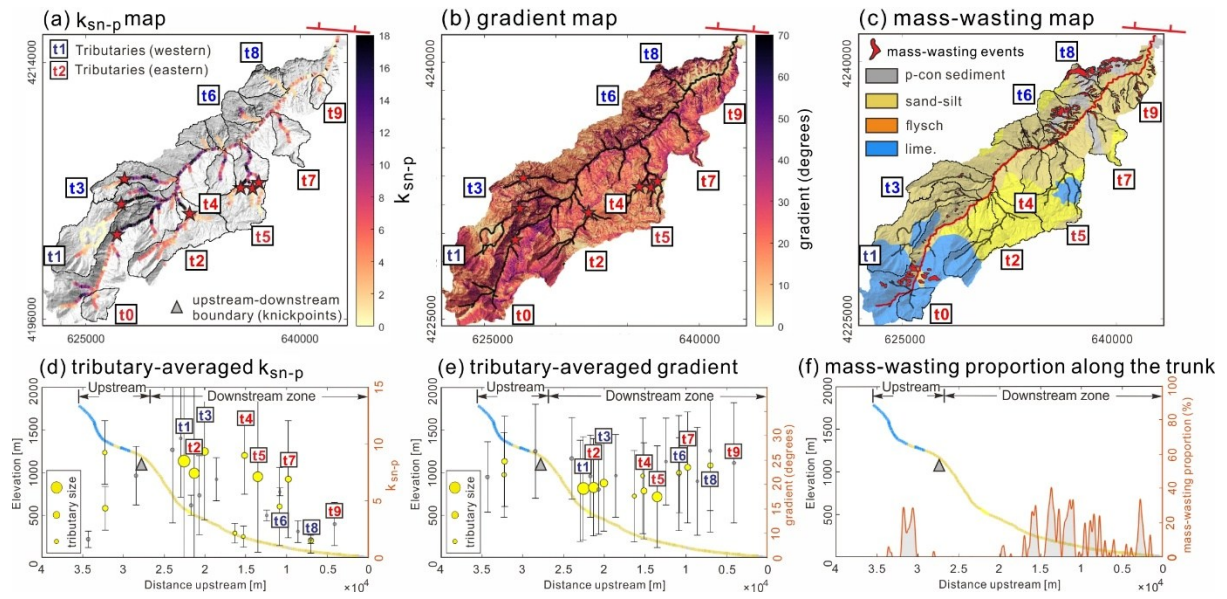
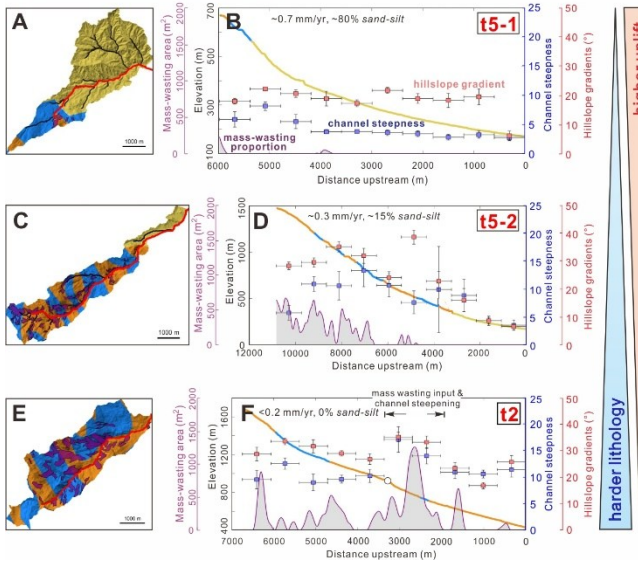


Figure 3.6. The Sithas river k_{sn-p} map (a), gradient map (b) and mass-wasting map (c). Tributary-averaged k_{sn-p} , gradient (standard deviations as error margins; d, e), and mass-wasting density (f), along the trunk channel, shown as a profile coloured by lithology in the background. Here, mass-wasting density is defined as the proportion of the area affected by mass wasting.

Consequently, in response to tectonic perturbations and heterogeneous bedrock, hillslopes and channels in the Phoenix and Sithas catchments behave differently. To further examine the hillslope-channel coupling, we selected three key tributaries from each catchment that exhibit systemic variations in uplift rates and lithology. Figure 3.7a-f show hillslope gradients, mass wasting, and k_{sn-p} for 3 key tributaries in the Phoenix catchment (see Figure 3.5a for location). In such tributaries, channels and hillslopes mostly stabilize at lithology-dependent values (termed here ‘background’ values, Figure 3.7a-f). For example, the areas dominated by the sand-silt mostly have hillslope gradients of $\sim 20^\circ$ and k_{sn-p} of ~ 3 , whereas the Pindos limestone and flysch often exhibit steeper hillslopes ($\sim 25^\circ$) and higher k_{sn-p} (~ 10) (Figure 3.7b, 3.7d, 3.7f). Notably, mass-wasting events in the Pindos limestone- and flysch- dominated areas are pervasive, rather than being limited to steep hillslopes along channels (Figure 3.7d, 3.7f). In tributary t2 (Figure 3.5a), where mass-wasting sediment inputs occur, channels and hillslopes become steeper than their ‘background’ values (Figure 3.7f). In contrast, the three Sithas tributaries (Figure 3.6a) are characterised by migrating tectonic knickpoints (Figure 3.7g-l). These knickpoints are followed by k_{sn-p} spikes and continuous steepened hillslopes (Figure 3.7g-l). The downstream tributary t9 (Figure 3.6a), with a knickpoint that has nearly migrated through, exhibits the steepest tributary hillslope gradients ($\sim 25^\circ$) (Figure 3.7h), along which mass-wasting events are concentrated (Figure 3.7g-l; Figure 3.S2). The local impact of knickpoints on k_{sn-p} , however, is exceeded by the global impact of lithology (Figure 3.7h, j, l), allowing tributaries with higher mean k_{sn-p} values (~ 5 , Figure 3.7j, l) to be found in more upstream locations with harder bedrock (t2 and t7) (Figure 3.6a).

The Phoenix tributaries



The Sithas tributaries

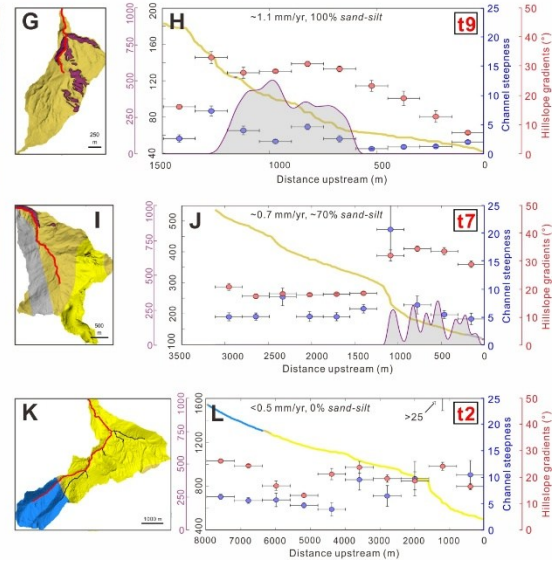


Figure 3.7. Representative tributaries in map view (A, C, E, G, I, K) and their trunk river profiles (B, D, F, H, J, L). Tributaries from the Phoenix and Sithas catchments are shown in the left (A-F) and right (G-L) panels, respectively (see Figure 3.5 and Figure 3.6 for locations and legends). Here, the text labels in the figure indicate that both the estimated uplift rate and the proportion of soft sand-silt decrease from A to C to E and from G to I to K. For both k_{sn} and gradient, horizontal bars indicate the river reach over which the mean value was calculated, while vertical error bars represent the corresponding standard deviation.

3.4.2 Sediment delivery from hillslopes and tributaries

Hillslope and channel processes, illustrated by the metrics in the section above, ultimately combine to convert intact bedrock into mobile sediments. Figure 3.8 shows the grain size distributions of hillslope and tributary sediments. In general, hillslope sediments supplied from the *lime.* and *flysch* geological units are coarser than those from the *cong.* and *sand-silt* (Figure 3.8a). This is consistent with the hypothesis that stronger bedrock fragments into coarser grains (Figure 3.4, Roda-Boluda et al., 2018). The wide range of sediment sizes observed on hillslopes underlain by the same geological unit (Figure 3.8a) is also interpreted to result partly from rock strength variations within that unit (Figure 3.4). For example, the grain size patchiness observed in a landslide deposit within the *lime.* geological unit may also reflect the interlayering of hard and soft rock layers in the Pindos unit (Figure 3.8a, b). Field observations also indicate that if conglomerate bedrock is weak, individual clasts within the conglomerate can be recycled; however, conglomeratic boulders up to meters in size can also be produced from hillslopes underlain by strong, well-cemented conglomerate bedrock (Figure 3.8c; Table 3.S1). Furthermore, our data shows that mass-wasting processes produce coarser hillslope sediments than those generated by weathering processes (Figure 3.8a). This is well-illustrated by the *sand-silt* hillslopes which can produce grains with $D_{84} > 200$ mm through landsliding, although these grains are friable (Figure 3.8d). In addition to those hillslope sediments, sediment grain size data from key tributary channels (t9 and t7) and from adjacent hillslopes tens of meters from the tributary

channel (t2) in the Sithas catchment were also collected. The results show that for the Sithas river, the tributary with steeper slopes and more pervasive landslides, but dominated by the *sand-silt* geological unit, does not export coarser grains than the other two we consider (i.e. t9; Figure 3.8a); instead, coarser grains come from those tributaries underlain by harder lithologies, albeit with lower slope gradients (i.e. t7 and t2; Figure 3.8a).

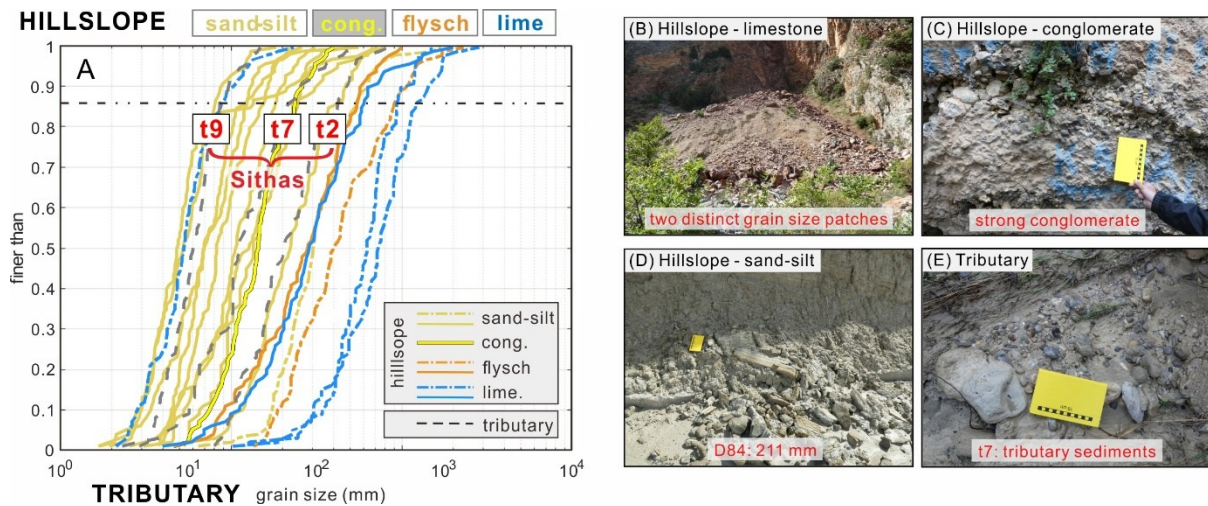


Figure 3.8. Sediment delivery from weathering (solid lines) and mass-wasting (dotted dashed lines) processes on hillslopes and from tributary channels (dashed lines) (A), shown with photos of their characteristic sediments (B, C, D, E).

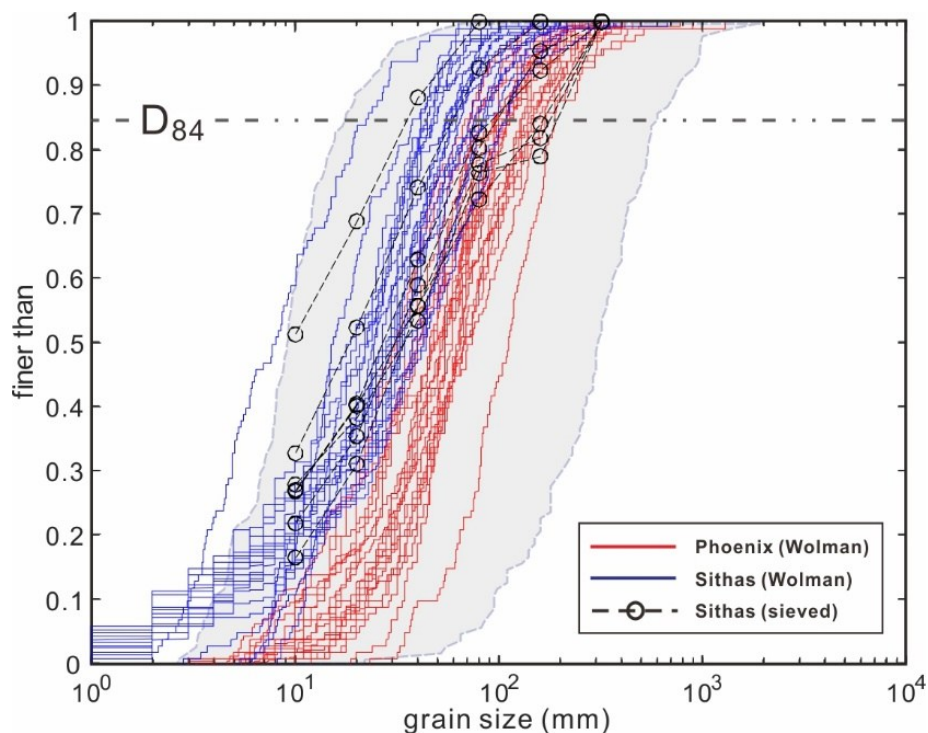


Figure 3.9. Trunk-channel sediment grain size for both the Sithas and Phoenix rivers. The shaded area represents the grain size range of hillslope sediments available for channel transport.

3.4.3 Downstream variations in sediment grain size

Figure 3.9 shows that compared to hillslope sediments (grey shaded zone), trunk channel sediments exhibit a markedly narrower grain size range. This observation clearly illustrates that channels selectively entrain hillslope sediments of certain sizes. In addition, the trunk-channel sediment grain size in the Phoenix river is systematically coarser than that in the Sithas river (Figure 3.9). Even within the same river, variations in trunk-channel sediment grain size are evident (Figure 3.9). Notably, the grain-size variations cannot be attributed solely to estimation method, as the two approaches are cross-validated by their positive correlation (Figure 3.9; Figure 3.S3). Here, we hypothesise that additional factors, such as local sediment supply, impact the trunk-channel sediment grain size (e.g. Sklar et al., 2020).

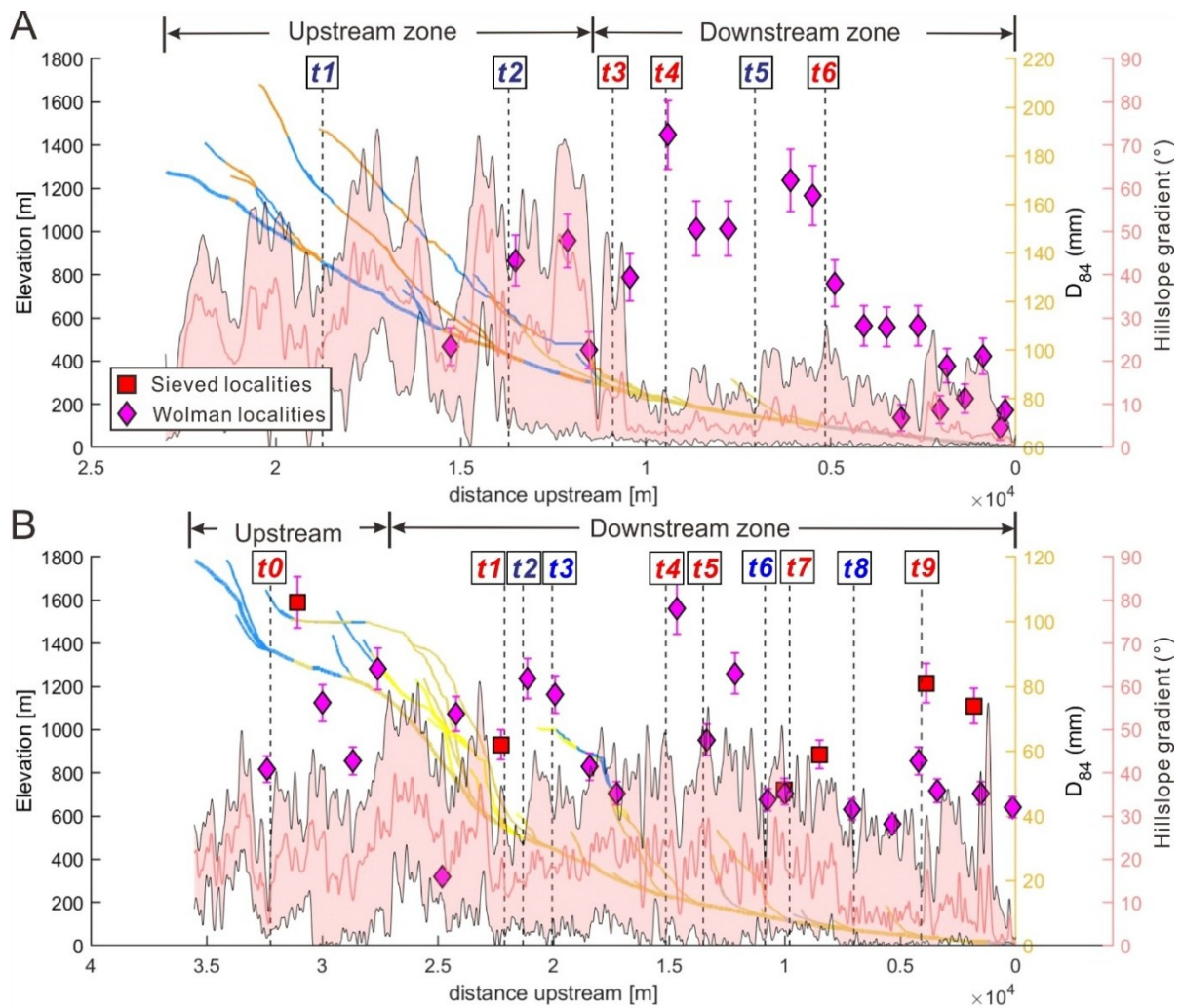


Figure 3.10. Downstream variations in trunk-channel sediment D_{84} for the Phoenix (A) and Sithas (B) catchments, shown with hillslope gradients in a 1,000 m-wide swath (pink bands) and key tributary junctions (dashed lines) along trunk channels coloured by lithology. Note, the Wolman grain size is used for both sieved (red squares) and Wolman (purple diamonds) localities to enable direct comparisons (Figure 3.S3). The eastern and western tributaries are marked in red and blue text, respectively. The minimum, mean, and maximum hillslope gradients in a 1,000 m-wide swath along the trunk channel are presented.

To examine sediment supply control on grain size variations along the trunk channels, trunk-channel sediment D_{84} was plotted alongside hillslope gradients within a 1,000 m-wide swath and tributary confluence locations (Figure 3.10). In general, the trunk-channel sediment D_{84} from both rivers vary with local hillslope gradients in the upstream zones, where channels are strongly coupled with hillslopes (Figure 3.10). This correlation leads to downstream coarsening D_{84} in the Phoenix catchment (Figure 3.10a) and fluctuating D_{84} in the Sithas catchment (Figure 3.10b). In the downstream zones where tributary input dominates, however, trunk-channel sediment D_{84} exhibit punctuated downstream fining (Figure 3.10). This is well-illustrated in the downstream Phoenix catchment, where local grain size peaks punctuating the overall fining pattern (Figure 3.10a) occur immediately downstream of tributary confluences (e.g. t4, t5). The tributary t6, located closest to the active fault but composed of only the *sand.-silt.*, is an exception that appears to accelerate the downstream fining (Figure 3.10a). In the downstream Sithas catchment, most eastern tributaries (highlighted in red text, e.g. t2, t4, t7, t9) coarsen trunk-channel sediments, between which downstream fining prevails (Figure 3.10b). This analysis thus reveals which hillslopes and tributaries are expected to supply coarse sediments—that is, those that contribute to the coarsening of trunk-channel sediments.

In this context, we further assess whether lithology or tectonics is the dominant control on coarse sediment delivery from these sources. Figure 3.11 shows that for upstream localities characterised by strongly coupled hillslopes, a strong positive correlation between hillslope gradient and trunk-channel sediment D_{84} can be observed, both across the two catchments and within each catchment (Figure 3.11). Importantly, these upstream localities exhibit spatially variable hillslope gradients: such spatial variations do not follow the systematic upstream decrease in uplift rates (Figure 3.2) but may instead be explained by differences in rock strength, as shown in Figure 3.4 (e.g. Roda-Boluda et al., 2018). For downstream localities characterised by intermittent tributary inputs, trunk-channel sediment D_{84} in the Phoenix catchment exhibits no correlation with hillslope gradients (Figure 3.11). In the Sithas catchment, the downstream localities cluster at 8° and 20° hillslope gradients. For such downstream localities, no consistent correlation is observed between D_{84} values and hillslope gradient (Figure 3.11). This finding supports that hillslope inputs no longer dominate trunk-channel sediment in the downstream zones, as tributary inputs become the main sediment supply mode. In this case, we distinguished the tributaries exhibiting coarsening effects on trunk sediments (Figure 3.10), and found they mostly have at least 30% *cong.*, *flysch* or *lime.* in their tributary catchments (Figure 3.S4). In addition, such tributaries are not preferentially distributed in areas experiencing higher uplift (i.e. near active faults, Figure 3.2) (Figure 3.10). These observations demonstrate that lithology, rather than tectonics, is the dominant control on tributary sediment grain size.

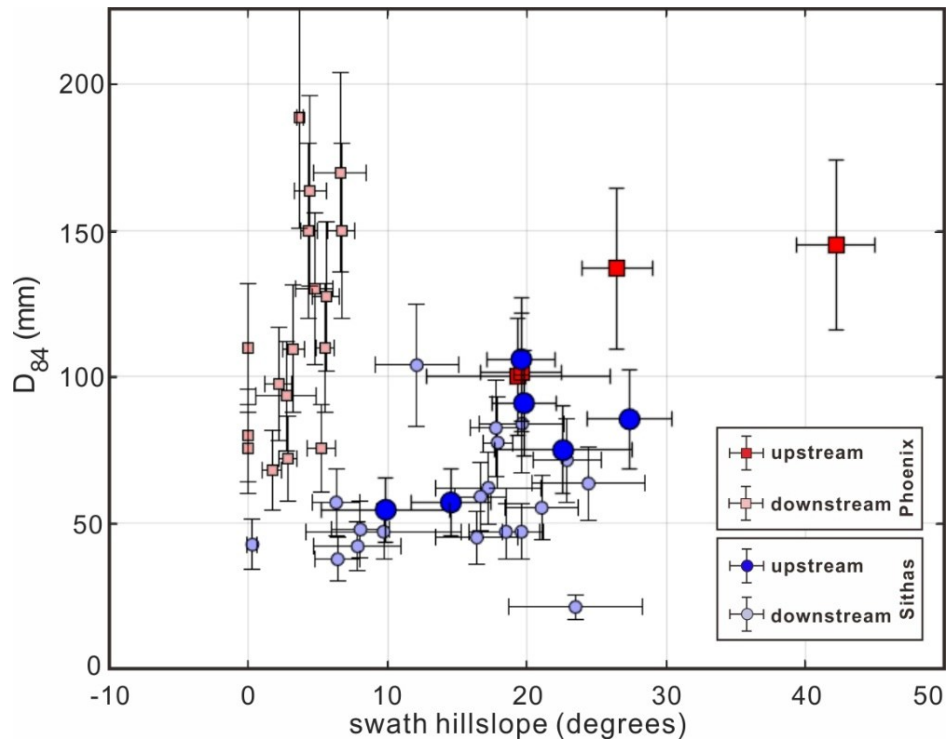


Figure 3.11. Trunk-channel sediment D_{84} versus swath hillslope gradient. Hillslope gradients were calculated as the mean gradient over a 300 m-long and 1,000 m-wide swath centred on the trunk-channel localities.

3.5 Discussion

3.5.1 Landscape dynamics driven by tectonics and lithology

In this study, we use discharge-normalised channel steepness (k_{sn-p}) and hillslope gradients to characterise tectonic and lithological controls on geomorphic processes. The broad-scale results show that tributary-mean k_{sn-p} values in both the Phoenix and Sithas catchments primarily correlate with lithological variations (Figures 3.5d and 3.6d); however, hillslope gradients are mainly associated with lithology in the Phoenix catchment and with tectonics in the Sithas catchment (Figures 3.5e and 3.6e). The topographic contrast suggests that landscape drivers and hillslope-channel couplings likely differ in the two catchments. Below, we exploit our tributary analysis (Figure 3.7) to examine hillslope-channel couplings and their influence on topographic response to tectonics and lithology.

In the Sithas catchment, channels and hillslopes are steepened by tectonic knickpoints that migrated from the downstream active fault (Figure 3.7; Zhou et al., 2024). This is supported by the systematic decrease in tributary-averaged hillslope gradients (Figure 3.6e), illustrating that upstream-migrating knickpoints rejuvenated the downstream tributaries earlier, and therefore more fully, than upstream ones (e.g. Gallen et al., 2011). However, unlike hillslope gradients, tributary-averaged k_{sn-p} does not exhibit a clear upstream decreasing pattern (Figure 3.6d). This discrepancy is likely due to channels responding more rapidly to knickpoints than hillslopes, as evidenced by transient channel steepening

limited to the reach immediately downstream of knickpoints: channels further downstream quickly recovered from transient steepening (Figure 3.7). The interpretation that channels outpaced hillslopes suggests knickpoint-driven channel incision regulates adjacent hillslopes (e.g. Hurst et al., 2012), not the reverse. In addition, mass-wasting is concentrated on the steepened hillslopes along channels (Figure 3.7; Figure 3.S2) supporting the inference that channel steepening propagates into hillslopes. Together, our observations in the Sithas catchment conform to the model in which migrating knickpoints undercut adjacent hillslopes and therefore increase hillslope gradients and mass-wasting frequency (e.g. Bennett et al., 2016; Bigi et al., 2006; Gallen et al., 2011). Thus, knickpoints initiated by tectonic perturbation drive the landscape evolution in the Sithas catchment.

In contrast, for the Phoenix catchment, both channels and hillslopes are much steeper in the upstream zone than in the downstream zone (Figure 3.5). This cannot be explained by upstream decreases in tectonic uplift (Figure 3.2) or increases in precipitation (Figure 3.1), leaving bedrock lithology as the explanation. In general, the upstream *lime*. (the Pindos limestone and flysch) exhibits rock strength much stronger than the *sand-silt* in the downstream zone (Figure 3.4). The stronger bedrock, with greater resistance to erosion, can facilitate the development of steep hillslopes and channels in the upstream zone (e.g. Gallen, 2018; Jansen et al., 2010). In addition, mass wasting is much more frequent in the upstream zone than in the downstream zone (Figure 3.5), with mass-wasting deposits being strikingly pervasive, i.e. not limited to steep hillslopes along channels (Figure 3.7). Here, we suggest that pervasive mass wasting reflects the locally interlayered and deformed nature of the Pindos limestone and flysch (Wicker et al., 2024), as mass wasting can be facilitated where the soft layers in interlayered rock are preferentially degraded (e.g. Eberhardt et al., 2005), or where intense local deformation from folding and thrusting occurs (Wicker et al., 2024). The lithology-induced mass wasting can release coarse sediment that steepens channels (higher k_{sn-p}), thereby promoting steeper hillslopes (e.g. Korup, 2006). This aligns with the observation that both channel and hillslope gradients exceed 'background' levels where mass wasting sediment input has occurred (Figure 3.7f). Thus, we suggest the bedrock lithology, through its resistance to erosion and its influence on mass wasting, is an important driver of the landscape in the Phoenix catchment.

Overall, our results illustrate how tectonics and lithology are translated into topography through hillslope-channel couplings. In areas with migrating knickpoints, such as the Sithas catchment, tectonic control is introduced into the topography through steepened channels following knickpoints and their destabilizing effects on adjacent hillslopes. In areas with mass-wasting-prone lithologies, bedrock lithology can also steepen channels and hillslopes, but through the impact of mass wasting deposits, as illustrated in the upstream zone of the Phoenix catchment.

3.5.2 Tectonic versus lithological controls on exported sediment grain size: lost in translation?

When channel and hillslope processes incise in response to tectonics and lithology, sediments are also released from intact bedrock. This underpins the possibility that tectonic and lithological forcings can translate into exported sediment grain size, but the extent of this translation remains contentious (e.g. Jerolmack and Paola, 2010). In the Gulf of Corinth region, Watkins et al. (2020) reported that exported sediment grain size at river outlets correlates primarily with catchment lithology and only weakly with tectonics. Here, we exploit our data sets to answer where the tectonic signal was shredded and how the lithological signal came to dominate.

First, we note that tectonics dominate hillslope gradients in the downstream Sithas catchment through upstream migration of knickpoints (Figures 3.6 and 3.7g-l). In contrast, in the Phoenix catchment where knickpoints have not yet developed, the transmission of tectonic signals into the landscape is impeded (Figures 3.5 and 3.7a-f). This contrast underscores the central role of knickpoints in translating tectonic forcing into topography and hence sediment grain size. In particular, the elevated gradients generated by knickpoint migration can promote the occurrence of mass-wasting events, as illustrated in the downstream *sand-silt* dominated hillslopes (Figure 3.7g-j). Combined with the observation that mass-wasting processes produce coarser sediment grains than weathering (Figure 3.8), our results clearly indicate that tectonic forcing can be transferred into hillslope sediment grain size (Attal et al., 2015), even for soft bedrock. However, this coarsening effect on sediments generated on *sand-silt* hillslopes does not extend into the trunk channel. Indeed, if this coarsening effect on hillslope sediments is maintained throughout tributary channel transport, we expect tributaries with steeper hillslopes and more frequent mass-wasting (i.e. near the fault) to export coarse sediments and therefore coarsen the trunk sediments. However, the opposite was observed: 1) the tributary with overall steeper hillslopes and more intense landslides (*t9*) exports finer grains than two other tributaries (*t7* and *t2*) (Figure 3.8), and 2) tributaries near the fault do not systematically coarsen trunk sediments (Figure 3.10). Thus, we hypothesize that the coarsening effect on such hillslope sediments is erased by abrasion during transport (Figure 3.S2). In an experimental study, Attal and Lavé (2009) found the abrasion rate for poorly consolidated sandstones can be up to 31%/km, 12 times higher than for limestones (2.6%/km). Their experimental abrasion rates confirm that intense size reduction through abrasion is geologically plausible (e.g. Attal and Lavé, 2006; Dingle et al., 2017). In addition, the lithology-dependent abrasion rates reconcile our results with those of Whittaker et al. (2010), where exported sediment grain size correlates with tectonic perturbations in the *lime*-dominated catchments. There, limited abrasion of the *lime*-dominated sediments likely attenuates, rather than obliterates, tectonic signals.

In contrast, lithological signals survived through the hillslope and channel processes (Watkins et al. 2020). Here, our results suggest that lithological control on sediment grain size is first introduced through its distinct rock strength and influence on the occurrence of mass wasting. For example, the harder *lime*. and *flysch* are weathered into coarser hillslope sediments than the softer *cong.* and *sand-silt* (Figure 3.8a). This is interpreted to reflect that harder bedrock has stronger resistance to fragmentation, thereby generating coarser hillslope sediments (Allen et al., 2015; Locat et al., 2006; Sklar et al., 2017). The strong positive correlation between hillslope gradient and trunk-channel sediment grain size in the upstream zones (Figure 3.11) likely also testifies that rock strength, through its influence on hillslope gradient, controls sediment grain size (e.g. Roda-Boluda et al., 2018). In addition, lithological control on the occurrence of mass wasting plays a significant role in controlling sediment grain size (e.g. Roda-Boluda et al., 2018). For example, the sediments transported in the Phoenix catchment, where mass wasting prevails, are systematically coarser than those in the Sithas catchment (Figure 3.9). Importantly, this observation also indicates that lithological signals imprinted in hillslope sediments extend into the trunk channels and even reach catchment outlets (Watkins et al., 2020). This is partly evidenced by the finding that tributaries dominated by hard bedrock coarsen the trunk-channel sediments (Figure 3.10). But what mechanism is responsible for signal propagation? Even though limestone and flysch are resistant to abrasion (Attal and Lavé, 2006), selective transport in the Phoenix catchment (Figure 3.S2) can filter hillslope sediments above a certain critical size and may therefore shred the carried signals (e.g. Litty et al., 2017). This filter effect is well-illustrated by the narrower size range of channel sediments than that of hillslope sediments (Figure 3.9). However, the critical size for entrainment is partly set by channel steepness (Bekaddour et al., 2013) and mass-wasting sediment supply itself can steepen channels (Figure 3.7f). Thus, we interpret that the channel steepness response to coarse sediment input facilitates the propagation of signals imprinted in sediment grain size.

Together, our process-based analysis reveals that tectonic perturbations alone do not guarantee the export of coarser sediments from perturbed catchments (c.f. Armitage et al., 2011; Attal et al., 2015; Whittaker et al., 2010). This could arise because the corresponding tectonic knickpoints have not yet been developed or because of strong abrasion, both of which can shred tectonic signals. In the case where the perturbed catchments are also lithologically heterogeneous, lithology can come to dominate the exported sediment grain size.

3.5.3 Implications for hangingwall stratigraphy

It becomes clear that sediment generation and transport processes occurring on hillslopes and within channels mediate tectonic and lithological controls on the grain size of sediments exported to the

basin (e.g. Jerolmack and Paola, 2010; Romans et al., 2016). This holds fundamental implications for stratigraphic interpretations (e.g. Allen et al., 2015). In our study, weathering and mass wasting, the two dominant sediment generation processes, are controlled by lithology (Figure 3.12a, e.g. Roda-Boluda et al., 2018) and tectonics (Figure 3.12b, e.g. Attal et al., 2015). Thus, along-strike variations in tectonics or lithology can markedly change sediment grain size exported to basins (e.g. Watkins et al., 2020) and thereby influence stratigraphy. However, only a few studies have invoked tectonically- and, in particular, lithologically- induced sediment grain size variations to account for stratigraphic changes (e.g. Allen et al., 2015; Armitage et al., 2011). In the North Alpine foreland basin, Schlunegger and Norton (2015) suggest that the contemporaneous progradation of the proximal facies and retrogradation of the distal gravel front cannot be explained by tectonic or climatic changes, but instead requires increased sediment grain size and variability, driven by the emergence of resistant bedrock. In the extensional Inner Moray Firth Basin, Scotland, McArthur et al. (2012) attribute a shift from siliciclastic to carbonate dominated sedimentation to lithological change in upland catchments. Here, our study shows how and why exported sediment grain size can vary in response to tectonics and lithology (Figures 3.12a, b). Taking the Phoenix as an example, we expect continuous incision would unroof the *sand-silt* and expose the underlying *lime*. and *flysch*, leading to the development of an upward-coarsening stratigraphy (Figure 3.12a 12c).

If the bedrock is very soft, such as the *sand-silt*, exported sediment grain size is likely to be insensitive to tectonics due to abrasion (t2 in Figure 3.12b) (c.f. Attal et al., 2015; Whittaker et al., 2010). Thus, tributaries with only soft bedrock in their catchments do not coarsen the trunk (compare T2-T3 and T3-T4 in Figure 3.12b). In many mountain areas, basement rocks exposed at their centres are often fringed by sedimentary rocks (e.g. Finnegan et al., 2017). In this case, knickpoints initiated from bounding faults would first erode the soft sedimentary rocks, whose erosional products are likely susceptible to intense grain size reduction. In such catchments, sediment export immediately after a tectonic perturbation is likely to remain fine-grained despite an increased flux (c.f. Armitage et al., 2011). Only when knickpoints migrate into hard basement rocks, are coarser sediments likely to be exported into adjacent basins (Figure 3.12b). Our results thus imply that the first export of coarser sediments to the basin may be significantly delayed after a tectonic perturbation. In the Corinth rift, for example, the time required for knickpoints to propagate from active faults into hard bedrock (i.e. the lag time) is on the million-year timescales (Zhou et al., 2024). If extrapolated to long-term stratigraphic developments in such settings, we expect a characteristic stratigraphic architecture with coarse sediment being preceded by fine sediment would develop after an increase in uplift rates (Figure 3.12d). In fact, this might have been recorded in the deltas supplied by the Sithas river, and careful field studies (e.g. Duller et al., 2019) can directly test this hypothesis. Of course, this long-term

stratigraphic pattern will be superimposed by shorter-term variations caused by autogenic processes (e.g. Muto et al., 2007), climate (e.g. McNeil et al., 2019), and tectonics (e.g. Backert et al., 2010). Still, our study represents an encouraging first step to comprehensively decompose source and sink controls on stratigraphic architectures.

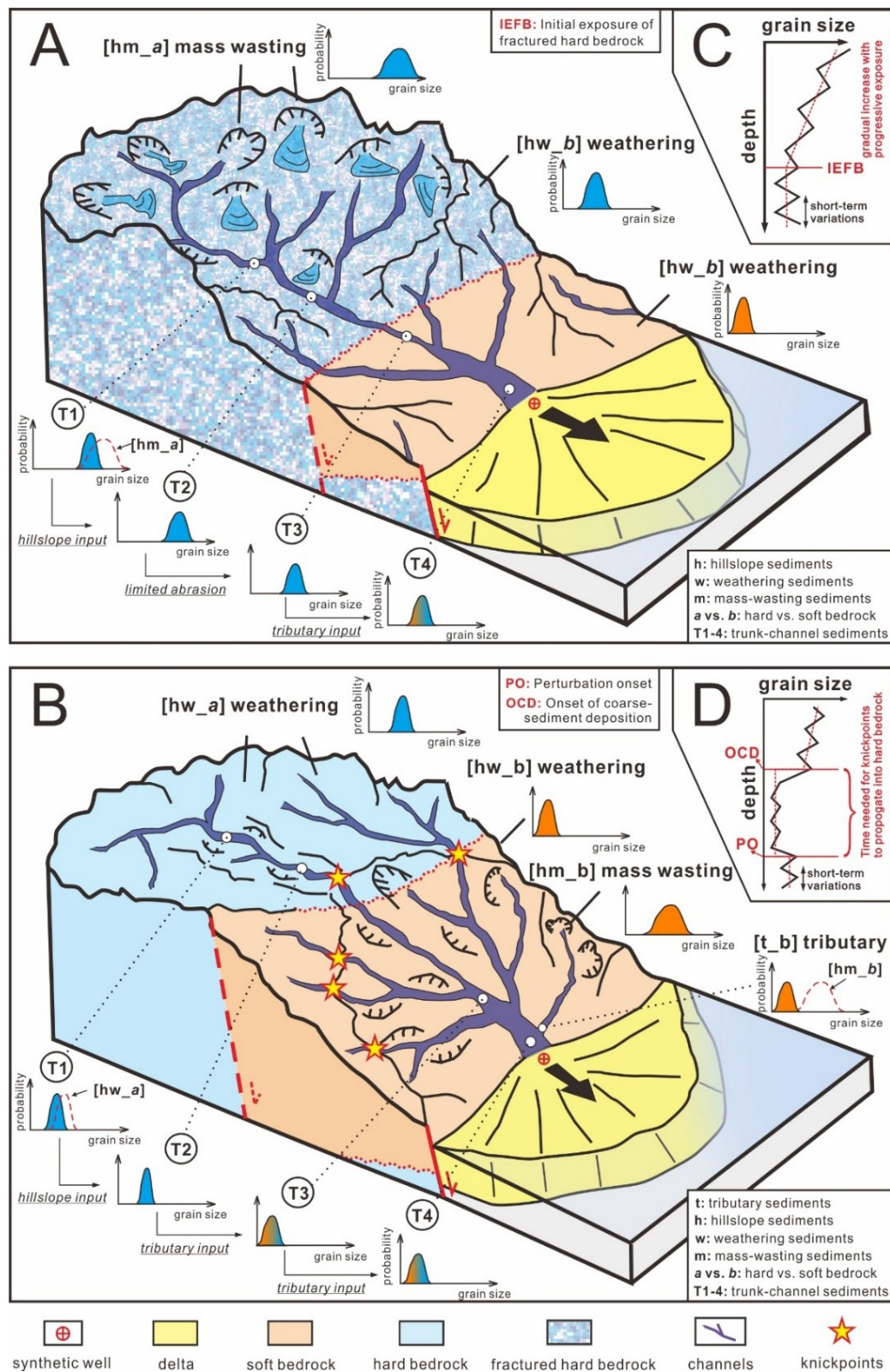


Figure 3.12. Schematic diagram illustrating process-based tectonic and lithological controls on exported sediment grain size (A, B) and their speculative implications for stratigraphic grain size evolution (C, D). In catchments where knickpoints have not yet developed, hard and fractured bedrock triggers pervasive mass wasting, and lithology-induced coarse hillslope sediment can directly coarsen trunk-channel sediment via coupled hillslopes and indirectly through tributary input (A); an upward-coarsening stratigraphy (IEFB onwards) is expected if continued incision unroofs the hard and fractured bedrock (C). In catchments where tectonic knickpoints are developed within soft bedrock, mass wasting is concentrated downstream of the knickpoints, and tectonically induced coarsening, limited by lithology-dependent abrasion, does not extend into trunk-channel sediments (B); a characteristic stratigraphy with coarse sediment (OCD onwards) being preceded by fine sediment (PO-OCD) is expected due to the time required for knickpoints to propagate into hard bedrock (D).

Finally, selective transport appears to attenuate, rather than completely obliterate, the signals imprinted in hillslope sediments (compare h1m and T1 in [Figure 3.12a](#)). This finding supports the assumption implicit in many studies that sediment grain size achieved in stratigraphy can be used to reconstruct external forcing (e.g. [Foreman et al., 2012](#); [Peizhen et al., 2001](#)). In short, we argue more upstream insights (e.g. [Whittaker et al., 2010](#)) are key to better interpret downstream stratigraphic record.

3.6 Conclusions

In this study, we focused on two rivers with well-constrained tectonic and lithological frameworks, to probe how tectonic and lithological forcings compete to control exported sediment grain size via hillslope and channel processes. The Phoenix catchment, with knickpoints in response to <100 ka faulting not yet developed, exhibits hard and mass-wasting-prone bedrock in its headwater and soft bedrock downstream. In the Sithas catchment, however, the hard bedrock in its headwater is less mass-wasting-prone and knickpoints representing >600 ka perturbations have propagated into the downstream soft bedrock.

Topographic analysis reveals that both tectonic and lithological forcings can drive hillslope and channel processes to shape landscapes. In the Sithas catchment, tectonic perturbations initiate upstream migrating knickpoints that steepen downstream hillslopes and trigger concentrated mass wasting along channels. In contrast, mass-wasting-prone bedrock in the Phoenix catchment steepens channels and hillslopes via pervasive mass wasting and rock resistance to erosion. Combined with the observations that 1) mass wasting generates coarser sediments than weathering, and 2) stronger bedrock lithologies are fragmented into coarser grains, our results show how tectonic perturbations and lithological variations can both be propagated into hillslope sediments. Upon entering the channels, hillslope sediment can undergo lithology-dependent abrasion that attenuates or even obliterates the tectonic or lithological signals. In our case, tectonic perturbations imprinted in hillslope sediments are obliterated in the Sithas catchment, as these sediments are dominated by a highly abradable *sand-silt* lithology. In addition, selective transport can attenuate but not completely

obliterate tectonic or lithological signals. This is supported by the observation that despite channel sediments in both catchments exhibit a narrower size range than hillslope sediments, channel sediments in the Sithas catchment are still systematically coarser than in the Phoenix catchment. Here, we attribute this non-obliteration effect to sediment input modulating channel entrainment by influencing channel steepness.

Finally, we argue that tectonically and lithologically induced grain size variations in exported sediments are likely to influence stratigraphic architectures in neighbouring depocenters. For instance, an upward-coarsening stratigraphy is expected, if continued incision unroofs the hard and fractured bedrock. In areas where basement rocks are fringed by soft bedrock, such as the sandstone-siltstones geological unit in our case, a characteristic stratigraphy with coarse sediment being preceded by fine sediment is expected due to the time required for knickpoints to propagate into hard bedrock.

3.7 Acknowledgements

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3.8 Supplementary material

3.8.1 Figures 3.S1-S4

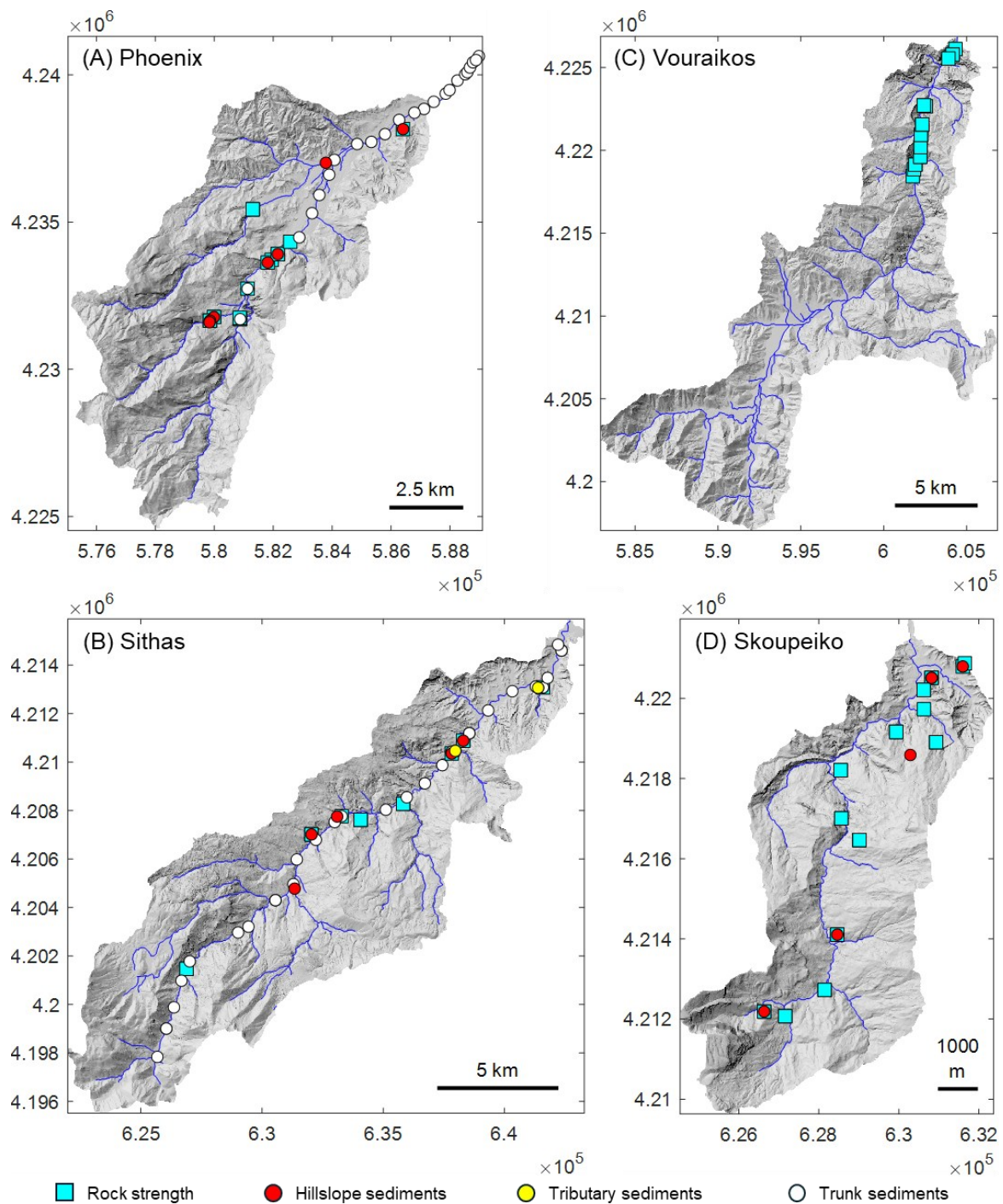


Figure 3.S1. Field localities measuring rock strength (cyan squares), tributary sediments (yellow circles), hillslope sediments (red circles), and trunk sediments (white circles) in both the Phoenix (A) and Sithas (B), as well as adjacent Vouraikos (C) and Skoupeiko (D) catchments where identical bedrock is well-exposed. See Table 3.S1 and 3.S2 for raw data.

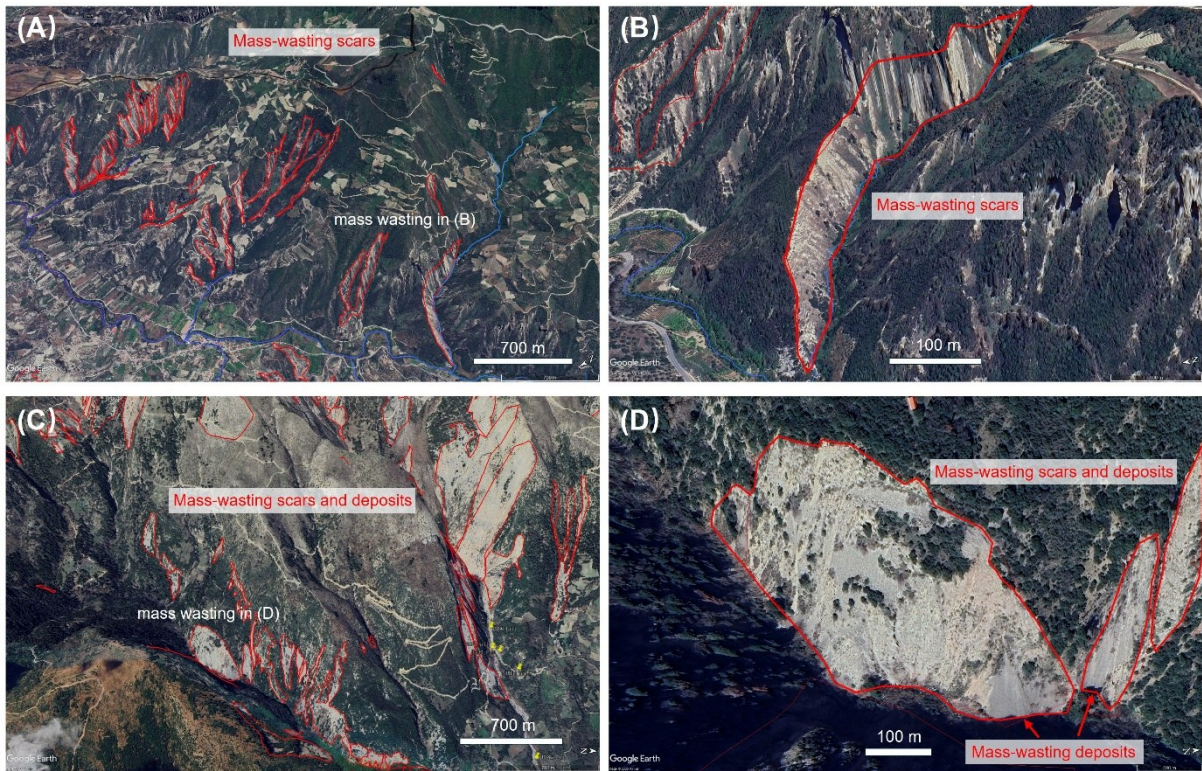


Figure 3.S2. Typical mass wasting events (red polygons) mapped from Google Earth: (A) mass wasting concentrated on sand-silt-dominated hillslopes along steepened channels in the Sithas, where tectonic knickpoints are present, (B) a zoomed-in view of the mass-wasting scars, (C) pervasive mass wasting on the lime- or flysch-dominated hillslopes in the Phoenix, and (D) a zoomed-in view of the mass-wasting scars and deposits.

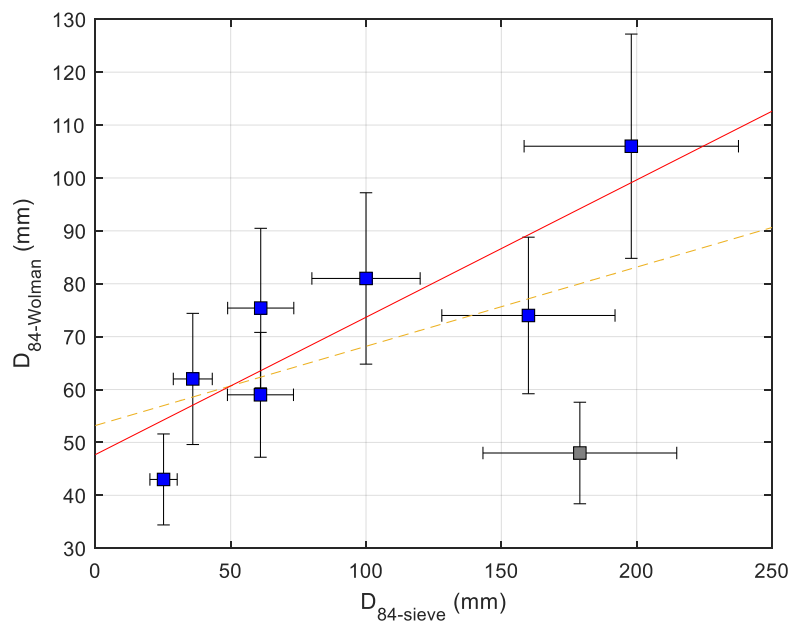


Figure 3.S3. $D_{84-Wolman}$ versus $D_{84-sieve}$. The solid and dashed red lines represent the best-fit lines, with and without considering the potential outlier (grey square). Note that the outlet localities from Watkins *et al.* (2020) for the two study rivers were also included (Table 3.S1).

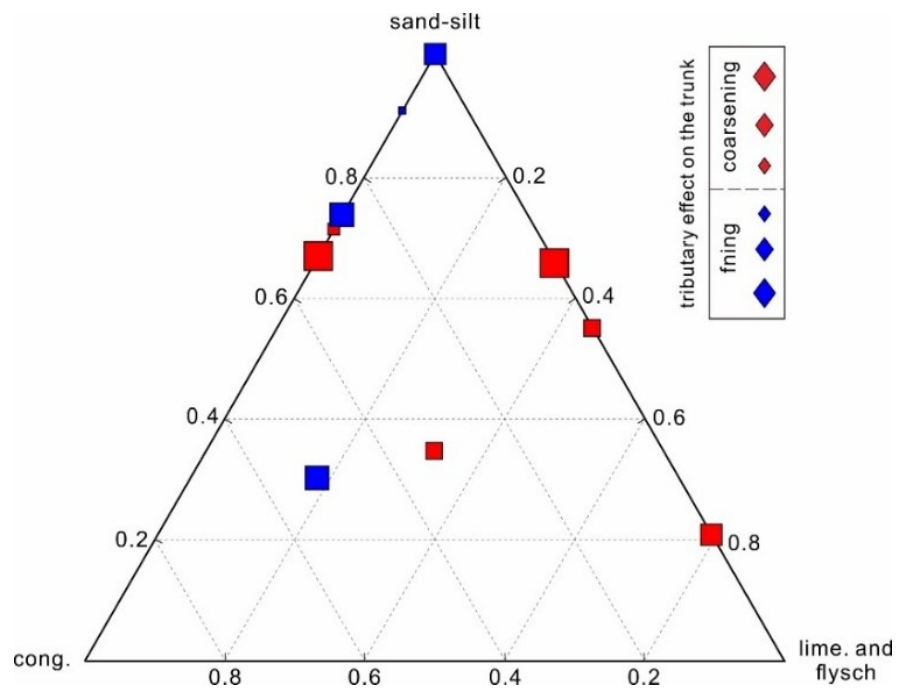


Figure 3.S4. Ternary diagram of lithological proportions in tributary catchments. Note the colour and size of the symbols indicate the impact of tributary input on trunk sediments (Table 3.S1).

3.8.2 Table 3.S1. Schmidt hammer rock strength.

River	Locality	GPS (decimal degrees)		UTM (m)		Lithology	Number of measurements	Schmidt hammer		references
		Latitude	Longitude	X	Y			MEAN	STD	
Phoenix	Ph2	38.247750	21.936528	581948.10	4233718.61	limestone	20	44	5.1	this study
Phoenix	Ph3	38.246917	21.935000	581815.35	4233624.79	chert in limestone	22	48	7.7	
Phoenix	Ph6	38.249472	21.938917	582155.20	4233911.83	limestone (more deformed)	23	41	5.7	
Phoenix		38.249472	21.938917	582155.20	4233911.83	limestone (strongly bedded)	22	57	4.0	
Phoenix	PH7	38.252139	21.943444	582578.00	4234330.00	limestone	22	48	4.5	
Phoenix	PH27	38.287333	21.988000	586405.22	4238157.65	sand_silt	22	13	2.0	
Phoenix	PH33	38.263250	21.929389	581306.15	4235432.20	limestone	22	42	8.1	
Phoenix		38.263250	21.929389	581306.15	4235432.20	limestone	22	38	8.8	
Phoenix	PH34	38.230389	21.913917	579988.57	4231772.44	flysch (bedded)	22	29	5.3	
Phoenix		38.230389	21.913917	579988.57	4231772.44	flysch (bedded)	22	34	8.9	
Phoenix	PH35	38.229306	21.912361	579853.60	4231650.89	flysch (less fractured)	22	37	7.3	
Phoenix		38.229306	21.912361	579853.60	4231650.89	flysch (very fractured)	22	29	8.1	
Phoenix	PH39	38.229667	21.924056	580876.77	4231701.11	flysch (very coarse sand)	22	41	9.5	
Phoenix		38.229667	21.924056	580876.77	4231701.11	flysch (very coarse sand)	22	41	4.1	
Phoenix	PH40	38.230083	21.923917	580864.15	4231747.23	flysch (sheared)	22	17	3.9	
Phoenix	PH43	38.238972	21.927056	581129.00	4232736.28	limestone (not in situ)	22	63	5.1	
Skoupeiko	SK1	38.125417	22.501417	631599.18	4220795.08	sand_silt	22	21	2.5	
Skoupeiko	SK1.A	38.126111	22.502000	631649.06	4220872.96	sand_silt	22	21	6.0	
Skoupeiko	SK2	38.122972	22.492472	630819.51	4220511.18	sand_silt	22	21	4.9	

Skoupeiko		38.122972	22.492472	630819.51	4220511.18	sand_silt	22	26	4.2
Skoupeiko	SK3	38.120278	22.490167	630622.22	4220208.95	sand_silt	22	17	2.6
Skoupeiko		38.120278	22.490167	630622.22	4220208.95	sand_silt	22	15	4.1
Skoupeiko	SK4	38.115944	22.490139	630627.50	4219728.07	cong. (fallen boulder cemeted by lime)	22	30	8.2
Skoupeiko	SK5	38.108500	22.493472	630933.02	4218906.71	sand_silt	22	16	2.9
Skoupeiko		38.108500	22.493472	630933.02	4218906.71	sand_silt	22	17	4.0
Skoupeiko	SK10	38.086778	22.471167	629015.57	4216465.12	limestone	22	47	5.4
Skoupeiko		38.086778	22.471167	629015.57	4216465.12	limestone	22	31	10.1
Skoupeiko		38.086778	22.471167	629015.57	4216465.12	shale in limestone	22	19	5.2
Skoupeiko	SK12	38.047444	22.449222	627159.11	4212070.35	cong (Gilbert delta, weathered)	22	35	5.9
Skoupeiko		38.047444	22.449222	627159.11	4212070.35	cong (Gilbert delta, fresch)	22	55	6.4
Skoupeiko	SK13	38.048583	22.443250	626633.10	4212188.57	cong	22	43	9.6
Skoupeiko		38.048583	22.443250	626633.10	4212188.57	cong	22	39	12.7
Skoupeiko	SK15	38.065611	22.464444	628463.06	4214107.10	sand_silt	22	23	6.9
Skoupeiko		38.065611	22.464444	628463.06	4214107.10	sand_silt	22	20	5.5
Sithas	SI1	38.054361	22.613806	641587.14	4213080.13	sand_silt	22	25	1.6
Sithas	SI3a	38.030389	22.570361	637824.19	4210351.78	sand_silt	23	22	4.8
Sithas	SI4	38.035111	22.575889	638298.08	4210881.47	sand_silt (rock fall?)	22	27	1.6
Sithas		38.035111	22.575889	638298.08	4210881.47	sand-silt (very coarse sand, in situ)	23	25	2.5
Sithas	SI7	38.001222	22.503917	632039.96	4207015.20	sand-silt (conglomerate bed)	22	12	1.4

Sithas		38.001222	22.503917	632039.96	4207015.20	sand_silt (crumbly sand bed)	22	16	2.2	
Sithas	SI9	38.007750	22.518222	633283.59	4207767.94	cong (big boulder)	22	51	5.0	

Vouraikos	65-3	38.176944	22.191186	604332.95	4226118.02	cong	22	28	6.0	Zondervan (2017)
Vouraikos	65-4	38.173912	22.189142	604158.24	4225779.29	cong	22	28	5.0	
Vouraikos		38.173912	22.189142	604158.24	4225779.29	cong	22	29	9.6	
Vouraikos	65-5	38.171851	22.186207	603904.07	4225547.30	cong	22	34	2.6	
Vouraikos	65-8	38.108019	22.160638	601752.94	4218436.15	limestone	18	64	5.2	
Vouraikos	65-10	38.111829	22.161584	601830.59	4218859.94	limestone	18	55	6.2	
Vouraikos	65-11	38.114329	22.162510	601908.29	4219138.35	limestone	22	50	5.8	
Vouraikos		38.114329	22.162510	601908.29	4219138.35	limestone	22	57	5.9	
Vouraikos	65-12	38.118608	22.165986	602207.06	4219616.98	limestone	18	58	7.0	
Vouraikos		38.118608	22.165986	602207.06	4219616.98	limestone	18	57	6.7	
Vouraikos	65-13	38.123798	22.166265	602224.27	4220193.16	limestone	18	55	2.7	
Vouraikos		38.123798	22.166265	602224.27	4220193.16	limestone	18	55	6.2	
Vouraikos	65-14	38.130480	22.166616	602245.72	4220934.98	limestone	19	64	8.9	
Vouraikos		38.130480	22.166616	602245.72	4220934.98	limestone	18	61	2.5	
Vouraikos	65-15	38.136065	22.167602	602324.33	4221555.79	limestone	18	26	10.2	
Vouraikos		38.136065	22.167602	602324.33	4221555.79	limestone	18	47	14.0	
Vouraikos	65-39	38.145977	22.170334	602549.89	4222658.63	limestone	22	46	11.4	
Vouraikos	65-40	38.146465	22.168862	602420.22	4222711.16	limestone	22	37	9.2	
Vouraikos		38.146465	22.168862	602420.22	4222711.16	limestone	22	56	6.3	
Sithas	54-12	37.954748	22.445804	626885.64	4201471.31	sand_silt	22	47	6.9	
Sithas		37.954748	22.445804	626885.64	4201471.31	sand_silt	22	33	6.4	

Sithas	54-19	38.008727	22.529225	634067.29	4207616.78	sand_silt	22	34	2.1	
Sithas		38.008727	22.529225	634067.29	4207616.78	sand_silt	22	23	3.0	
Sithas	54-21	38.014424	22.548954	635820.64	4208276.31	sand_silt	22	32	2.3	
Sithas		38.014424	22.548954	635820.64	4208276.31	sand_silt	22	35	2.3	
Skoupeiko	57-4B	38.055934	22.461958	628145.64	4212726.31	sand_silt	27	26	8.5	
Skoupeiko		38.055934	22.461958	628145.64	4212726.31	sand_silt	22	30	12.3	
Skoupeiko	57-7	38.094567	22.467776	628563.00	4217007.22	limestone	22	61	9.3	
Skoupeiko		38.094567	22.467776	628563.00	4217007.22	limestone	23	53	6.7	
Skoupeiko	57-8	38.105122	22.467471	628545.64	4218206.31	sand_silt	22	19	3.0	
Skoupeiko		38.105122	22.467471	628545.64	4218206.31	sand_silt	22	20	5.4	
Skoupeiko	57-9	38.113274	22.483613	629935.64	4219161.31	sand_silt	22	27	3.0	
Skoupeiko		38.113274	22.483613	629935.64	4219161.31	sand_silt	22	26	6.6	

3.8.3 Table 3.S2. Sediment grain size.

River	Locality	site type*	GPS (decimal degrees)		UTM (m)		Wolman		Notes (if any)
			Latitude	Longitude	X	Y	pebble counting	D84 (mm)	
Phoenix	Ph1	m	38.293444	21.996250	587119.43	4238843.49	physical	72	
Phoenix	Ph3	m	38.246917	21.935000	581815.35	4233624.79	physical	145	
		h	38.246917	21.935000	581815.35	4233624.79	physical, photo	495	limestone landslide sediments
Phoenix	Ph6	m	38.249472	21.938917	582155.20	4233911.83	physical	100	
		h	38.249472	21.938917	582155.20	4233911.83	photo	225	limestone weathering sediments
Phoenix	Ph8	m	38.254528	21.947306	582883.50	4234480.28	physical	130	
Phoenix	Ph9	m	38.261889	21.952361	583317.44	4235301.61	photo	189	
Phoenix	Ph10	m	38.267444	21.955250	583563.81	4235920.66	physical	150	
Phoenix	Ph11	m	38.273611	21.959167	583899.35	4236608.47	physical	150	
Phoenix	Ph12	m	38.282917	21.970139	584848.29	4237651.03	physical	170	
Phoenix	Ph13	m	38.283528	21.975667	585331.04	4237723.93	photo	164	
Phoenix	Ph14	m	38.285861	21.981000	585794.76	4237987.77	photo	128	
Phoenix	Ph15	m	38.290167	21.986528	586273.11	4238470.67	physical	110	
Phoenix	Ph16	m	38.292333	21.992500	586792.82	4238716.67	photo	110	
Phoenix	Ph17	m	38.295611	22.000056	587449.62	4239087.50	photo	110	
Phoenix	Ph19	m	38.298056	22.004722	587854.76	4239363.17	physical	76	
Phoenix	Ph20	m	38.299139	22.006222	587984.62	4239484.80	photo	94	
Phoenix	Ph21	m	38.302028	22.009333	588253.16	4239808.32	photo	80	
Phoenix	Ph23	m	38.304556	22.013194	588587.70	4240092.50	physical	98	
Phoenix	Ph25	m	38.307556	22.015444	588780.77	4240427.55	physical	68	
Phoenix	Ph27	h	38.287333	21.988000	586405.22	4238157.65	photo	27	sand-silts weathering sediments
Phoenix	Ph34	h	38.230389	21.913917	579988.57	4231772.44	photo	392	flysch rockfall sediments
Phoenix	Ph35	h	38.229306	21.912361	579853.60	4231650.89	photo	203	flysch weathering sediments
Phoenix	Ph37	h	38.228833	21.912167	579837.10	4231598.33	photo	91	chert scree sediments
Phoenix	Ph39	m	38.229667	21.924056	580876.77	4231701.11	physical	102	
Phoenix	Ph43	m	38.238972	21.927056	581129.00	4232736.28	physical	137	

Phoenix	69(0)	m	\	\	588884.00	4240504.00	sieve	61	the outlet site from Watkins et al. (2020)
			\	\	588884.00	4240504.00	physical	75	

Sithas	si1	m	38.054361	22.613806	641587.14	4213080.13	sieve	100	
			38.054361	22.613806	641587.14	4213080.13	physical	81	
Sithas	si2	t	38.054333	22.611500	641385.53	4213065.53	physical	45	
Sithas	si3	m	38.030750	22.570333	637814.66	4210396.02	sieve	179	
			38.030750	22.570333	637814.66	4210396.02	physical	48	
Sithas	si3a	h	38.030389	22.570361	637824.19	4210351.78	photo	47	sand-silts weathering sediments
Sithas	si4	h	38.035111	22.575889	638298.08	4210881.47	photo	65	sand-silts weathering sediments
Sithas	si5	m	38.068028	22.622778	642350.45	4214603.00	sieve	43	
			38.068028	22.622778	642350.45	4214603.00	physical	74	
Sithas	si7	h	38.001222	22.503917	632039.96	4207015.20	photo	23	sand-silts weathering sediments
Sithas	si9	m	38.007750	22.518222	633283.59	4207767.94	sieve	91	
Sithas	si12	t	38.031444	22.572056	637971.52	4210465.26	physical	58	
Sithas	si13	h	38.007667	22.516111	633099.40	4207753.84	photo	16	sand-silts weathering sediments
Sithas	si14a	h	37.981167	22.495583	631345.42	4204780.55	photo	145	the conglomerate weathering sediment on hillslopes is used to represent the tributary for their proximity (see more in the main text)
Sithas	54(2)	m	38.005664	22.514868	632994.92	4207526.16	physical	47	
Sithas	54(3)	m	38.009924	22.539025	635108.03	4208033.67	physical	104	
Sithas	54(4)	m	38.019493	22.557553	636716.89	4209122.55	physical	84	
Sithas	54(5)	m	38.031108	22.570487	637830.50	4210430.47	physical	47	
Sithas	54(6)	m	38.046208	22.587884	639328.78	4212131.95	physical	42	
Sithas	54(7)	m	38.054769	22.610432	641290.92	4213115.96	physical	57	
Sithas	54(8)	m	37.967291	22.473623	629441.11	4203210.09	physical	72	
Sithas	54(9)	m	37.965185	22.468712	629013.40	4202969.60	photo	21	
Sithas	54(10)	m	37.937755	22.438043	626366.24	4199883.92	photo	75	
Sithas	54(11)	m	37.947590	22.441771	626676.97	4200980.27	physical	57	

Sithas	54(12)	m	37.954748	22.445804	627019.01	4201780.02	physical	86	
Sithas	54(13)	m	37.919411	22.429917	625683.38	4197837.50	physical	55	
Sithas	54(14)	m	37.929891	22.434324	626052.86	4199006.30	sieve	198	
			37.929891	22.434324	626052.86	4199006.30	physical	106	
Sithas	54(15)	m	37.976970	22.486476	630553.01	4204302.03	sieve	36	
			37.976970	22.486476	630553.01	4204302.03	physical	62	
Sithas	54(16)	m	37.982850	22.495053	631295.87	4204966.55	physical	83	
Sithas	54(17)	m	37.991994	22.496814	631434.22	4205983.66	physical	78	
Sithas	54(18)	m	37.999185	22.505918	632220.76	4206794.49	photo	55	
Sithas	54(20)	m	38.026136	22.565890	637436.32	4209871.96	physical	45	
Sithas	54(21)	m	38.014424	22.548954	635971.42	4208547.47	physical	64	
Sithas	54(22)	m	38.037855	22.578829	638549.95	4211191.54	sieve	61	
			38.037855	22.578829	638549.95	4211191.54	physical	59	
Sithas	54(23)	m	38.053198	22.599368	640323.16	4212924.86	photo	38	
Sithas	54(24)	m	38.057916	22.616059	641778.56	4213473.73	photo	48	
Sithas	54(25)	m	38.070199	22.621245	642209.78	4214844.63	physical	47	
Sithas	54(0)	m	\	\	#####	#####	sieve	25	the outlet site from Watkins et al. (2020)
							physical	43	

Skoupeiko	sk1a	h	38.125417	22.501417	631599.18	4220795.08	photo	36	sand-silts weathering sediments
Skoupeiko	sk2	h	38.122972	22.492472	630819.51	4220511.18	photo	123	sand-silts weathering sediments
			38.122972	22.492472	630819.51	4220511.18	photo	190	sand-silts landslide sediments
Skoupeiko	sk7	h	38.105722	22.486056	630287.69	4218588.04	photo	68	sand-silts weathering sediments
Skoupeiko	sk13	h	38.048583	22.443250	626633.10	4212188.57	physical, photo	67	conglomerate weathering sediments
Skoupeiko	sk15	h	38.065611	22.464444	628463.06	4214107.10	photo	20	sand-silts weathering sediments

Note 'h' refers to hillslope sediments, 'm' refers to trunk-channel sediments, and 't' refers to tributary sediments.

4 SEDIMENT SUPPLY CONTROLS ON CHANNEL MORPHOLOGICAL ADJUSTMENTS TO TECTONICS AND LITHOLOGY ¹

Abstract

In nature, rivers not only incise bedrock but also transport sediment supplied to them; however, incision is often assumed to dominate over sediment transport in shaping channel geometry so as to exploit the simplicity of the stream power law. This likely flawed assumption raises fundamental concerns about inverting channel morphology to map external forcing, such as active deformation. Central to this issue is how relative sediment flux (i.e. sediment supply relative to transport capacity, Q_s/Q_t) modulates the efficacy of excess shear stress in incising bedrock, known as the relative sediment flux function. Two competing functions have been proposed, but resolving them in the field has proven challenging to date. Here, we address this issue by contrasting two rivers in the Gulf of Corinth, Greece that traverse comparable gradients in tectonics and lithology with distinct relative sediment fluxes. We show that the sediment-rich Phoenix river, with an estimated Q_s/Q_t value ~ 0.8 , has a much lower sensitivity of excess shear stress to tributary sediment input than the sediment-poor Sithas river. To our knowledge, this represents the first field evidence supporting the function with a markedly decreased sensitivity to Q_s/Q_t at high Q_s/Q_t values. This allows us to clarify the importance and mechanisms of channel slope versus width adjustment in reaching the excess shear stress required for equilibrium. As a result, we outline the conditions under which the widely used stream power law and its associated metrics, such as steepness and knickpoints, can (or cannot) be sensibly employed.

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4.1 Introduction

In non-glaciated upland areas, bedrock rivers are the main geomorphic engines that transmit tectonic, climatic and lithological forcings into landscapes (e.g. Allen, 2008; Whipple et al, 2013). In doing so, these rivers not only incise bedrock but also transport sediment supplied to them (e.g. Whipple and Tucker, 2002; Finnegan et al., 2017). However, in most studies that invert channel morphology to map active deformation (e.g. Kirby et al., 2008), probe climatic variations (e.g. Leonard et al., 2023), and infer lithological controls (e.g. Gallen et al., 2013), channel dynamics are assumed to be dominated by incision (i.e. detachment-limited erosion). This assumption exploits the simplicity of the stream power law (SPL), in which incision is expressed as a function of drainage area and slope, both of which are readily available (Whipple and Tucker, 1999). Yet, this appealing assumption is likely flawed in most rivers that carry significant sediment (e.g. Whipple and Tucker, 2002). The expected importance of sediment supply at best complicates, and at worst may invalidate, the inferences drawn from prior studies using the SPL (Yanites, 2018). It is thus of fundamental importance to elucidate the extent to which sediment supply modulates long term channel dynamics in real rivers.

Sediments moving through channels have long been hypothesised to modulate incision and channel adjustment to tectonic forcing (Equation 4.1, e.g. Sklar and Dietrich, 2006). If bedload sediment supply (Q_s) is low relative to channel transport capacity (Q_t), moving sediment tends to abrade and loosen bedrock to enhance incision (e.g. Sklar and Dietrich, 2004); in contrast, excessive relative sediment flux (Q_s/Q_t) covers the bedrock and inhibits incision (e.g. Johnson et al., 2009). This is often known as the “tools and cover” effect (Figure 4.1), which has been examined in theoretical (e.g. Turowski et al., 2007), experimental (e.g. Sklar and Dietrich, 2001; Johnson and Whipple, 2010; Chatanantavet and Parker, 2008), modelling (e.g. Yamanishi and Naruse, 2026; Zhang et al., 2020), and some field perspectives (e.g. Cook et al., 2013; Jansen et al., 2011; Johnson et al., 2009). Yet, most field studies to date have focussed on the qualitative effects of sediment flux, leaving its quantitative function ($f(Q_s/Q_t)$) poorly constrained (Figure 4.1). An important exception is Cowie et al. (2008), who supported a hump-shaped function for relative sediment flux in governing the erosional history of channels over timescales $>10^5$ years, based on well-constrained rivers crossing active faults in Italy and Greece. These authors ruled out an “undercapacity model” for governing long-term river incision (Beaumont et al., 1992), but their data did not resolve between published sediment flux functions that include tools and cover terms (Figure 4.1, Sklar and Dietrich, 2004; Turowski et al., 2007).

To address this fundamental issue, we contrast two well-constrained rivers draining into the Gulf of Corinth, Greece. Both rivers cross marked gradients in bedrock erodibility and uplift rate (e.g. Zhou et al., 2024), and have good constraints on sediment supply (e.g. McLeod et al., 2024). We characterise

the contrasting responses of these two rivers with distinct sediment fluxes to comparable tectonic and lithological gradients. In particular, the sensitivity of excess shear stress to tributary input is used to resolve competing relative sediment flux functions (c.f. Sklar and Dietrich, 2004; Turowski et al., 2007). In this context, we clarify the competing importance of channel slope versus width in reaching the excess shear stress required for equilibrium. The results give new insights into sediment flux-dependent river incision and help clarify when and why the SPL can be reasonably used for rivers carrying some or abundant sediment.

4.2 Flux-dependent incision

In general, flux-dependent incision can be modelled using Equation 4.1:

$$E = Kf(Q_s/Q_t)(\tau_b - \tau_c)^a, \quad (4.1)$$

where E is incision rate, K is erodibility, $f(Q_s/Q_t)$ describes the “tools and cover” effect of relative sediment flux (i.e. Q_s/Q_t), τ_b is bankfull shear stress, τ_c is critical shear stress for incipient sediment motion, and a is a constant associated with the incision process (Hobley et al., 2011). The relative sediment flux function $f(Q_s/Q_t)$ can be interpreted as modulating the efficacy of excess shear stress (i.e. $\tau_b - \tau_c$) in incising bedrock (hereafter termed *erosional efficacy*, Johnson et al., 2009). Notably, the SPL ($E = KA^mS^n$, where A is drainage area, S is channel slope, and m and n are constants) can also be stated in terms of shear stress as $E = K(\tau_b - \tau_c)^a$ under common hydraulic scaling assumptions (Whipple and Tucker, 1999). In this sense, the SPL represents a special case of flux-dependent incision, in which $f(Q_s/Q_t)$ is a constant.

The relative sediment flux function is central to constraining river incision, but its specific form remains contentious (Figure 4.1, e.g. Sklar and Dietrich, 2006). The basic form was first conceptualised (Sklar and Dietrich, 2004) and later validated by experimental studies (e.g. Johnson and Whipple, 2010) to depend on both sediment supply and the fraction of exposed bedrock (F_e). In Sklar and Dietrich (2004)’s conceptual model, the fraction of exposed bedrock is assumed to be linearly proportional to relative sediment flux (i.e. $F_e = 1 - Q_s/Q_t$), supported by flume experiments by Chatanantavet and Parker (2008). This results in a parabolic curve for $f(Q_s/Q_t)$ (Figure 4.1), with the rising limb (tools domain) and falling limb (cover domain) being symmetrical with respect to the $Q_s/Q_t = 0.5$ axis (Equation 4.2).

$$f(Q_s/Q_t) = Q_s(1 - Q_s/Q_t), \quad (4.2)$$

Based on a probabilistic argument, however, Turowski et al. (2007) derived an exponential relationship between the fraction of exposed bedrock and relative sediment flux (i.e. $F_e = e^{-\phi \frac{Q_s}{Q_t}}$,

where φ represents the difference in the probability of excess sediment accumulation per unit bed area between exposed and covered areas). In this exponential relationship, the rate of decrease in the fraction of exposed bedrock diminishes as Q_s/Q_t increases. This leads to an asymmetrical $f(Q_s/Q_t)$ curve (Figure 4.1), featuring a high- Q_s/Q_t range in which erosional efficacy becomes increasingly insensitive to Q_s/Q_t variations (Equation 4.3).

$$f(Q_s/Q_t) = Q_s e^{-\varphi \frac{Q_s}{Q_t}}, \quad (4.3)$$

This equation is derived for channels with $Q_s < Q_t$, in which all sediment is mobile, but dynamic cover can occur as sediment concentration increases, inhibiting the number of grain-bed collisions by increasing the frequency of grain-grain collisions (Turowski et al., 2007).

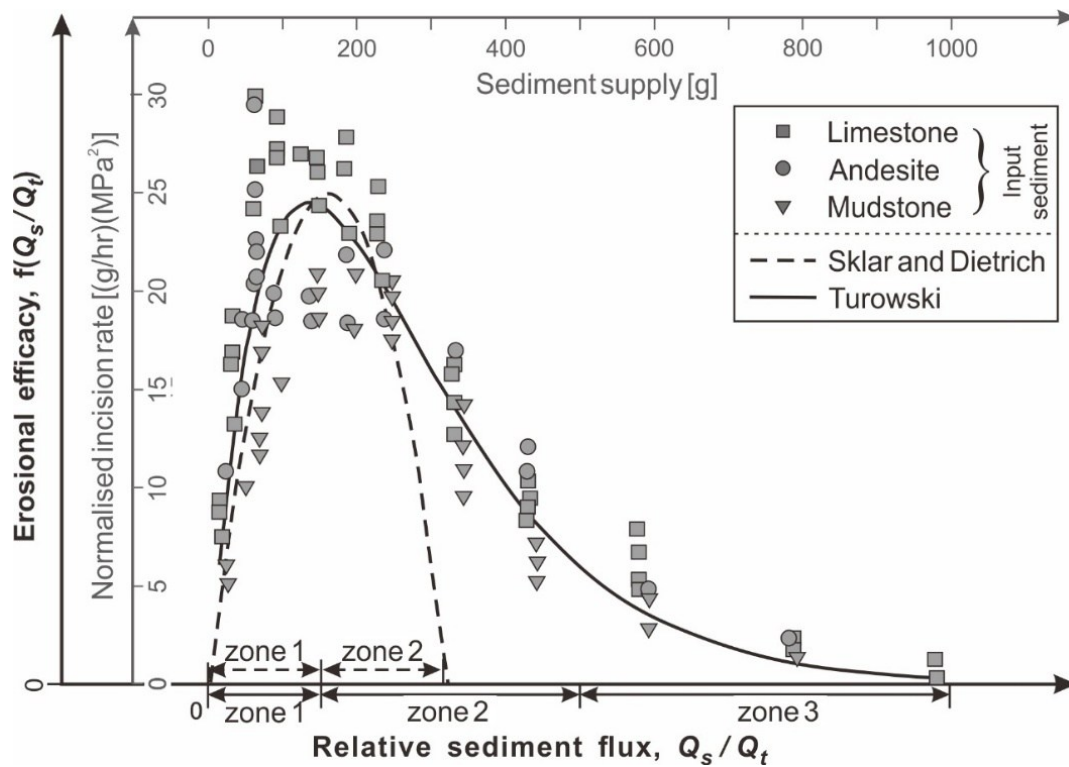


Figure 4.1. Two competing, hump-shaped relative sediment flux functions—dashed (Sklar and Dietrich, 2004) and solid (Turowski et al., 2007)—are fitted to short-term experimental data from an erosion mill, in which sediments of different lithologies are input to abrade an underlying bedrock disk (scatters on grey axes; Sklar and Dietrich, 2001). Here, Turowski et al.'s (2007) function fits the scatters better than Sklar and Dietrich's (2004), but its significance is limited by differences in timescale and hydrodynamics between the experimental mill and natural rivers. Identifying the correct function is crucial, as the two predict distinct river dynamics (see main text). Note the Q_s/Q_t range can be conceptually divided into zones 1, 2, and 3, based on the gradient or sensitivity of the function to Q_s/Q_t variations.

Even though experimental data can help constrain the form of $f(Q_s/Q_t)$ (Figure 4.1; e.g. Sklar and Dietrich, 2004; Chatanantavet and Parker, 2008; Johnson and Whipple, 2010), their significance is limited by the extent to which the short-term erosion mill or flume experiments capture long-term

river dynamics. To our knowledge, field evidence from natural rivers has yet to be found to resolve between the two formulations (e.g. [Hobley et al., 2011](#)). This is of fundamental importance, as the two functions predict distinct incision dynamics, partly due to differences in erosional efficacy gradients across the Q_s/Q_t range. Up to three zones occur as a function of relative sediment flux: 1) a steep positive gradient (zone 1), 2) a steep negative gradient (zone 2), and 3) a shallow negative gradient (zone 3) ([Figure 4.1](#)). Notably, zone 1 and zone 2 exist in both functions, but zone 3 is unique to [Turowski et al.'s \(2007\)](#) function, where the increasingly shallow gradient of the curve allows erosional efficacy to level off and therefore be effectively subsumed into erodibility K , whose dimensions are model-dependent ([Equation 4.1](#)). Thus, the existence of zone 3 in natural rivers may obscure the expected distinction between flux-dependent incision and the SPL ([Whipple and Tucker, 1999](#)). In addition, the two functions also differ in the Q_s/Q_t values at which zone 1 transitions into zone 2: this at 0.5 in [Equation 4.2](#) for [Sklar and Dietrich \(2004\)](#)'s formulation and at $1/\varphi$ in [Equation 4.3](#) for [Turowski et al. \(2009\)](#)'s. The value of φ is unknown, but the experiments suggest the transition occurs at $Q_s/Q_t \sim 0.2$, based on the assumption that incision ceases when $Q_s = Q_t$ (i.e. $Q_t = 1000$ g in [Figure 4.1](#)). This transition point holds important implications for predicting incision rates in flux-dependent rivers ([Turowski et al., 2007](#)). In turn, the key distinctions described above provide the most effective means of resolving the two competing functions ([Figure 4.1](#)).

4.3 Study area

The Phoenix and Sithas are two rivers draining into the Gulf of Corinth, Greece, one of the fastest spreading active rifts in Europe ([Figure 4.2](#); [Bell et al., 2008](#)). The Phoenix river crosses the onshore Selianitika and Erineos-Lampiri normal faults with a Late Quaternary footwall uplift rate of 1–2 mm/yr ([Figure 4.3c](#); [Palyvos et al., 2010](#)). The Sithas river is bounded by the offshore East Xylokastro fault, which has a comparable Late Quaternary uplift rate of 1.3–1.6 mm/yr ([Figure 4.3d](#); [Armijo et al., 1996](#); [de Gelder et al., 2019](#)). The back-tilting of these fault blocks causes strike-perpendicular uplift rates to decline systematically, as evidenced by marine terraces (e.g. [Armijo et al., 1996](#)) and landscape characteristics (e.g. [Fernández-Blanco et al., 2020](#)). A linear decay model with a rate of $\sim 0.08 \text{ mmyr}^{-1} \text{ km}^{-1}$ broadly reproduces the strike-perpendicular uplift rate variations in Corinth ([de Gelder et al., 2019](#); [Zondervan et al., 2020](#)).

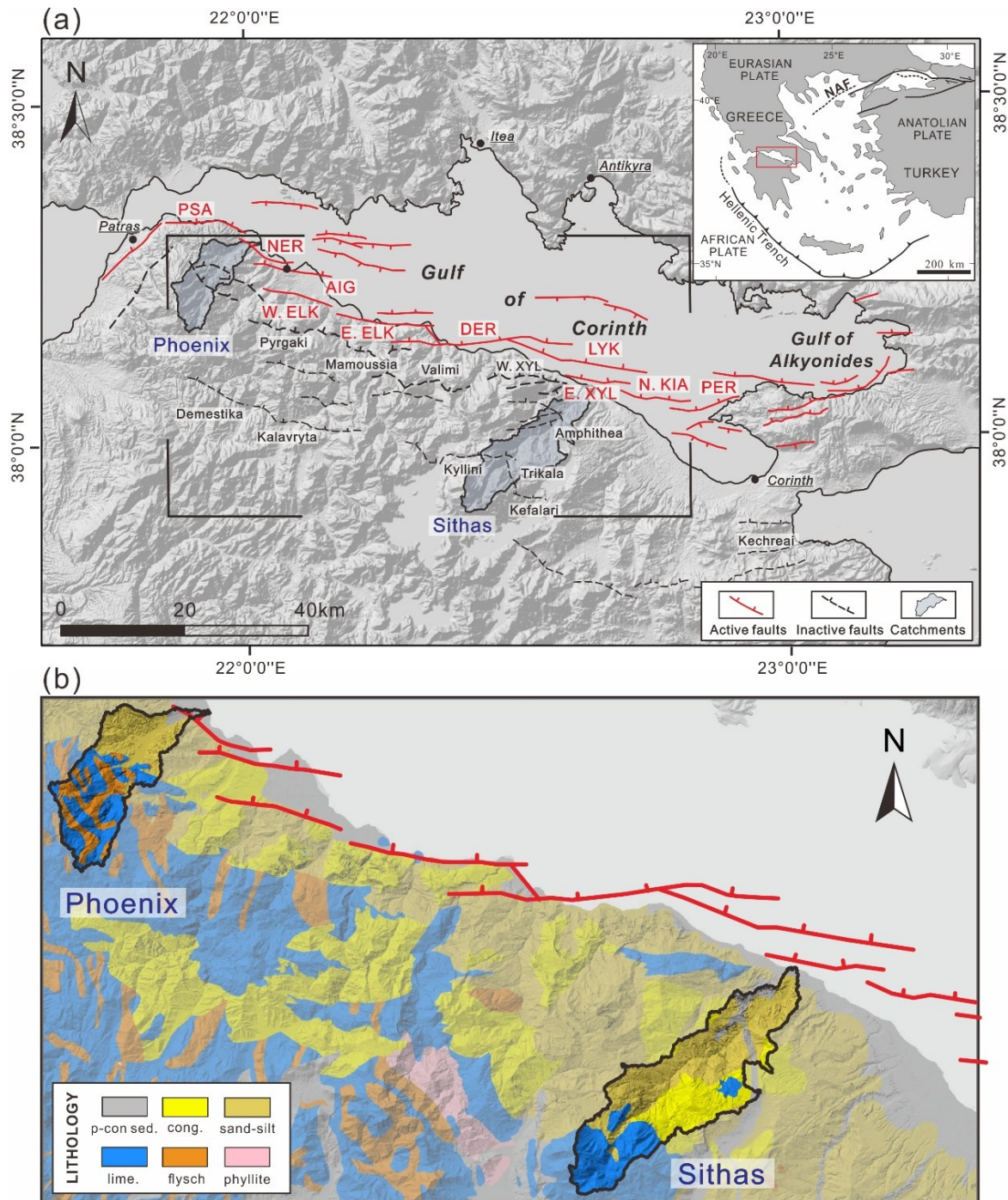


Figure 4.2. Gulf of Corinth in maps of hillshaded relief (a), and lithology (b), with active faults (red lines) and the two study catchments highlighted. Abbreviations: p-con sed. – poorly consolidated sediment; cong. – conglomerate; sand-silt – sandstone-siltstone; lime. – limestone. AIG-Aigion fault, DER-Deveni fault, E. ELK-East Eliki fault, E. XYL-East Xylokaastro fault, LYK-Lykopia fault, N. KIA-North Kiato fault, PER-Perachora fault, PSA-Psathopyrgos fault, LNER-Neo Erineos fault, W. ELK-West Eliki fault.

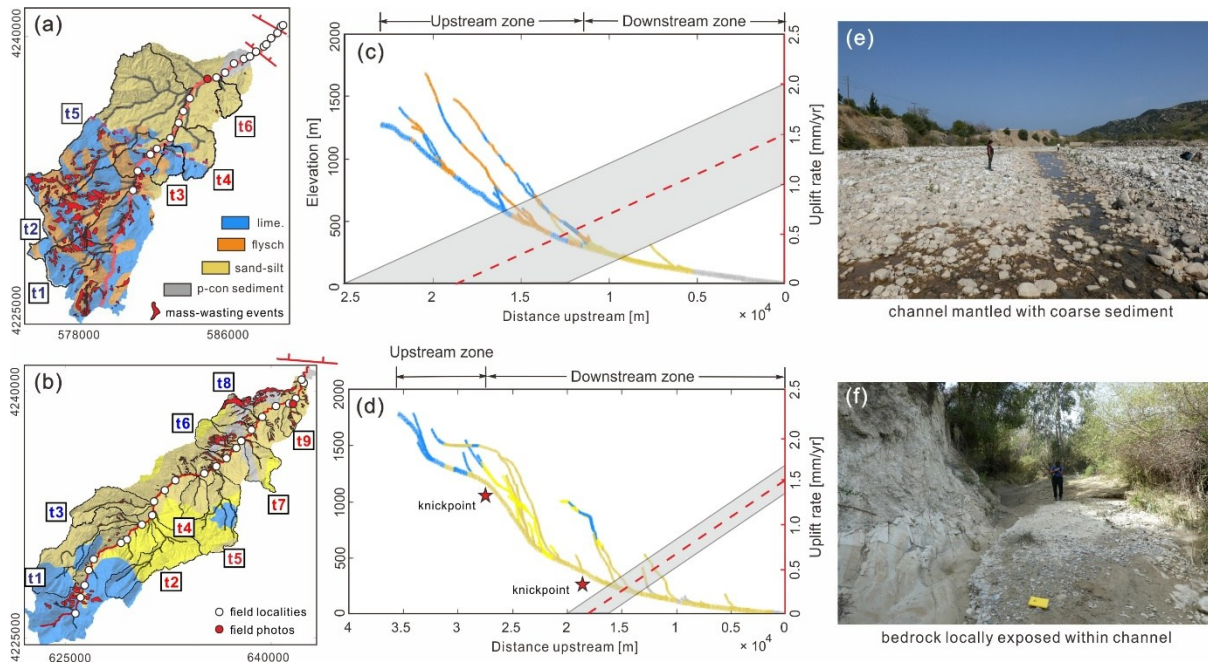


Figure 4.3. Main characteristics of the Phoenix (a, c, e) and Sithas (b, d, f) rivers: (a, b) trunk and tributary channels cutting across different lithologies and normal faults; (c, d) trunk-channel river profiles, coloured by lithology from Figure 4.2, shown with first-order variations in uplift rates. Dashed red lines and grey bands represent the mean and range of fault uplift rate constraints, with their upstream decrease based on the linear back-tilting model derived from two well-constrained faults in Corinth (see main text); (e, f) low-flow stage field observations (see locations in a and b). Note both rivers exhibit the transition between the upstream zone, where sediment input comes directly from adjacent coupled hillslopes, and the downstream zone, where tributary input dominates. The upstream-downstream transition coincides with the boundary between hard and soft bedrock in the Phoenix river and with a knickpoint in the Sithas river.

Footwall rivers, such as the Phoenix and Sithas, experience a comparable downstream increase in uplift rate (Figure 4.3c-d). The Phoenix river does not develop knickpoints in response to the onshore faults, but in the vicinity of the faults, the channel appears to have shifted toward alluvial conditions due to back-tilting (Figure 4.3c). In contrast, the Sithas river is bounded by the older offshore fault and has developed two tectonic knickpoints (Figure 4.3d, Zhou et al., 2024). Both rivers also traverse a similar bedrock erodibility gradient, from pre-rift basement upstream to syn-rift stratigraphy downstream (Figure 4.2b, Zhou et al., 2024). The pre-rift basement, including limestone, flysch, and phyllites-quartzites (e.g. Wicker et al., 2024) is documented to be less erodible than the syn-rift stratigraphy comprising conglomerate, sandstone-siltstone, and poorly consolidated sediments (Figure 4.2b, e.g. Zhou et al., 2024). In particular, limestone, flysch, and conglomerate, generate most of the bedload sediment, whereas sandstone-siltstone is rapidly abraded into sand-sized particles (Zhou et al., 2026).

Consequently, while Phoenix and Sithas rivers cross marked yet comparable gradients in bedrock erodibility and uplift rate (Figure 4.3), both qualitative and quantitative observations indicate distinct

bedload sediment supply. For example, [Deiss et al. \(2025\)](#) reported a 100-kyr bedload sediment budget for the Sithas of $\sim 25\text{km}^3/\text{Myr}$, based on mapping of a deep-water fan offshore. This translates to $\sim 2\text{kg/s}$ using a mean sediment density of 2650 kg/m^3 and is several orders of magnitude lower than typical bankfull transport capacity (e.g. [McLeod et al., 2024](#)). As such, the Sithas river likely lies within the tools domain. Field observations show that during low-flow stages, coarse sediments only partially cover the Sithas channel ([Figure 4.3f](#)), yet they fully mantle the Phoenix channel ([Figure 4.3e](#)), suggesting considerable sediment supply in the latter catchment. This aligns with published findings that mass wasting, the main bedload generation process, is more pervasive in the Phoenix river than in the Sithas river ([Figure 4.3a-b](#), [Zhou et al., 2026](#)). In addition, based on seismic and field data, [McLeod et al. \(2024\)](#) estimated the Holocene bedload volume and bankfull sediment transport capacity for both rivers. Their results indicated that over the last 12 kyr, the Q_s/Q_t of the Phoenix river was ~ 2.5 times that of the Sithas river ([their figure S6c](#)), consistent with our qualitative field observations. In short, the two rivers exhibit distinct relative sediment fluxes, with the Sithas and Phoenix rivers likely falling in the tools and cover domains, respectively ([Figure 4.1](#)).

The study rivers thus represent a unique opportunity to resolve between two competing relative sediment flux functions ([Figure 4.1](#)). However, directly determining bankfull Q_s/Q_t downstream in rivers is non-trivial over long periods. This is because we typically lack hydrographs over 100-kyr timescales to constrain long-term bankfull flood intermittency and sediment transport capacity. Thus, we leverage variations in excess shear stress, and turn to the defining feature of zone 3, if present, i.e. its markedly lower sensitivity to Q_s/Q_t variations compared to the other two zones ([Figure 4.1](#)) to reconstruct how these rivers are influenced by sediment flux dependent incision. In addition, available shorter-term Q_s/Q_t estimates introduced above are presented as corroborative evidence.

4.4 Methods

To investigate sediment supply controls on river incision, we analysed channel morphological responses to variations in tectonics, lithology and sediment supply. For each river, we: i) systematically measured bankfull channel geometry; ii) evaluated excess shear stress responses to expected Q_s/Q_t variations across tributary confluences; and iii) constrained Q_s/Q_t values.

4.4.1 Bankfull channel geometry

Bankfull channel geometry, including width, slope, and depth, represents the main variable rivers adjust to accommodate variations in Q_s/Q_t , bedrock erodibility and uplift rates. These measurements are also essential to quantify excess shear stress and sediment transport capacities. In the field, we used flood markers, such as bleaching surfaces, vegetation boundaries, and debris to estimate bankfull stage ([Figure 4.4a](#), e.g. [Whittaker et al., 2007b](#)). Then, channel width was measured as the distance

between the bounding channel banks (Figure 4.4b) using a laser rangefinder. Slope was determined in degrees over ~30 m along the river (Figure 4.4c). Depth was measured as the elevation differences from thalweg points to flood markers (Figure 4.4d). Errors associated with these measurements are provided in [supplementary methods](#).

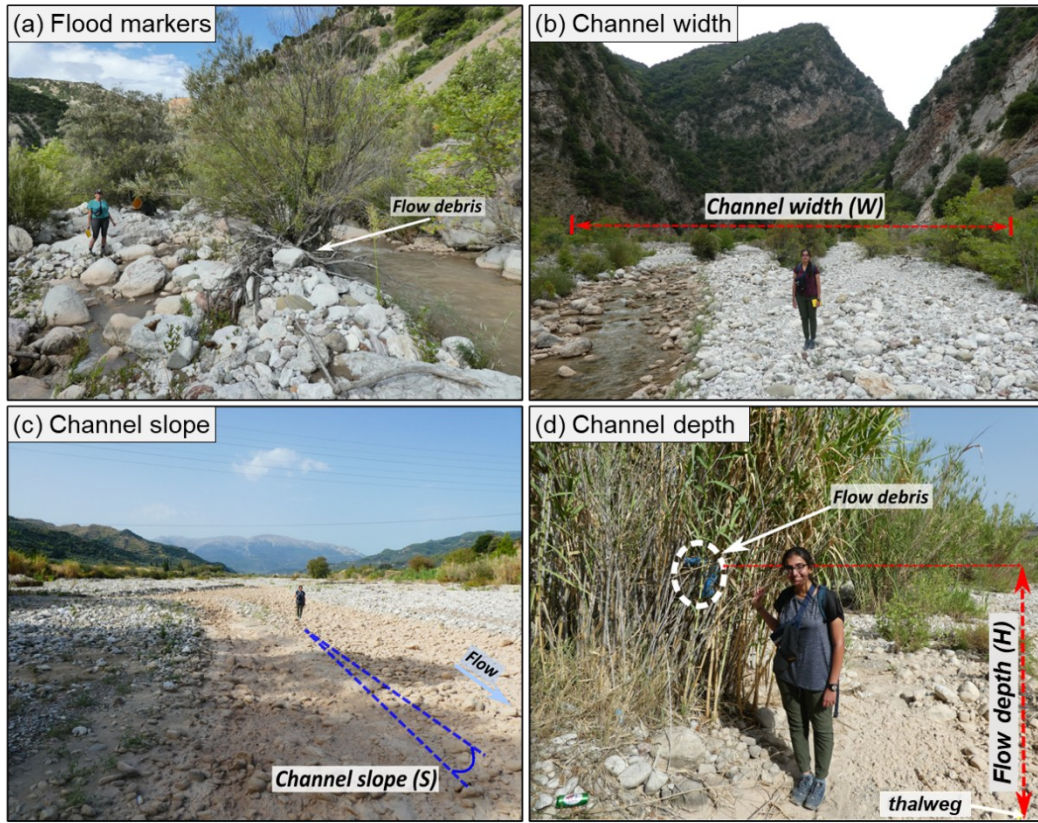


Figure 4.4. Field photos illustrating bankfull flood markers (A), width (B), slope (C), and depth (D).

To characterise channel responses to marked variations in lithology and tectonics, bankfull width, slope, and width: depth (W/D) were plotted along the trunk channel in both rivers (Figure 4.5a-d). Because channel width and slope scale with drainage area, area-normalised slope (steepness, k_{sn}) (Equation 4.4) and width (wideness, k_{wn}) (Equation 4.5) were introduced to account for area-induced variations:

$$k_{sn} = SA^b, (4.4)$$

$$k_{wn} = WA^{-c}, (4.5)$$

where S is the channel slope, A is the upstream drainage area, W is the channel width, b and c are exponents relating drainage area to slope and width, respectively (e.g. Allen et al., 2013). In our calculations, b and c were set as 0.5 for both rivers to enable direct comparison (e.g. Yanites, 2018), with uncertainties propagated. In this study, we compared the relative variability of wideness and steepness to infer the dominant mechanism (width versus slope adjustment) a river uses to

accommodate gradients in bedrock erodibility and uplift rate: slope adjustment prevails if steepness is more variable; alternatively, width adjustment dominates.

4.4.2 Excess shear stress and tributary input

To test the hypothesis that the sediment-rich Phoenix river may lie in zone 3 (Figure 4.1), the sensitivity of erosional efficacy to Q_s/Q_t variations was examined. Erosional efficacy cannot be measured directly, but for channels that are in erosional equilibrium with identical uplift rates and erodibility, it is inversely related to excess shear stress (Equation 4.1). Thus, in such channel reaches, the objective can be reframed as examining the sensitivity of excess shear stress to Q_s/Q_t variations. Excess shear stress was calculated here as bankfull shear stress minus critical shear stress, both of which can be derived directly from our bankfull channel geometry measurements (supplementary methods).

In both rivers, the downstream zones are point-sourced by tributary inputs (Figure 4.3; Zhou et al., 2026), making tributary confluences an ideal context for investigation. This is because the trunk channel Q_s/Q_t may change markedly across confluences due to tributary sediment input, while uplift rates and lithology remain effectively constant over such short distances. Between confluences, uplift rates are expected to increase towards the faults, but we make the simplifying assumption that Q_s/Q_t is approximately spatially uniform as sediment and water discharge is not expected to change. From the perspective of sediment mixing, the impact of tributary input on the trunk channel's Q_s/Q_t depends mostly on their differences in Q_s/Q_t and drainage area. The latter is readily available, but the former is hard to constrain directly. Here, we leveraged the published finding that bedload sediment in our study rivers is dominantly generated from limestone, flysch, and conglomerate, as the softer sandstone–siltstone is documented to be rapidly abraded into sand-sized particles and therefore does not contribute to bedload (Zhou et al., 2026). This allows us to use the combined area of limestone, flysch, and conglomerate, relative to the total drainage area (bedload-generating-lithology proportion), as a qualitative proxy for relative sediment flux. Note, this proxy is not used here to quantify Q_s/Q_t along the trunk channels; rather, it is used in conjunction with drainage area to assess which tributaries are most or least likely to change the trunk-channel Q_s/Q_t at confluences. Specifically, tributaries with distinct bedload-generating-lithology proportions and large drainage areas relative to the trunk-channel are most likely to change the trunk-channel Q_s/Q_t at confluences (supplementary Table 4.S2).

Consequently, we plotted excess shear stress (supplementary methods) along the trunk-channel profiles for both rivers, with the positions of tributary confluences shown (Figure 4.6). First, we examined excess shear stress variations between confluences to test whether the surveyed channels are in equilibrium with the gradients in uplift rate and erodibility. In this context, we then compared

the magnitude of excess shear stress variations with that of the expected Q_s/Q_t changes (supplementary Table 4.S2) across confluences to assess the sensitivity of excess shear stress to Q_s/Q_t changes: if a tributary expected to significantly disturb the trunk-channel Q_s/Q_t does not alter excess shear stress, the erosional efficacy of the river is considered insensitive to Q_s/Q_t variations (zone 3). In contrast, if a tributary assessed to have limited impact on the trunk-channel Q_s/Q_t corresponds to a marked change in excess shear stress, the erosional efficacy of the river is considered highly sensitive to Q_s/Q_t variations (zone 1 or zone 2).

4.4.3 Q_s/Q_t estimates

The interpretation drawn from the sensitivity analysis can be complemented by our independent constraints on bankfull Q_s/Q_t values of the two rivers. We exploited the observation that the Phoenix river appears to transition to dominantly alluvial conditions near the onshore active faults (Zhou et al., 2026). In transport-limited alluvial reaches, excess shear stress is, by definition, all spent to transport sediment (i.e. $Q_s = Q_t$), and transport capacity can be estimated using the Meyer-Peter and Müller equation (supplementary methods). Thus, we estimated maximum Q_s from the dominantly alluvial channels (transparent points in Figure 4.5a, c, e) and Q_t from channels adjacent to those dominantly alluvial ones (supplementary methods and Table 4.S1). The Q_s/Q_t can be therefore constrained for the Phoenix river and then extrapolated to the Sithas river, based on its ~2.5-fold difference over the last 12 kyr (McLeod et al., 2024) and the likely equal number of bankfull floods in the two adjacent rivers.

4.5 Results

4.5.1 Bankfull channel geometry

Figure 4.5 shows downstream variations in bankfull width, slope, and W/D , as the Phoenix and the Sithas rivers traverse comparable gradients in bedrock erodibility and uplift rate (Figure 4.3) with distinct sediment fluxes. In the sediment-rich Phoenix river, channels first widen markedly from the hard-bedrock dominated upstream zone (~15m) to the soft-bedrock dominated downstream zone (~90 m), and then narrow towards active faults, where uplift rates increase (Figure 4.5a). This narrowing fails to manifest just upstream or downstream of the faults, where the four widest channels in the river are located (~120 m) (Figure 4.5a). This is interpreted as a shift in their channel dynamics toward dominantly alluvial conditions due to fault back-tilting (e.g. Humphrey and Konrad, 2000). Notably, the dramatic variations in width (~10-150 m) persist even after normalisation by drainage area (i.e. wideness ~1.5-16 m/km) (Figure 4.5a). In comparison to width, across the same gradients in bedrock erodibility and uplift rate, channel slope exhibits less variation (~0.014-0.1), most of which occurs in the hard-bedrock-dominated upstream zone (Figure 4.5c). This pattern is also clear in

steepness, where the upstream channel k_{sn} fluctuates between ~ 300 – 500 m while the downstream channels stabilise at ~ 210 m (Figure 4.5c). The relative variability of wideness and steepness suggests that, except in the upstream zone, width adjustment is the main mechanism the sediment-rich Phoenix river uses to accommodate variations in bedrock erodibility and uplift rate. W/D is mostly < 10 in the upstream zone and increases to ~ 50 in the downstream zone, with the dominantly alluvial reaches reaching ratios up to 110 (Figure 4.5e).

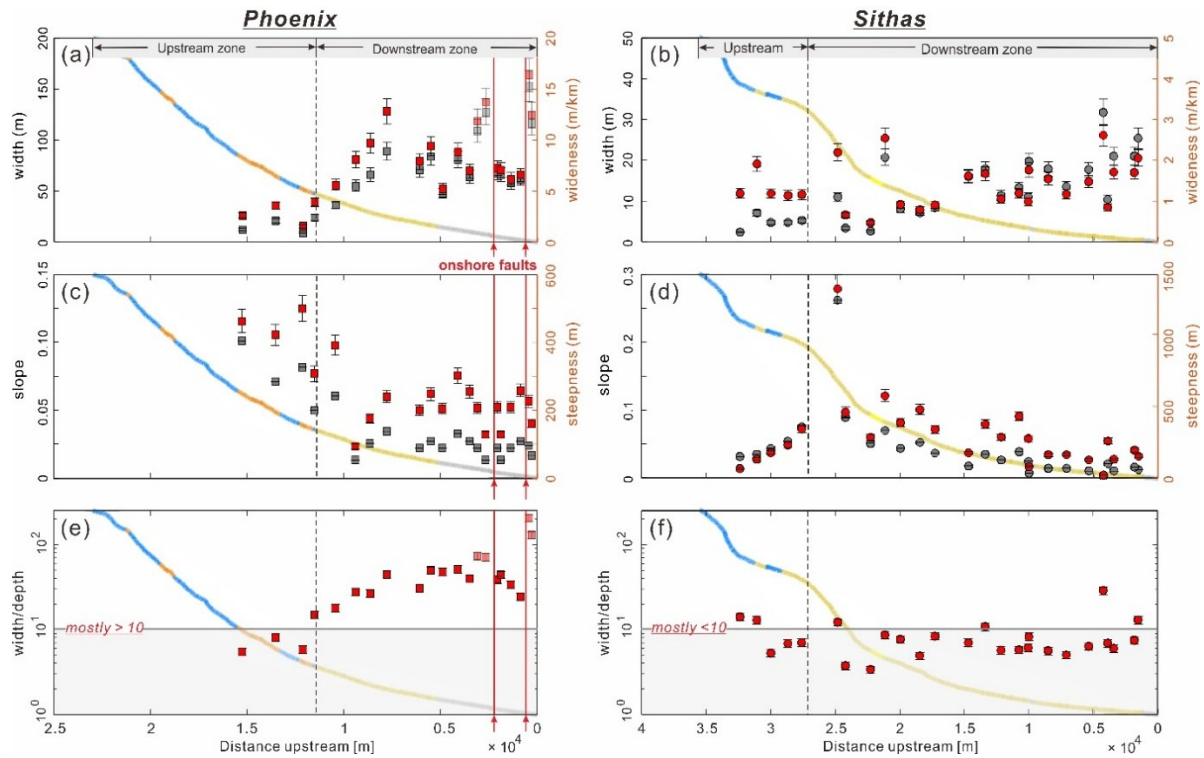


Figure 4.5. Bankfull width (grey) / wideness (red) (a, b), slope (grey) / steepness (red) (c, d), and width:depth, W/D (e, f) along the lithology-coloured trunk-channel river profiles for the Phoenix (a, c, e) and Sithas (b, d, f) rivers. Note the Phoenix river is dominated by width adjustments, whereas the Sithas river is dominated by slope adjustments. The onshore faults (Figure 4.3a) are marked as red lines, and the four transparent markers in the Phoenix river represent channels interpreted to be close to transport-limited alluvial conditions (see main text for details). See Figure 4.3 for lithology legend.

In contrast, for the sediment-poor Sithas river, channel width increases gradually from 2 m to 25 m downstream (Figure 4.5b). Yet, most of the width increase can be explained by drainage area, given only limited variation remains in wideness (0.5 – 2.5 m/km) (Figure 4.5b). Compared to width, channel slope increases from 0.03 to 0.25 near the knickpoint and then decreases downstream to 0.01 (Figure 4.5d). These variations persist in steepness, even when the impact of drainage area is accounted for (~ 20 – 1300 m, Figure 4.5d). The much larger relative variability in slope compared to width suggests

that slope adjustment is the main mechanism by which the sediment-poor Sithas river tries to achieve equilibrium. W/D is <10 across most channels, with no clear pattern observed (Figure 4.5f).

Taken together, the two rivers, with distinct sediment fluxes, differ in their morphological responses to comparable gradients in bedrock erodibility and uplift rate: the sediment-rich Phoenix mostly adjusts width, except in its upstream zone; in contrast, the sediment-poor Sithas predominantly adjusts slope (Figure 4.5a-d). In both rivers, the slope-dominated channel response appears to occur when W/D is approximately less than 10. This contrast in morphological response points to the crucial role of sediment supply in shaping the channel geometry.

4.5.2 Downstream variations in excess shear stress

Figure 4.6 plots excess shear stress along the trunk of the Phoenix and Sithas rivers, with tributary confluences marked as dashed blue lines. In the sediment-rich Phoenix, excess shear stress is ~ 3 times higher upstream than in the downstream zone (Figure 4.6a). This aligns with estimated variations in bedrock erodibility (Zhou et al., 2024), supporting the idea that the river has equilibrated with changing bedrock erodibility. In the downstream zone, excess shear stress increases almost linearly towards the faults (Figure 4.6a), consistent with the variations in uplift rates (Figure 4.3c). Notably, excess shear stress does not deviate significantly downstream of tributary confluences, where Q_s/Q_t is expected to change (Figure 4.6a). For example, tributary t5 has a bedload-generating-lithology proportion of 45% and a drainage area of 27 km², both of which are significant relative to those of the trunk channel (86%, 48 km², respectively in supplementary Table 4.S2). Its confluence with the trunk channel is expected to markedly disturb the trunk Q_s/Q_t , yet excess shear stress exhibits little corresponding response (Figure 4.6a). This implies that erosional efficacy remains approximately constant despite variations in Q_s/Q_t due to tributary input. This could occur either if the river approaches detachment-limited conditions or if the river lies in zone 3, where erosional efficacy is insensitive to Q_s/Q_t variations (Figure 4.1); however, the former can be excluded based on the clear observation that the channel is covered by coarse sediment (Figure 4.3e). Consequently, we interpret that the sediment-rich Phoenix river lies in zone 3.

In the sediment-poor Sithas river, excess shear stress shows a general downstream increase within reaches between tributary confluences, where Q_s/Q_t is approximately constant (Figure 4.6b). This appears to track the trend in uplift rates (Figure 4.3d), consistent with the notion that the channels are in equilibrium. Yet, when sediment-rich tributaries known to coarsen trunk-channel sediment grain size join the trunk channel, such as t2, t4, and t7, excess shear stress decreases markedly (Figure 4.6b) (c.f. Zhou et al., 2026). In the case that these channels are in equilibrium, the reduction in excess shear stress must have been compensated by a corresponding increase in erosional efficacy across

these tributary confluences (e.g. Cowie et al., 2008). Tributary t7, with a bedload-lithology proportion of 29% but a small drainage area of 6 km² relative to the trunk channel (52%, 125 km², respectively in Table 4.S2), is expected to have limited impact on trunk-channel's Q_s/Q_t ; however, excess shear stress drops to one-third of its initial value downstream of the confluence (Figure 4.6b). This observation suggests that excess shear stress and, therefore erosional efficacy, is highly sensitive to Q_s/Q_t variations. Thus, we interpret that the sediment-poor Sithas river lies in zone 1 (Figure 4.1).

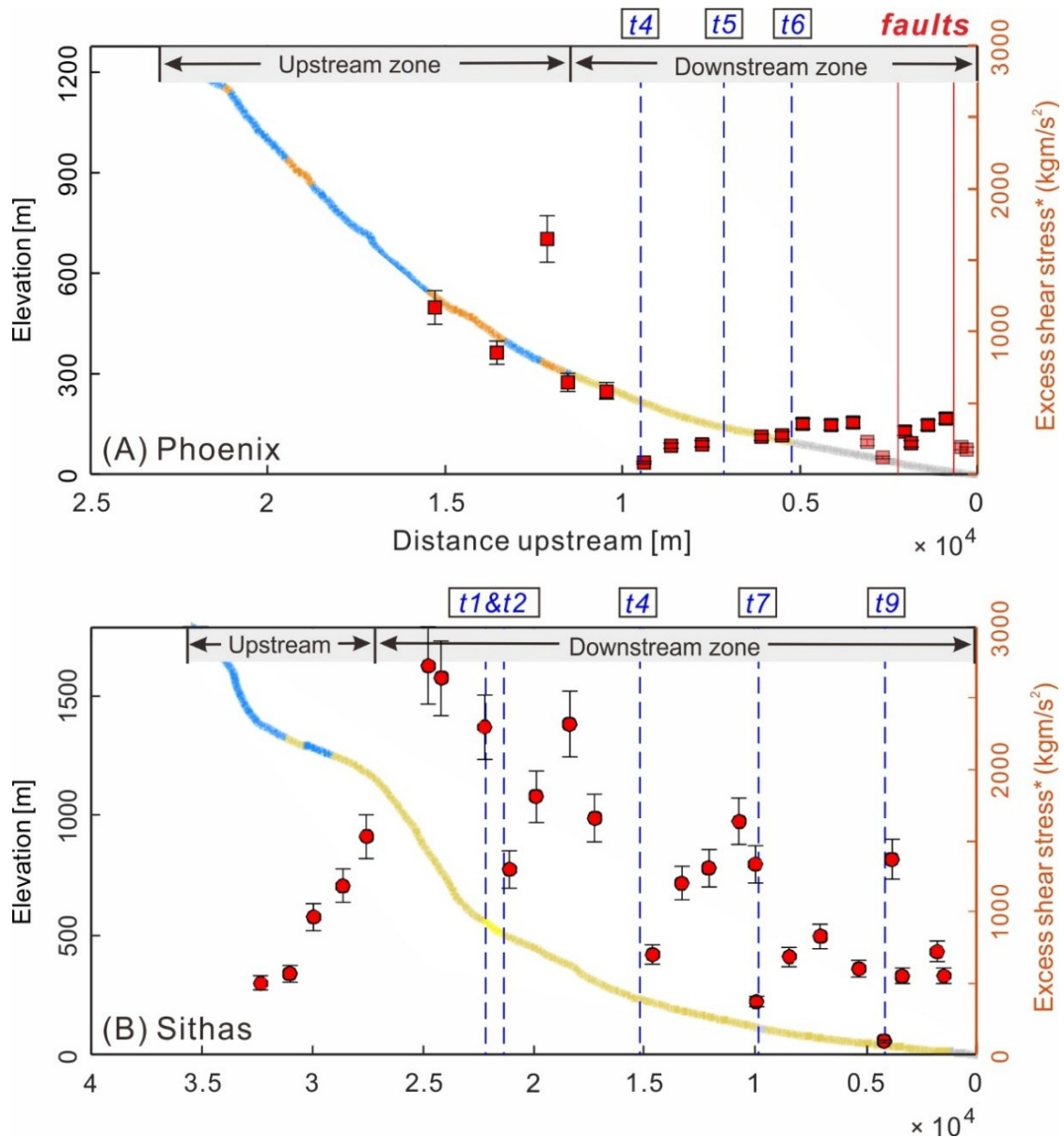


Figure 4.6. Downstream variations in excess shear stress along the lithology-coloured trunk-channel river profiles for the Phoenix (a) and Sithas (b) rivers, with the locations of tributary confluences marked (dashed blue lines). Note the sensitivity of excess shear stress to tributary input differs significantly between the Phoenix and Sithas rivers. The four transparent markers in the Phoenix river represent channels interpreted to be close to transport-limited alluvial conditions (see main text for details). See Figure 4.4 for lithology legend.

4.5.3 Constraints from Q_s/Q_t estimates

To further assess the interpretation of the Q_s/Q_t zones in which the two rivers fall, the Q_s/Q_t values are approximated. In the Phoenix river, bankfull Q_s can be estimated from dominantly transport-limited reaches at ~ 6000 kg/s, and bankfull Q_t , estimated from adjacent flux-dependent reaches, is ~ 7700 kg/s (see supplementary methods and Table 4.S1 for details). The Phoenix river's Q_s/Q_t is therefore constrained at ~ 0.8 , aligning with its interpreted position in zone 3 (Figure 4.1). Combined with the finding that the Q_s/Q_t of the Phoenix river is ~ 2.5 times that of the Sithas river (McLeod et al., 2024), our constraint on the Phoenix river can be extrapolated to the Sithas river ($Q_s/Q_t \sim 0.3$). This is consistent with its interpreted position in the tools domain (zone 1) (Figure 4.1).

4.6. Discussion

4.6.1 Constraints on the relative sediment flux function

We show two rivers with distinct sediment fluxes exhibit contrasting morphological responses to comparable gradients in bedrock erodibility and uplift rates (Figure 4.5). This is inconsistent with predictions from the SPL for detachment-limited rivers (e.g. Finnegan et al., 2006) but aligns with the effects of flux-dependent dynamics (e.g. Yanites and Tucker, 2010). In both rivers, excess shear stress variations across tributary confluences (Figure 4.6) are interpreted to reveal the sensitivity of erosional efficacy to Q_s/Q_t variations, with the sediment-rich Phoenix River being much less sensitive than the sediment-poor Sithas River. In particular, the sediment-rich Phoenix river is estimated to have a Q_s/Q_t of ~ 0.8 .

The high Q_s/Q_t zone (with markedly decreased sensitivity to Q_s/Q_t variations) is absent from Sklar and Dietrich's (2004) function but is integral to Turowski et al.'s (2007) function (zone 3 in Figure 4.1). Essentially, the existence of zone 3 is a direct result of the diminishing rate of decrease in the fraction of exposed bedrock as Q_s/Q_t increases. This feature, however, is challenging to discern directly in experimental studies (Johnson and Whipple, 2010). Hobbey et al. (2011) found their optimal sediment flux functions for three rivers flowing over coarse and loose postglacial debris are comparable in form to Turowski et al.'s (2007) function. Yet, this study is interpreted to provide ambiguous support for Turowski et al.'s (2007) function as the rivers they document never incise into bedrock and could therefore be transport-limited (Hobbey et al., 2011). Thus, to our knowledge, our evidence for a zone 3 river (i.e. the Phoenix river) represents the first unambiguous field support for Turowski et al.'s (2007) function over Sklar and Dietrich's (2004) for relative sediment flux (Figure 4.1).

The decreased sensitivity of erosional efficacy to Q_s/Q_t variations at high Q_s/Q_t values (Turowski et al., 2007) helps reconcile some long-standing conflicting observations. Below, we focus on the elusive

role of width versus slope in channel response to external forcing, the mixed success of the SPL, and the implications for interpreting flux-dependent river incision.

4.6.2 Channels increase shear stress: to steepen or to narrow?

In principle, rivers can narrow or steepen to increase shear stress (Equation 4.S2), but how this adjustment is partitioned between changes in width and slope remains elusive (e.g. Yanites and Tucker, 2010). It has been explored in numerical models incorporating flux-dependent incision, but such models often use overly simplified relative sediment flux functions or assume that channels reach their steady-state width at which slope is minimised (Turowski et al., 2009; Yanites and Tucker, 2010). Moreover, rivers with well-constrained sediment fluxes have been scarce for model validation. Here, we take the opportunity to explore what our results may imply about the contrasting dynamics of narrowing compared to steepening in sediment-influenced bedrock rivers.

In our study, the sediment-poor Sithas river has a W/D mostly <10 (Figure 4.5f). This W/D is close to a threshold value, below which shear stress maxima shift from the channel centre to the banks due to increased friction near the banks (e.g. Turowski et al., 2008). This mechanism, first examined in detachment-limited conditions, has been invoked to limit channel narrowing (e.g. Wobus et al., 2006). This is because further narrowing results in shear stress on the banks exceeding that on the bed, which increases bank incision and counteracts the narrowing. In contrast, channel slope—affecting only the magnitude, rather than the cross-sectional distribution, of shear stress—is still free to adjust. Consequently, rivers at a threshold width mostly steepen, rather than narrow, to increase their erosive power (e.g. Wobus et al., 2006). The same pattern is observed in the Sithas river (Figure 4.5b, d), in the zone 1 tools domain (Figure 4.6b), in which erosional efficacy increases as sediment concentration increases. This motivates us to propose how the threshold width model may be extended to flux-dependent rivers: when the channel bed and banks share limited sediment, sediments are expected to concentrate in areas of higher shear stress (e.g. Meyer-Peter and Müller, 1948), meaning the erosional efficacy of the bed and banks is shear-stress dependent. Incision in such flux-dependent channels thus depends only on shear stress, similar to the SPL (Equation 4.1), and channel cross-section equilibrium is consequently reached at the threshold width, where shear stress between the bed and banks is balanced: any further narrowing or widening is counteracted by subsequent redistribution of shear stress (Figure 4.7a).

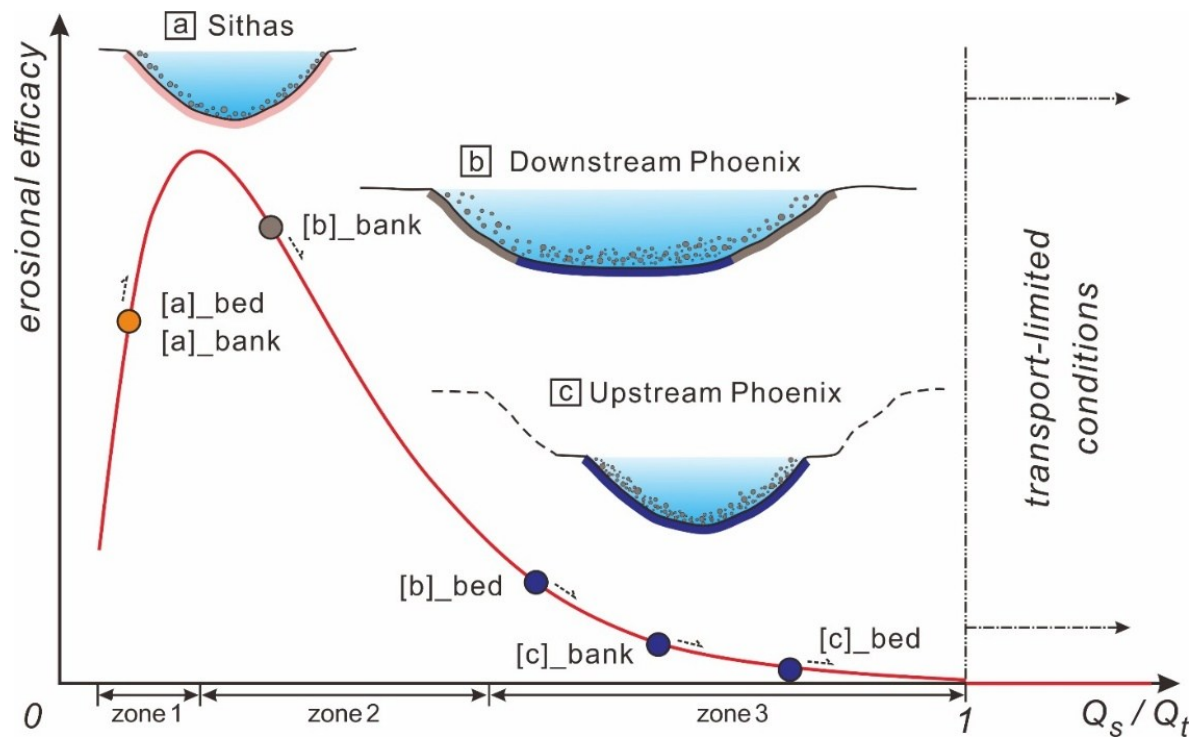


Figure 4.7. A schematic diagram illustrating the ‘tools and cover’ effect of sediment in shaping channel equilibrium geometry. Note, channels a and c exhibit similar channel geometry despite significant differences in relative sediment flux, because both have limited contrast in erosional efficacy between the channel bed and banks. Furthermore, this contrast does not change significantly when there is an increase in Q_s/Q_t , as indicated by dash arrows. In contrast, channel b, where the bank efficacy exceeds bed efficacy, exhibits a much wider channel. The increasingly shallow gradient of the curve in the cover domain causes a Q_s/Q_t increase to reduce bank efficacy more than bed efficacy, as indicated by dash arrows along the $f(Q_s/Q_t)$ curve.

Unlike the Sithas river, the sediment-rich Phoenix river in its downstream zone is significantly wider than a threshold width (Figure 4.5e), as observed in other studies (e.g. Baynes et al., 2020; Finnegan et al., 2017). In channels wider than the threshold width, shear stress is expected to be much higher on the bed than on the banks (e.g. Turowski et al., 2008). The Phoenix river as a whole is interpreted to lie in zone 3, so the channel bed with higher shear stress must also be in zone 3, as sediment concentration depends on shear stress (e.g. Meyer-Peter and Müller, 1948). The banks with lower shear stress should have lower sediment concentration and therefore higher erosional efficacy, likely zone 2 for the banks of the Phoenix river (Figure 4.7b). This may explain why sediment-rich rivers, such as the Phoenix river, are wider than sediment-poor rivers, such as the Sithas river: shear stress on the bed must be high enough to compensate for its lower efficacy (Figure 4.7b).

Importantly, channels in this condition are free to narrow or steepen to increase erosive power. In the downstream Phoenix river, however, the channels mostly adjust width, rather than slope, to accommodate variations in bedrock erodibility and uplift rate (Figure 4.5a,c). Similar observations have also been reported in both ancient (Amos and Burbank, 2007) and modern (Yanites et al., 2010;

Allen et al., 2013) sediment-rich rivers. This raises the question: why does a wide channel mostly adjust width to modulate its erosive power? Here, we hypothesise that, for channels such as the downstream-most reach of the Phoenix river, steepening alone can, in theory, elevate erosive power to balance increasing uplift rates towards the fault, but such steepened channels are not in full equilibrium: any perturbations that increase sediment concentrations can trigger an incision imbalance across the channel cross-section. This is because, for a given increase in Q_s/Q_t , bank efficacy in zone 2 decreases more strongly than bed efficacy in zone 3, due to the increasingly shallow gradient of the $f(Q_s/Q_t)$ in Turowski et al. 's (2007) formulation (Figure 4.7c). In this case, bank incision is surpassed by bed incision, promoting narrowing that increases sediment concentrations further (Figure 4.7c). This positive feedback loop continues until the channel narrows to a width where bank incision catches up with bed incision (Equation 4.1), and a new equilibrium is reached. In this process, the initial steepening is partly or fully erased as channels maintain equilibrium with the imposed uplift rates. Thus, wide channels would preferentially adjust their width to increase erosive power. If channels are narrowed to near their threshold width, however, further narrowing is inhibited and steepening becomes important again, as in the upstream Phoenix and other sediment-rich rivers (e.g. Amos and Burbank, 2007; Yanites et al., 2010; Allen et al., 2013). Here, we propose that in such channels, both the bed and banks have entered zone 3, where their difference in erosional efficacy approaches zero (Figure 4.7c). This again causes incision to depend mostly on shear stress, similar to the SPL (Equation 4.1). As a result, any narrowing can be counteracted by the subsequent increase in shear stress on the bed relative to the banks (Figure 4.7c).

Thus, a conceptual channel geometry model that is built upon the constrained flux-dependent incision and the equilibrium principle is proposed (Figure 4.7; c.f. Yanites and Tucker, 2010; Turowski et al., 2007). The model predicts the relative sediment flux conditions under which width or slope dominates channel morphological adjustments to changing boundary forcing (c.f. Yanites and Tucker, 2010; Yanites, 2018). In this model, sediment-rich rivers can be as narrow as bedrock rivers, a somewhat counterintuitive observation recently derived from ~1300 well-documented channels (Buckley et al., 2024). However, testing and validating this conceptual model requires more field, numerical, and experimental work that compares channel width and slope responses to known forcing.

4.6.3 Implications of flux-dependent incision for river interpretation

In general, our results illustrate that moving sediment exerts a dual control on channel geometry: first, by determining the excess shear stress required to maintain equilibrium with boundary conditions (Figure 4.6), and second, by influencing how that shear stress is partitioned between width and slope adjustments (Figure 4.7). This recognition consolidates the theoretical importance of sediment supply

and raises questions about the applicability of the SPL (Yanites, 2018). Yet, the SPL works surprisingly well in many rivers, even in sediment-rich ones (e.g. Lavé and Avouac, 2001; Yanites et al., 2010). Thus, we consider when and why the SPL can be reasonably used in river interpretation.

The SPL can be considered as an upscaled version of Equation 4.1, when erosional efficacy is constant and channel width scales only with drainage area (e.g. Whipple and Tucker, 1999). Its application therefore hinges on the extent to which these two conditions are met. For rivers with some sediment, such as the Phoenix river, their erosional efficacy is highly sensitive to variations in Q_s/Q_t (zone 1 in Figure 4.1). Here, maintaining equilibrium requires that excess shear stress and, implicitly, bankfull channel geometry must adjust significantly in response to variations in Q_s/Q_t . Furthermore, the excess shear stress in such rivers is adjusted mostly by slope, rather than width changes (Figure 4.7a), allowing the SPL to be effective at a reach scale where Q_s/Q_t is constant, but not across confluences where Q_s/Q_t might change (e.g. Zhou et al., 2024). In rivers carrying abundant sediment (e.g. the Phoenix river), however, channels may enter zone 3 of Turowski et al. (2007)'s function in which erosional efficacy becomes insensitive to changes in Q_s/Q_t (Figure 4.1). In this case, Q_s/Q_t variations are not the limiting condition as erosional efficacy is nearly constant; instead, the critical constraint is channel-width dynamics. If the banks are yet to reach zone 3, width changes are the preferred mechanism to adjust shear stress (e.g. the downstream Phoenix; Figure 4.7b). As a result, the SPL cannot be used to indicate incision rates across the whole river, but other shear-stress based incision laws that explicitly incorporates channel width can be used (e.g. Lavé and Avouac, 2001; Yanites et al., 2010). Once the banks also lie within zone 3, channel morphological adjustments are again mostly manifested as slope changes (e.g. the upstream Phoenix; Figure 4.7c). In such rivers, the SPL can capture the main characteristics of flux-dependent rivers.

Our results explain the mixed success of the SPL when tested against independent field evidence. For example, only in channels dominated by slope adjustments are channel steepness (i.e. drainage area-normalised slope) and knickpoints likely to be effective in mapping active deformation. In addition, the linear correlation between incision rate and excess shear stress in both the Phoenix and other zone 3 rivers (e.g. Lavé and Avouac, 2001; Yanites et al., 2010) suggest that the commonly observed nonlinearity in the incision rate – steepness relationship might arise from i) the inappropriate use of steepness in rivers dominated by width adjustments, and ii) variations in Q_s/Q_t . These insights are important because the SPL, due to its simplicity, has been, and will likely continue to be, a “default” model for river incision. Our approach helps to assess the confidence we should place in previous studies that used the SPL, and to choose appropriate rivers for its future use.

Taking flux-dependent dynamics fully into account is impeded by the lack of quantitative constraints on Q_s/Q_t ; however, its impact on river incision interpretation can be minimised. A critical step in this effort is to identify the Q_s/Q_t zone of the study rivers (cf. [Figure 4.1](#)). This can be achieved by analysing W/D and the sensitivity of excess shear stress to tributary input. As discussed in section 6.2, rivers with W/D approaching ~ 10 are likely to lie either within zone 1 or zone 3, with the latter further distinguished by insensitivity of excess shear stress to tributary input. The Q_s/Q_t range of the rivers can then be used to guide further analysis. For example, relationships between tectonics, lithology, and topography should be examined in zone 3 rivers. In addition, excess shear stress serves as a recommended proxy for external forcing, especially in zone 2 rivers.

4.7 Conclusions

This study shows that two rivers, with distinct sediment fluxes, exhibit contrasting morphological responses to comparable tectonic and lithological gradients. In the sediment-poor Sithas river, excess shear stress is sensitive to tributary input, consistent with the tools effect of sediment supply (zone 1 in [Figure 4.1](#)). In contrast, in the sediment-rich Phoenix river, with a Q_s/Q_t of ~ 0.8 , excess shear stress exhibits little response to tributary input. This supports the existence of a high Q_s/Q_t zone, in which erosional efficacy approaches a constant (zone 3 in [Figure 4.1](#)). This behaviour is absent from [Sklar and Dietrich's \(2004\)](#) tools versus cover effect model of sediment flux dependent incision but is an integral part of the [Turowski et al. 's \(2007\)](#) model, in which the rate of decrease in the exposed bedrock fraction diminishes as Q_s/Q_t increases. This, to our knowledge, represents the first unambiguous field evidence favouring [Turowski et al. 's \(2007\)](#) function and suggests that the erosional efficacy of rivers is best described by an asymmetric sediment flux function that incorporates tools and cover effects.

In reaching the excess shear stress required to maintain equilibrium, the sediment-poor Sithas river mostly adjusts its slope, whereas the sediment-rich Phoenix river mostly adjusts its width, except in its upstream sections. Importantly, the changing importance of slope versus width can be explained by flux-dependent behaviour in channels that follow Turowski's relative sediment flux equation. Consequently, we conclude that sediment controls channel geometry not only by determining the excess shear stress required to balance boundary conditions, but also by influencing how that excess shear stress is partitioned between width and slope adjustments.

The stream power law is a special case of sediment flux-dependent incision, and we argue it is most effective where Q_s/Q_t is relatively constant down-system or where the river lies in the high Q_s/Q_t zone characterised here (zone 3 in [Figure 4.1](#)), in which additional sediment flux does not significantly change erosional efficacy. In addition, the finding that channel response to external forcing is

sediment flux-dependent challenges common tectonic interpretations of long profiles because some rivers mostly adjust width, rather than slope, in response to forcing such as active faulting. To minimise the impact of flux-dependent dynamics, we advocate examining the Q_s/Q_t zone of rivers as a routine part of fluvial geomorphology.

4.8 Acknowledgements

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4.9 Supplementary material

4.9.1 Supplementary methods

4.9.1.1 Errors associated with channel geometry measurements

For our channel geometry measurements, instrumental errors for the laser rangefinder are minimal; most errors arise from selecting the bankfull stage to measure ([Whittaker et al., 2007b](#)). In our study, the estimation of bankfull stage mostly impacts the measurement of depth, rather than slope and width, due to their mostly rectangular cross-sectional shape. Here, we put a $\pm 15\%$ error bar to individual slope and width measurements based on previous field studies ([e.g. Whittaker et al., 2007a](#)). No error bar was assigned to individual depth measurements; instead, depth was used statistically in calculating excess shear stress based on the expectation that water discharge correlates linearly with drainage area (see below for more details; [Table 4.S1](#)).

4.9.1.2 Excess shear stress calculation

Excess shear stress quantifies the force exerted by flowing water in excess of the threshold required to mobilise sediment ([e.g. Yanites et al., 2010](#)). The first step was to estimate bankfull water discharge using our channel geometry data, as the two study rivers are not gauged. Here, bankfull water discharge (Q_w) is estimated from flow velocity using Manning's equation and channel cross-section area:

$$Q_w = vC = \frac{1}{n} H^{2/3} S^{1/2} C, \quad (4.S1)$$

where v is flow velocity in m/s, C is the cross-sectional area in m^2 , and n is the Manning friction factor, ~ 0.03 ([e.g. Kent et al., 2021](#)). Note, the value of n mostly impacts the specific magnitude of estimated bankfull discharge. In subsequent calculation of shear stress, its influence cancels out ([Equation 4.S2](#)). The point-based estimates of bankfull water discharge can be rather variable over short length scales ([e.g. Kent et al., 2021](#)), partly due to the uncertainties in the measured depth as mentioned above. In

small catchments with minor precipitation variations (e.g. Allen et al., 2013), as in our study (Zhou et al., 2024), a linear relationship between water discharge and drainage area is expected. Here, we exploited this relationship to obtain an average scaling ratio of $Q_w: A$ for both rivers (Table 4.S1) to smooth local variations in estimated water discharge. The calibrated water discharge was then used to calculate bankfull shear stress (τ_b):

$$\tau_b = \rho_w g \left(\frac{n Q_w}{W} \right)^{0.6} S^{0.7}, \quad (4.S2)$$

where ρ_w is the water density and g is gravitational acceleration (e.g. Allen et al., 2013). The critical shear stress (τ_c) required to entrain sediment grains was further estimated as (e.g. Baynes et al., 2020):

$$\tau_c = \tau_c^* (\rho_s - \rho_w) g D_{50}, \quad (4.S3)$$

where τ_c^* is the critical Shield stress, ρ_s is the density of sediment, $\sim 2650 \text{ kg/m}^3$, and D_{50} is the median grain size, as reported in Zhou et al. (2026). Here, we set the critical Shield stress at 0.06, given that the studied rivers are steep (e.g. Lamb et al., 2009; Carr et al., 2023). The choice of τ_c^* is unlikely to significantly affect either the spatial patterns or the absolute magnitude of excess shear stress, because the current critical shear stress constitutes only $\sim 10\%$ of the bankfull shear stress.

4.9.1.3 Estimation of sediment supply and transport capacity

To estimate relative sediment flux (Q_s/Q_t), we need to constrain both Q_s and Q_t . The former cannot be directly measured, but the latter can be estimated for any given channel using the Meyer-Peter and Müller equation (Equation 4.S4):

$$Q_c = \rho_s \left(\frac{\rho_s - \rho_w}{\rho_w} g D_{50}^3 \right)^{0.5} C (\tau^* - \tau_c^*)^{1.5} W, \quad (4.S4)$$

where $\tau^* = \tau_b / [(\rho_s - \rho_w) g D_{50}]$ is the Shield stress, τ_c^* is the critical Shield stress (taken as 0.06), and the constant $C = 3.97$ based on Wong and Parker (2006). Here, we focused on the reach downstream of the t6 of the Phoenix river (Figure 4.6a), for which Q_s is expected to be approximately constant. In this reach, channels near the faults appear to have shifted to dominantly alluvial conditions (i.e. $Q_s \approx Q_t$), allowing us to use the average Q_t from individual channels (Equation 4.S4) to represent the reach-scale Q_s (Table 4.S1). For other channels not in alluvial conditions within the reach, we averaged their individual Q_t estimates to obtain the reach-scale Q_t (Table 4.S1).

4.9.2 Table 4.S1. Bankfull channel geometry, discharge, excess shear stress, and transport capacity.

River	Locality	GPS (decimal degrees)		Bankfull channel geometry			Excess shear stress (kgm/s ²)	Bankfull transport capacity* (kg/s)	Upstream area (km ²)	Bankfull discharge ¹ (m ³ /s)	Scaled depth*(m)	Bankfull shear stress (kgm/s ²)	D50 (mm)	Critical shear stress (kgm/s ²)	Shield stress
		Latitude	Longitude	width (m)	depth (m)	slope (y/x)									
Phoenix	Ph1	38.293444	21.996250	109.0	1.50	0.023	224	7343	84.8	1076	1.14	258	35.0	35	0.4
Phoenix	Ph3	38.246917	21.935000	9.3	1.60	0.082	1647	12537	37.2	195	2.07	1699	52.0	51	2.0
Phoenix	Ph6	38.249472	21.938917	24.0	1.60	0.051	639	7819	37.7	394	1.36	691	52.0	51	0.8
Phoenix	Ph8	38.254528	21.947306	36.0	2.00	0.061	582	10198	41.5	942	1.07	655	73.0	72	0.5
Phoenix	Ph9	38.261889	21.952361	55.2	2.00	0.014	81	813	46.4	690	1.38	192	112.5	111	0.1
Phoenix	Ph10	38.267444	21.955250	66.3	2.50	0.026	200	3785	46.6	1647	1.03	268	69.0	68	0.2
Phoenix	Ph11	38.273611	21.959167	88.8	2.00	0.035	213	5550	48.0	1756	0.80	280	68.5	68	0.2
Phoenix	Ph12	38.282917	21.970139	70.6	2.30	0.023	258	5914	79.9	1421	1.42	323	65.5	65	0.3
Phoenix	Ph13	38.283528	21.975667	84.1	1.70	0.028	271	7565	80.4	1135	1.21	338	67.6	67	0.3
Phoenix	Ph14	38.285861	21.981000	47.7	1.00	0.023	355	6430	83.4	240	1.85	420	65.3	65	0.4
Phoenix	Ph15	38.290167	21.986528	81.1	1.60	0.033	342	10331	84.5	1078	1.21	401	60.0	59	0.4
Phoenix	Ph16	38.292333	21.992500	63.9	1.60	0.028	362	8890	84.8	779	1.47	411	49.2	49	0.5
Phoenix	Ph17	38.295611	22.000056	127.7	1.80	0.014	117	3276	87.2	1340	1.22	170	53.0	52	0.2
Phoenix	Ph19	38.298056	22.004722	67.6	1.75	0.023	303	7183	87.3	863	1.54	350	47.5	47	0.4
Phoenix	Ph20	38.299139	22.006222	65.8	1.50	0.014	216	4200	87.3	509	1.81	253	37.9	38	0.4
Phoenix	Ph21	38.302028	22.009333	57.8	1.70	0.023	345	7468	87.4	703	1.69	385	40.0	40	0.6
Phoenix	Ph23	38.304556	22.013194	61.4	2.50	0.028	389	9490	87.5	1577	1.54	430	41.5	41	0.6
Phoenix	Ph25	38.307556	22.015444	153.0	0.75	0.024	194	8344	87.5	493	0.93	226	32.0	32	0.4
Phoenix	Ph39	38.229667	21.924056	12.0	2.20	0.102	1168	9655	21.0	474	1.18	1199	31.5	31	2.3
Phoenix	Ph43	38.238972	21.927056	21.0	2.60	0.072	851	10506	35.1	921	1.27	913	63.0	62	0.9
Phoenix	69(0)	/	/	116.6	0.90	0.017	171	5262	87.5	431	1.20	210	39.5	39	0.3
Sithas	54(2)	38.005664	22.514868	8.4	1.00	0.037	1658	11370	84.6	53	4.60	1685	27.0	27	3.8
Sithas	54(3)	38.009924	22.539025	16.1	2.30	0.017	699	5985	98.5	283	4.25	742	43.0	43	1.0

Sithas	54(4)	38.019493	22.557553	11.4	2.00	0.026	1307	10856	117.7	195	5.14	1346	40.0	40	2.0
Sithas	54(5)	38.031108	22.570487	19.7	2.40	0.007	371	2831	124.8	236	5.70	398	28.0	28	0.9
Sithas	54(6)	38.046208	22.587884	13.6	2.70	0.014	827	6502	135.0	279	6.08	849	21.5	21	2.4
Sithas	54(7)	38.054769	22.610432	31.8	1.10	0.002	94	580	147.7	52	7.18	125	32.0	32	0.2
Sithas	54(8)	37.967291	22.473623	3.4	0.90	0.089	2641	9224	26.1	28	3.00	2675	34.0	34	4.8
Sithas	54(9)	37.965185	22.468712	11.0	0.90	0.262	2727	31580	25.2	158	1.04	2735	7.9	8	20.9
Sithas	54(10)	37.937755	22.438043	4.8	0.90	0.044	964	2878	16.2	28	2.26	988	24.5	24	2.4
Sithas	54(11)	37.947590	22.441771	4.8	0.70	0.054	1181	3928	17.8	21	2.24	1211	30.0	30	2.4
Sithas	54(12)	37.954748	22.445804	5.3	0.75	0.075	1529	6331	20.5	30	2.09	1573	44.0	44	2.2
Sithas	54(13)	37.919411	22.429917	2.4	0.17	0.031	498	537	4.1	1	1.65	518	20.0	20	1.6
Sithas	54(14)	37.929891	22.434324	7.1	0.55	0.035	567	1935	13.8	16	1.73	605	38.0	38	1.0
Sithas	54(15)	37.976970	22.486476	2.6	0.77	0.051	2296	5813	31.3	13	4.60	2330	34.0	34	4.2
Sithas	54(16)	37.982850	22.495053	20.8	2.40	0.070	1298	19576	66.6	787	1.90	1329	30.5	30	2.6
Sithas	54(17)	37.991994	22.496814	8.1	1.05	0.044	1812	12548	77.7	61	4.23	1846	35.0	35	3.2
Sithas	54(18)	37.999185	22.505918	7.1	1.45	0.052	2316	16025	82.4	101	4.47	2341	24.7	24	5.7
Sithas	54(20)	38.026136	22.565890	13.3	2.30	0.038	1635	17727	124.2	348	4.32	1658	23.0	23	4.4
Sithas	54(21)	38.014424	22.548954	17.9	1.65	0.035	1202	15039	114.4	257	3.54	1235	33.0	33	2.3
Sithas	54(22)	38.037855	22.578829	17.9	3.17	0.014	686	6480	134.0	482	5.12	715	29.5	29	1.5
Sithas	54(23)	38.053198	22.599368	17.8	2.80	0.010	600	5260	144.7	337	5.87	615	15.3	15	2.4
Sithas	54(24)	38.057916	22.616059	21.0	3.50	0.010	548	5435	151.0	578	5.45	570	22.4	22	1.5
Sithas	54(25)	38.070199	22.621245	25.4	1.95	0.012	551	6617	152.8	285	4.67	571	20.5	20	1.7
Sithas	si1	38.054361	22.613806	10.4	1.50	0.021	1368	10615	150.2	99	6.72	1408	40.0	40	2.1
Sithas	si3	38.030750	22.570333	11.0	1.80	0.024	1334	10803	124.8	153	5.55	1357	24.0	24	3.4
Sithas	si5	38.068028	22.622778	21.0	2.80	0.016	720	8175	152.5	488	4.85	762	43.0	43	1.1

1, Rectangular channels, with depth used as a surrogate for hydraulic radius.

*, Discharge values scaled to a regional A (km²) : Q (m³/s) of 1 : 8, derived from applying Mannings equation to the two studied rivers, and using the assumption that discharge increases downstream proportionally to drainage area. These values in bold were used to constrain the Qs/Qt of the Phoenix river, with Qs taken from the averaged red values and Qt from the averaged black.

4.9.3 Table 4.S2. Tributary bedload-lithology proportion and drainage area relative to the trunk channel.

River	Tributary confluence	Tributary		Trunk upstream	
		bedload-generating-lithology proportion	drainage area (km2)	bedload-generating-lithology proportion	drainage area (km2)
Phoenix	t3	78%	2.4	97%	39.0
	t4	34%	4.6	96%	41.8
	t5	45%	27.0	86%	48.0
	t6	0%	1.9	66%	81.0
Sithas	t9	0%	2.0	46%	147.9
	t8	9%	5.0	50%	135.0
	t7	29%	5.9	52%	125.5
	t6	26%	4.3	54%	118.9
	t5	70%	14.1	55%	99.7
	t4	33%	5.6	57%	92.6
	t3	0%	8.3	83%	69.3
	t1 + t2	65%	32.6	79%	31.3

Note the tributaries in red are the two selected to evaluate the sensitivity of excess shear stress to expected trunk-channel's Q_s/Q_t variations, and those in bold are known to coarsen the trunk-channel's grain size according to Zhou et al. (2026).

5 DISCHARGE VARIABILITY CONTROLS CHANNEL WIDTH BUT NOT SLOPE ¹

Abstract

Much of Earth's surface is shaped by rivers, whose width and slope adjust in concert to equilibrate with external forcings such as climate. This equilibrium channel geometry, as a benchmark towards which rivers adjust, is central to predicting geohazards, ecosystem services, and landscape dynamics. Yet, the impact of discharge variability on channel geometry remains poorly constrained by field observation. Here, we focus on steady-state, metamorphic-rock-dominated rivers with conditions typical of most rivers worldwide, and examine their channel slope and width responses to aridity index and mean annual precipitation, both of which can be used as proxies for discharge variability. Notably, our results show that channel width, but not slope, correlates with these discharge variability proxies. Moreover, channel width is shown to increase with discharge variability, following a power-law relationship. This width dynamic, not fully captured in many current models, may have suppressed the expected slope response to discharge variability. Combined with climate models, we suggest that enhanced storminess in a warming world may drive many rivers across the globe to widen. As a result, cumulative bank erosion is likely to promote bank failures, intensify channel mobility, and exacerbate flood risk in the near future.

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Zhou, Z., Whittaker, A.C., Hampson, G.J., Bell, R.E., 2026. Discharge variability controls channel width but not slope. **To be submitted to *Geology*.**

5.1 Introduction

Much of Earth's surface is shaped by rivers that erode bedrock and distribute water, sediment, and nutrients (e.g. Allen, 2008). In doing so, their channel slope and width approach an equilibrium with external forcings, such as climate (e.g. Stark et al., 2006; Lague et al., 2005). This equilibrium channel geometry, as a benchmark towards which rivers adjust, is central to predicting erosion and flood risks, ecosystem services, and landscape dynamics (e.g. Phillips et al., 2022; Yanites, 2018). However, despite efforts to quantify controls on channel geometry, considerable variability remains (e.g. Leopold and Maddock, 1953; Whipple et al., 2013; Phillips et al., 2024). Part of this remaining variability has been hypothesised to arise from discharge variability (e.g. Stark et al., 2016; Whipple et al., 2013)—a parameter that is expected to increase for many rivers in the near future due to enhanced storminess in a warming world (e.g. IPCC; Trambly and Somot, 2018; Trenberth, 2011).

Discharge variability is expected to influence channel geometry, because river planform reflects cumulative erosion of both channel bed and banks during threshold-exceeding floods (Figure 5.1A and B). The frequency-magnitude relation of floods, exemplified in Figure 5.1C, can be described by a stretched exponential distribution, with the shape parameter c characterising discharge variability (Rossi et al., 2016). Lower values of c indicate higher discharge variability, allocating more of the total discharge to extreme events (Figure 5.1C). In turn, floods exceeding a particular erosion threshold, such as the bank threshold Q_{bank} (shaded area in Figure 5.1C), become more frequent, enhancing cumulative erosion that drives channels towards a new equilibrium geometry (Figure 5.S1). Notably, for a given increase in discharge variability, the resulting enhancement of erosion depends on the magnitude of erosion thresholds (compare Q_{bank} and the bed-erosion threshold, Q_{bed} , in Figure 5.S1). As a result, channel width in actively incising rivers is hypothesised to be more sensitive to discharge variability than slope, because Q_{bank} is expected to be higher than Q_{bed} (Figures 5.1 A, B and 5.S1).

The expectation that Q_{bank} is higher than Q_{bed} arises from the requirement for significant bank inundation to derive bank retreat, the stabilising effects of vegetation (e.g. Ielpi and Lap tre, 2019), and cross-sectional variations in shear stress (Figure 5.1D; e.g. Turowski et al., 2008). This is also supported by empirical evidence that moderate floods can incise channel beds, but only extreme floods cause erosion to channel banks (Hartshorn et al., 2002; Barbour et al., 2009). Yet, although channel width may be more sensitive to discharge variability (Figure 5.S1), current studies to date have mostly focussed on channel slope adjustments to discharge variability (e.g. DiBiase and Whipple, 2011; Marder and Gallen, 2023). This omission is problematic given that

currently modelled slope adjustments (e.g. Lague et al., 2005) will be modulated by width responses to discharge variability. In addition, channel width responses can be integral to understanding the impact of ongoing climate change on erosion and flood risks (e.g. Barneveld et al., 2025), as well as sediment and nutrient export (cf. Molnar et al., 2006; Barefoot et al., 2022).

To address this research gap, we test the hypothesis that channel width is more sensitive to discharge variability than slope. For this purpose, we first select rivers with similar lithology and with erosion rate and precipitation typical of most global rivers (Figure 5.2), based on a recent global river dataset by Marder and Gallen (2023). We then compare channel slope and width responses to mean annual precipitation (MAP) and aridity index ($AI = MAP/\text{mean annual potential evapotranspiration}$), both of which are used as effective proxies for discharge variability (e.g. Rossi et al., 2016; Phillip et al., 2024). This proxy approach is necessary and effective in larger-scale analyses such as ours, because most rivers are not gauged (e.g. Marder and Gallen, 2023) and these metrics are already documented to scale inversely with discharge variability as a power law (e.g. Rossi et al., 2016). Based on these results, we provide unique insights into the impact of discharge variability on landscape dynamics.

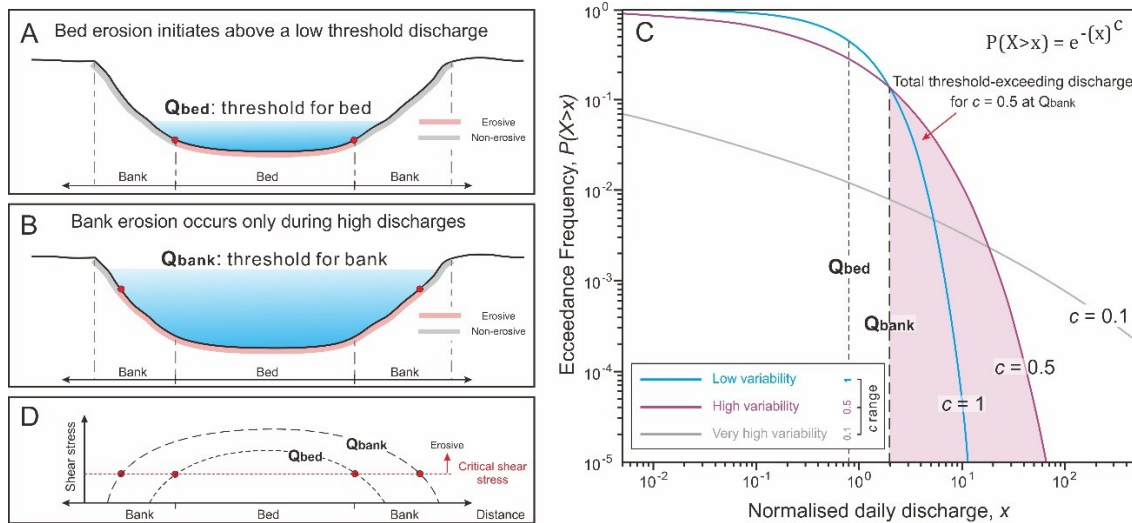


Figure 5.1. Erosion thresholds and their impacts on threshold-exceeding floods. (A, B) Threshold discharges for the channel bed, Q_{bed} (A) and for banks, Q_{bank} (B). (C) Flood frequency-magnitude relation described by a stretched exponential distribution, whose shape parameter c characterises discharge variability. (D) Cross-sectional variations in shear stress cause Q_{bank} to be higher than Q_{bed} for a given critical discharge. Here, an increase in discharge variability (e.g. from $c = 1$ to $c = 0.5$) redistributes more of the total discharge into bigger events, allowing more frequent threshold-exceeding floods (e.g. the area below $c = 0.5$ is greater than that below $c = 1$ for Q_{bank}). Note, the area increase (termed erosion enhancement) caused by a given increase in discharge variability depends on the magnitude of threshold discharges, allowing channel width in actively incising rivers to be more sensitive to discharge variability than slope (Figure 5.S1).

5.2 Methods

To test whether and to what extent channel width and slope respond to discharge variability, we carefully selected rivers across the globe based on three criteria: First, we restricted our analysis to well-constrained, fluvial-dominated rivers that are in steady-state and incise similar lithology. To this end, we exploited a global river dataset by [Marder and Gallen \(2023\)](#) that compiled such rivers, along with independent constraints on cosmogenic ^{10}Be -derived erosion rates E , lithology, and climate data — mean annual precipitation (MAP) and aridity index (AI) ([Supplementary material](#)). In this dataset, rivers incising >80% metamorphic rock were selected to control for lithology. Second, we focused on rivers with erosion rates and precipitation typical of rivers worldwide to ensure broad applicability. Thus, rivers with $E < 300$ m/Ma, comparable to the global median of ~ 60 m/Ma and $MAP \sim [300, 1800]$ mm/yr, a range that covers 76% of catchments worldwide ([Harel et al., 2016](#)) were chosen. Third, we limited our selection to those with identifiable channels in Google Earth imagery, to enable subsequent width measurements. Together, these objective criteria yielded a total of 72 catchments selected worldwide based on their physical characteristics, rather than their geographic position ([Figure 5.2](#); [Supplementary Table 5.S1](#)).

For each selected river, we measured channel slope and width, representing time-integrated channel bed and bank erosion. For channel width, we focused on the maximum distance between channel margins—identified from breaks in vegetation, colour contrasts, and other visible changes in surface texture—in high-resolution Google Earth imagery ([Figure 5.S1](#); e.g. [Scherler et al., 2017](#)). For each river, we sampled locations at ca. 5% increments of cumulative upstream drainage area along the trunk rivers, and measured width at up to five equally spaced cross sections (~ 50 m) at each sampling point to calculate the local mean and standard deviation ([Supplementary Material](#)). In total, ~ 2350 width measurements were made at 522 locations. We then plotted channel width against drainage area, coloured by AI to assess the robustness of width measurements and illuminate possible controls ([Figure 5.3](#); for similar with E and MAP see [Figure 5.S3](#)).

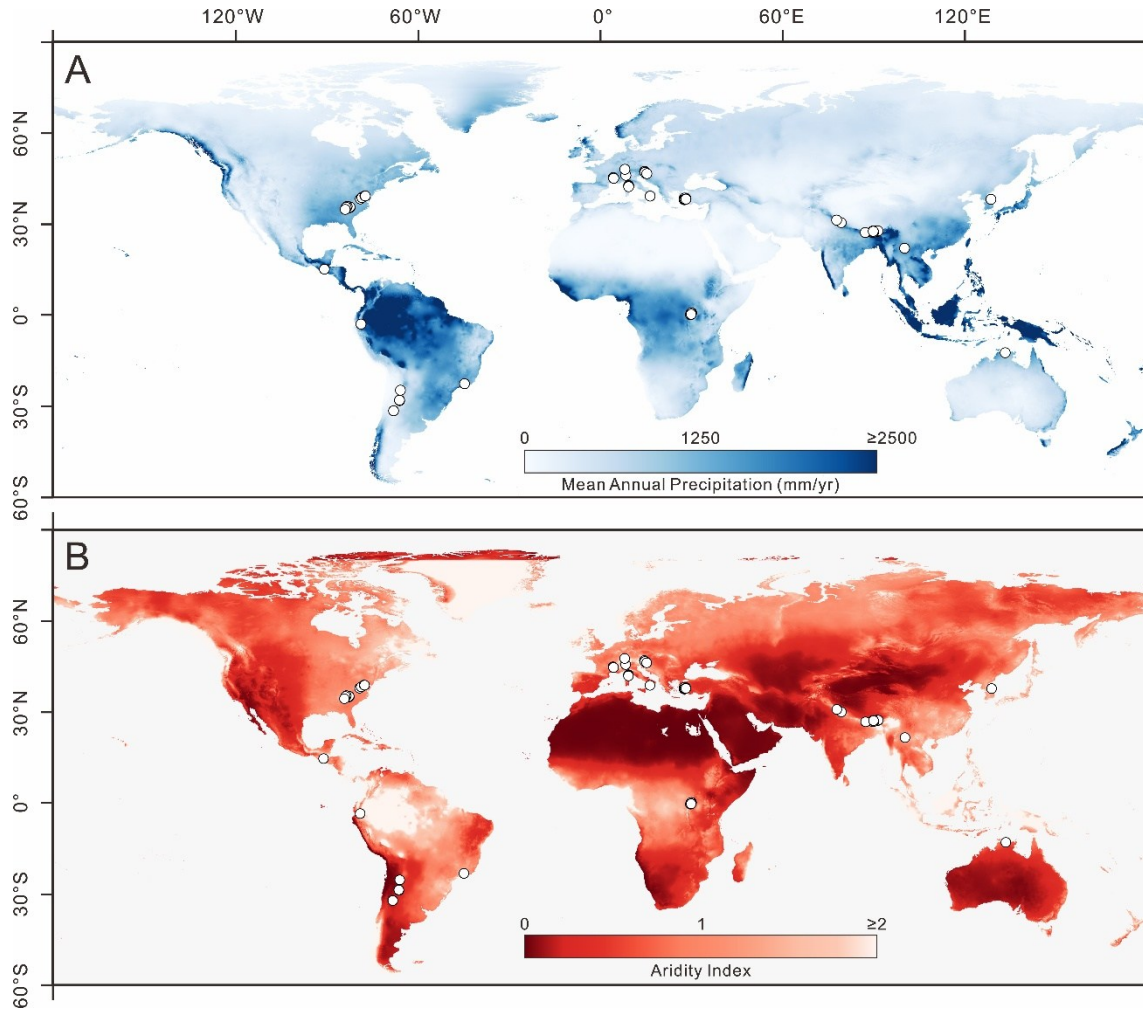


Figure 5.2. Locations of selected rivers on (A) global mean annual precipitation map (Fick and Hijmans, 2017), (B) global aridity index map (Trabucco and Zomer, 2009). Aridity index is the ratio of mean annual precipitation to potential evapotranspiration.

To isolate the link between channel geometry and discharge variability, we accounted for the effects of drainage area and erosion rate. For the well-established effect of drainage area on channel geometry, we calculated area-normalised channel slope (steepness, k_{sn}) and area-normalised width (wideness, k_{wn}) (e.g. Lague, 2014; Yanites, 2018).

$$k_{sn} = SA^a, \quad (5.1)$$

$$k_{wn} = WA^{-b}, \quad (5.2)$$

where A is upstream drainage area and a (also known as concavity index) and b are constants that fall within the range of 0.3-0.5 for most rivers worldwide (e.g. Wohl and David, 2008; Marder and Gallen, 2022). Here, we set a and b to 0.5, but conducted sensitivity analyses with both set to 0.3 and 0.4 (supplementary material; Table 5.S2). We calculated k_{sn} using an integrative approach (Supplementary material). We converted the mean width measured at

each location into k_{wn} , and then derived their mean and standard deviation to define the catchment-scale k_{wn} . For the effect of erosion rate on channel geometry, we examined both E - k_{sn} and E - k_{wn} relationships using Pearson correlation and, where appropriate, derived best-fit functions. For the significant E - k_{sn} relationship, we normalised k_{sn} by the predicted values from the function (termed k_{sn-E}), to adjust for the effect of erosion rate (e.g. Whipple, 2004; Lague, 2014). To contrast the responses of channel slope and width to discharge variability, we plotted k_{sn-E} and k_{wn} against MAP and AI , both of which are used as effective proxies for discharge variability (Figure 5.4), as in other large-scale analyses (e.g. Phillip et al., 2024; Marder and Gallen, 2023). This proxy approach is supported by both published power-law relationships linking MAP and AI to the discharge-variability parameter c (e.g. Rossi et al., 2016) and by numerical studies (Deal et al., 2018).

5.3 Results

Figure 5.3 shows that our measured channel width scales with drainage area, with a power-law exponent of 0.4. This exponent falls within the common range (0.3-0.5) reported by previous compilations (e.g. Whipple, 2004; Wohl and David, 2008), supporting the robustness of our width measurements. Another observation consistent with these studies is that the unexplained variability in width–area scaling can span up to an order of magnitude; for example, channel width ranges from $\sim 3\text{m}$ to $\sim 30\text{m}$ for a drainage area of $\sim 10\text{ km}^2$ (Figure 5.3). Significantly, we note that rivers with higher AI (lower discharge variability) in general appear to have narrower channels than those with lower AI (Figure 5.3), consistent with a hypothesised role of discharge variability in influencing channel width.

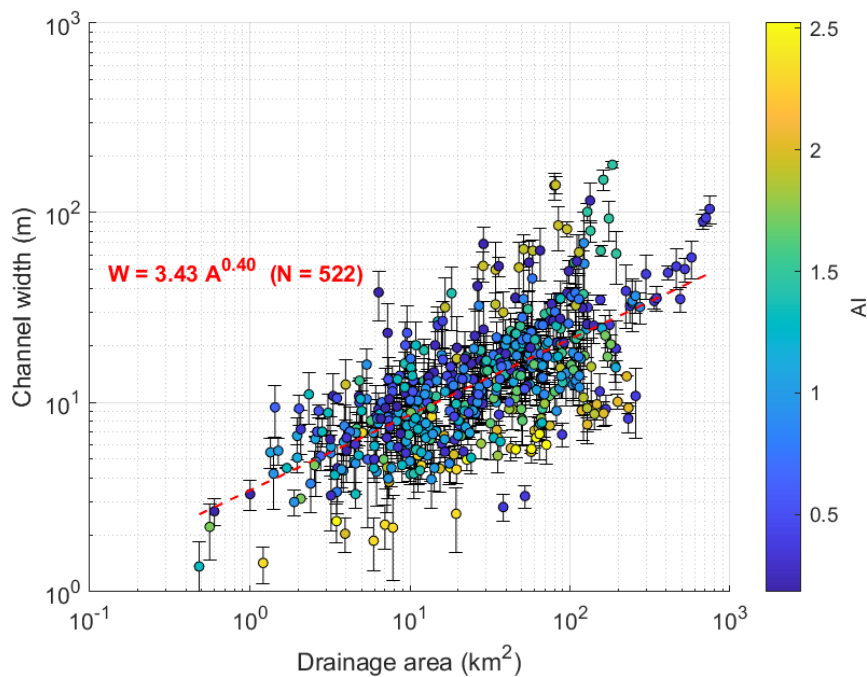


Figure 5.3. Channel width versus drainage area, coloured by aridity index (AI). Each data point represents the mean width at a given location, with standard deviation shown as error bars. In total, ~2350 width measurements were made at 522 locations across 72 rivers.

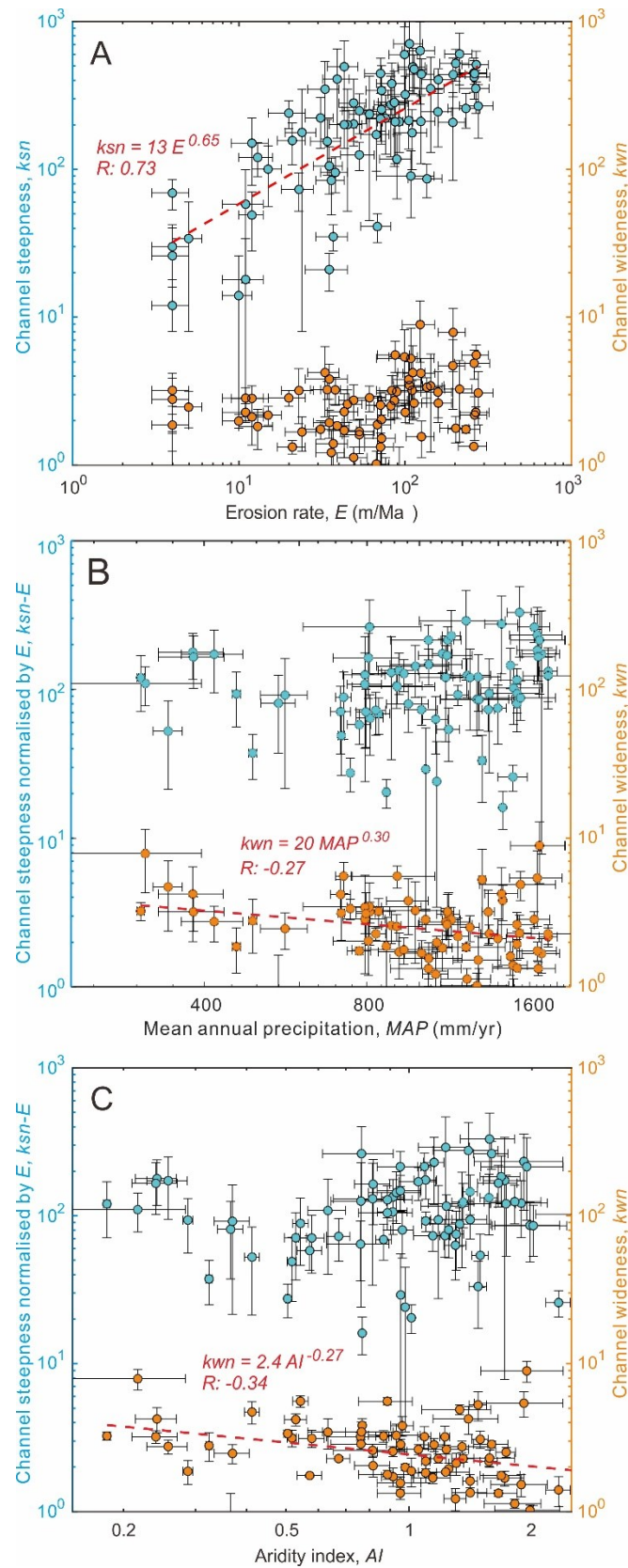


Figure 5.4. (A) k_{sn} and k_{wn} versus E , (B) k_{sn-E} and k_{wn} versus MAP , and (C) k_{sn-E} and k_{wn} versus AI . Here, k_{sn-E} is k_{sn} normalised by the predicted values from the best-fit function in (A).

Figure 5.4A evaluates the extent to which steepness (k_{sn}) and wideness (k_{wn}) may scale with erosion rate. This result illustrates that erosion rate does not markedly influence k_{wn} , but exerts an expected important control on k_{sn} , following the power-law relationship $k_{sn} = 13E^{0.65}$. This $E - k_{sn}$ relationship is expected from first principles (e.g. Whipple, 2004; Lague, 2014), so we normalise k_{sn} by the predicted values from the best-fit function (yielding k_{sn-E}) for later analyses (Figure 5.4B and C). However, even after accounting for the influence of erosion rate, k_{sn-E} shows no statistically significant correlation with MAP (Figure 5.4B) or AI (Figure 5.4C). In contrast, we find that k_{wn} does exhibit a clear correlation with both MAP (Figure 5.4B) and AI (Figure 5.4C), with power law scaling exponents of -0.30 and -0.27, respectively. Our sensitivity analysis reveals similar results for k_{sn} and k_{wn} for drainage area normalization exponents a and b (Equations 5.1 and 5.2) equal to 0.3 or 0.4, indicating our results are robust (Table 5.S2).

5.4 Discussion

5.4.1 Threshold-dependent channel slope versus width sensitivity to discharge variability

Our results indicate that channel wideness (k_{wn}), but not steepness (k_{sn-E}), decreases with both increasing MAP and AI (Figure 5.4B and C). This directly supports the hypothesis that channel width, where banks are expected to have a higher discharge threshold for erosion, is more sensitive to discharge variability (as proxied by MAP and AI) than slope (Figure 5.1). From a qualitative perspective, our results suggest that channel width, and hence long-term bank erosion, increases with higher discharge variability, as indicated by lower MAP and AI (e.g. Rossi et al., 2016). This is conceptually consistent with Stark et al. (2010), who found that bedrock river sinuosity, driven by bank erosion, increases with the frequency of extreme rainfall (> 50 mm/day). This dynamic is also reported in alluvial rivers, as Phillips et al. (2024) found wider channels in rivers with a lower AI , and Leenman et al. (2025) linked accelerated channel migration and avulsion to high discharge variability.

From a quantitative perspective, our analysis suggests that $k_{wn} \sim AI^{-0.27}$ (Figure 5.4C). Notably, AI has previously been shown to scale with the discharge-variability parameter c as $c \sim AI^{0.44}$ (Rossi et al., 2016). Combining these empirical relations yields $k_{wn} \sim c^{-0.6}$; that is, lower c (higher discharge variability) is associated with greater k_{wn} . Likewise, MAP can correlate with c via its scaling with mean annual runoff (MAR , Rossi et al., 2016), producing comparable results (supplementary material). Thus, our analysis suggests k_{wn} depends markedly on c (or discharge variability). The natural range of the discharge variability parameter across fluvial-dominated

rivers in the USA is broadly an order of magnitude (~ 0.1 -1, [Rossi et al., 2016](#)), therefore encompassing a five-fold variation in k_{wn} for an exponent of -0.6. This means, if c decreases, for example from 0.35 to 0.30 (equivalent for AI to decrease from 0.88 to 0.62; [Figure 5.2B](#)), the channel at a given location is predicted to widen by $\sim 10\%$.

How do climate metrics (AI and MAP) control channel geometry via discharge variability? A key mechanism is the occurrence of infrequent but intense precipitation events in areas with lower AI and MAP (e.g. [Trenberth, 2011](#)), leading to higher discharge variability and enhanced erosion ([Figure 5.1C](#)). In contrast, discharge variability may increase by reducing moderate flows via evapotranspiration ([Rossi et al., 2016](#)). The link between flow reduction and erosion enhancement seems counter-intuitive, but may be possible if sustained flows increase particle interlocking and critical shear stress (e.g. [Masteller et al., 2019](#)). This is consistent with observations that rivers experiencing same extreme events could exhibit distinct erosion rates that correlate inversely, rather than positively, with MAP ([Gayer et al., 2019](#)). The two mechanisms cannot be distinguished here, but the sensitivity of channel width to discharge variability may help constrain them.

5.4.2 Implications for landscape evolution and erosion hazards

In general, our current understanding of landscape responses to climate change and discharge variability is heavily centred on channel slope (e.g. [Lague, 2005](#); [Marder and Gallen, 2023](#); [Molnar et al., 2006](#)). Yet here, our results demonstrate that compared to channel slope, width is more sensitive to discharge variability ([Figure 5.4B and C](#)). There are several common explanations for the apparent insensitivity of channel slope to discharge variability, including: i) the trade-off between discharge variability and MAR (e.g. [DiBiase and Whipple, 2011](#)), ii) an insufficient range in discharge variability under consideration (e.g. [Adams et al., 2020](#)), and iii) negligible erosion thresholds (e.g. [Lague et al., 2005](#)). The first two can be largely excluded, because channel width shares the same limitations yet remains sensitive to proxies for discharge variability. The third explanation is possible, but its likelihood is reduced here because rivers with exceptionally high erosion rates relative to global median were excluded (> 300 m/Ma; e.g. [DiBiase and Whipple, 2011](#)). In this case, we argue that channel widening with increasing discharge variability may have suppressed the expected slope response to discharge variability (cf. [Scherler et al., 2017](#)). Note, however, we emphasise that the apparent insensitivity does not contradict discharge variability control on channel slope (e.g. [DiBiase and Whipple, 2011](#)). Finally, the evidence for enhanced bank erosion with increasing discharge variability supports the hypothesis of [Barefoot et al. \(2022\)](#) that high discharge variability may boost fine sediment

supply via intensified channel mobility. Overall, our analysis represents a crucial complement to current slope-centred landscape dynamics.

This work also yields insights into potential erosion hazards likely to be driven by climate change in the coming decades. Taking the Mediterranean region as an example, rivers are likely to experience higher discharge variability (i.e. more intermittent) by 2100, as extreme precipitation may increase by ~20% in a warming world (e.g. McLeod et al., 2024; IPCC). High discharge variability corresponds to wider channels (Figure 5.4B and C), so rivers in the warming world are expected to erode their banks until new equilibrium states are approached. This implies increased bank failures and exacerbated flood risk in the near future and helps guide how much space should be reserved for expected widening. In contrast, the insensitivity of channel slope to discharge variability (Figure 5.4B and C) suggests that channel-bed erosion is more transient (cf. Slater and Singer, 2013).

5.5 Conclusions

We tested the hypothesis that channel width is more sensitive to discharge variability than slope using carefully selected rivers worldwide. Our results support this hypothesis by showing that channel wideness, but not steepness, correlates with mean annual precipitation and aridity index, both of which are effective proxies for discharge variability. Their distinct sensitivities may allow the width response to modulate the slope response to discharge variability, a dynamic not fully captured in many river incision and landscape evolution models (e.g. Lague, 2005; DiBiase and Whipple, 2011). Combined with climate projections that predict increasing river discharge variability, we suggest that enhanced channel widening driven by bank erosion in response to this signal is an under-recognised hazard exacerbating flood risk for many rivers with typical rainfall and erosion rates globally.

5.6 Acknowledgments

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5.7 Supplementary material

5.7.1 Flood distributions and erosion thresholds

To describe the magnitude-frequency distribution of floods, we followed Rossi et al. (2016) to use the stretched exponential distribution:

$$P(X > x) = e^{-\left(\frac{x}{x_0}\right)^c} \quad (5.S1)$$

where $P(X > x)$ is the exceedance probability of discharge x , x_0 is a scale parameter, and c is a shape parameter that characterises discharge variability. Lower values of c correspond to higher discharge variability. For a given erosion threshold Q_c , only floods with discharge exceeding Q_c contribute to erosion. The total threshold-exceeding discharge, defined as the integral of discharge weighted by its probability density, can be approximated by integrating the exceedance probability:

$$Q_{tc} = \int_{Q_c}^{\infty} P(X > x) dx \quad (5.52)$$

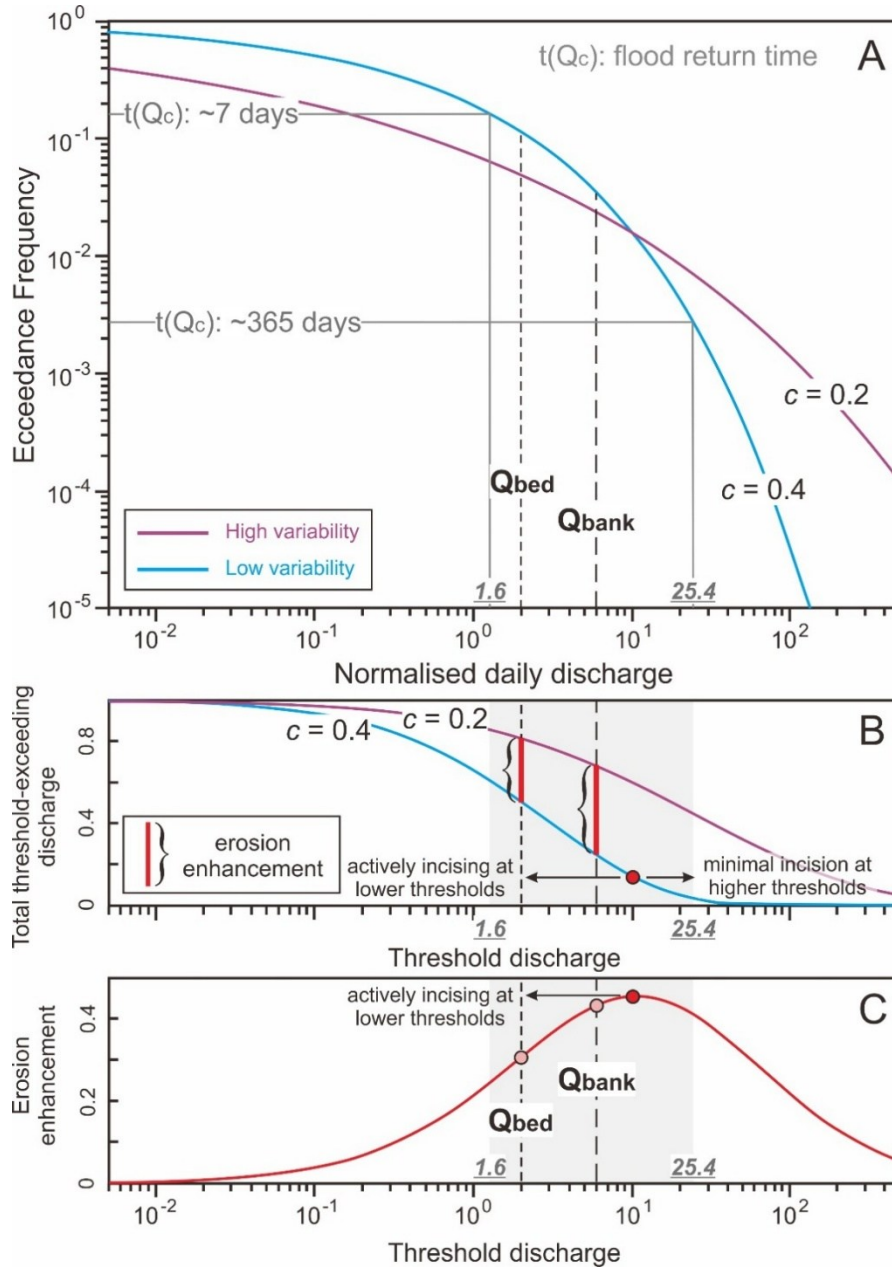


Figure 5.S1. (A) Two synthetic flood distributions with identical total discharge but distinct variability, parameterized by c ; (B) total threshold-exceeding discharge for both distributions, and (C) their difference as a function of threshold discharge Q_c . Gray lines in A and boxes in B and C constrain the common range of threshold discharges for $c = 0.4$, based on return times

reported by [Lague \(2014\)](#). For actively incising rivers ($Q_c < 10$), channel width is expected to be more sensitive to discharge variability than slope, because the bank threshold is higher than the bed threshold, as detailed in the main text (indicated as Q_{bed} and Q_{bank} in C).

To illustrate the impact of discharge variability and erosion thresholds on erosion, we constructed two synthetic flood distributions with two representative values of c (0.4 and 0.2; [Rossi et al., 2016](#); [Figure 5.S1A](#)). Both distributions have the same total discharge but differ in how that discharge is partitioned among small, moderate, and extreme events ([Figure 5.S1A](#)). Changing Q_c from 0 to infinite, we calculated the corresponding threshold-exceeding total discharge for both flood distributions, shown in [Figure 5.S1B](#).

Based on the common assumption that erosion scales approximately linearly with discharge (e.g. [Adams et al., 2020](#); [DiBiase and Whipple, 2011](#)), all else being equal, the calculated Q_{tc} can be used as a proxy for erosion. Thus, the difference in Q_{tc} between the two distributions (red lines in [Figure 5.S1B](#)) represents the impact of discharge variability on erosion, which we term erosion enhancement ([Figure 5.S1C](#)). This erosion enhancement depends on the magnitude of the erosion threshold, for a given change in the discharge-variability parameter c ([Figure 5.S1C](#)). For 10 well-constrained steady-state rivers across the globe, [Lague \(2014\)](#) estimated return times of threshold-exceeding floods (Q_c) between 7 and 365 days; in our case, these correspond to $1.6 < Q_c < 25.4$ for $c = 0.4$ ([Figure 5.S1A](#)). In this range of Q_c , erosion enhancement increases with increasing thresholds up to $Q_c \sim 10$, and then decreases slightly ([Figure 5.S1C](#)). Note, rivers with $Q_c > 10$ cause limited incision for $c = 0.4$ ([Figure 5.S1C](#)), and we expect rivers actively incising to balance uplift to have Q_c less than 10. In this case, for a given increase in discharge variability, higher thresholds lead to greater erosion enhancements, and thereby larger channel geometric adjustments ([Figure 5.S1C](#)). In turn, channel width is more sensitive to discharge variability than slope (compare Q_{bed} with Q_{bank} in [Figure 5.1C](#)).

5.7.2 Measurement of channel width and slope

For each selected river, the coordinates of the ^{10}Be sample were used to define a geographic extent expected to encompass the catchment, typically by extending $\pm 0.3^\circ$ in both longitude and latitude. This extent was then used to retrieve the 30m Advanced Land Observing Satellite (ALOS) DEM from OpenTopography via the function 'readopentopo' in MATLAB. For the retrieved DEM, we followed standard procedures to derive its drainage network, using a threshold area at 1 km^2 in TopoToolbox ([Schwanghart and Scherler, 2014](#)). We then snapped the ^{10}Be sampling location to the network, and the DEM was cropped to its upstream catchment. The resulting drainage area was compared with that reported in [Marder and Gallen \(2023\)](#) to ensure correct catchment extraction. This catchment boundary was then used to derive mean annual precipitation (MAP)

(WorldClim 2; [Fick and Hijmans, 2017](#)) and an up-to-date aridity index ($AI = MAP/\text{mean annual potential evapotranspiration}$) (Global-AI v3.1; [Trabucco and Zomer, 2009](#)).

To objectively select channel width measurement locations, we extracted the upstream drainage areas at the start and end nodes of the trunk channel. By dividing this range into 20 equal intervals, we obtained target drainage area values for each location. These values were then passed to the function ‘getlocation’ in TopoToolbox, which returns the coordinates along the trunk channel where a node has a drainage area closest to a specified value ([Schwanghart and Scherler, 2014](#)). Because tributary–trunk junctions introduce discrete jumps in drainage area, these nodes often do not match the target values exactly; consequently, we used these coordinates to extract the corresponding actual drainage areas. This procedure thus yields sampling locations at approximately 5% increments of cumulative upstream drainage area along the trunk channel ([Figure 5.S2A](#)). Where channels were identifiable in Google Imagery, we measured up to five, equally spaced (mostly ~50 m), cross sections at that location, depending on image quality ([Figure 5.S2B](#)). Following the approach in the main text, we plotted channel width against drainage area, coloured by E and MAP to illuminate possible controls ([Figure 5.S3](#)).

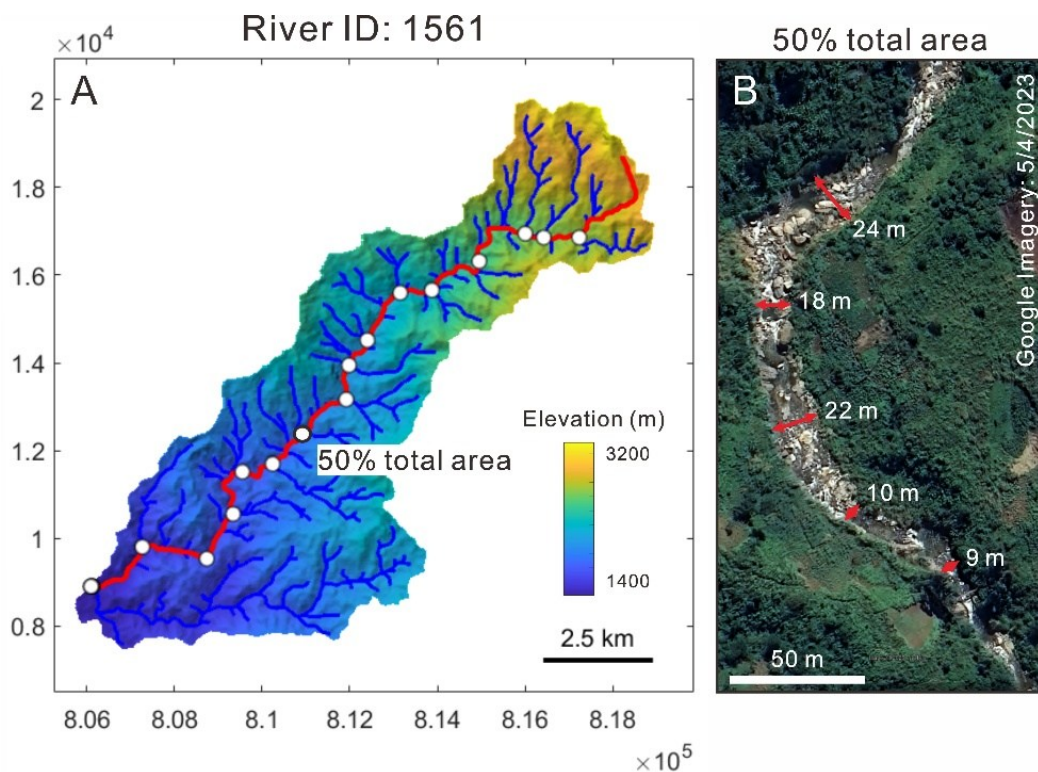


Figure 5.S2. Illustration of channel width measurement in an example river (A), with five measurements taken at the location corresponding to 50% of the total drainage area (B).

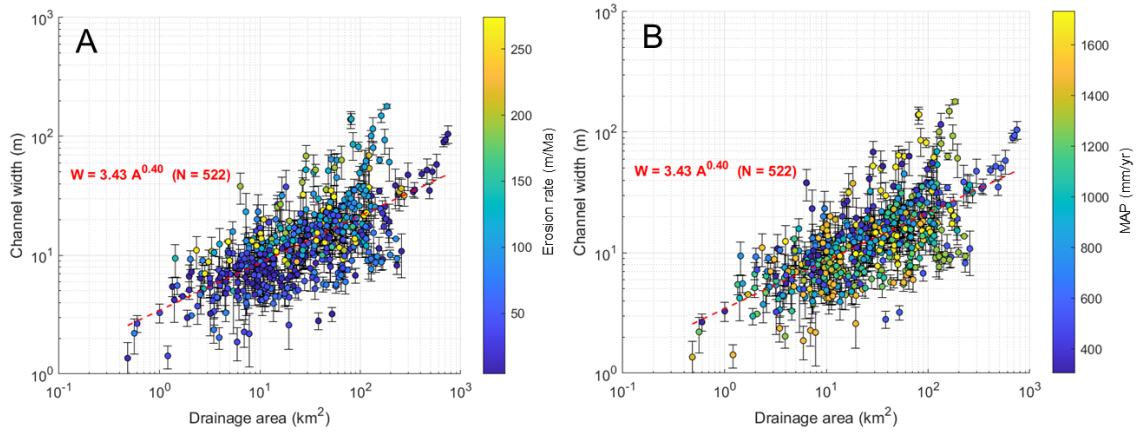


Figure 5.S3. Channel width versus drainage area, coloured by erosion rate (A) and mean annual precipitation (B).

To quantify area-normalised channel slope (k_{sn} , channel steepness), one approach is to regress through channel slope (S) and drainage area (A) using Equation 5.S3:

$$k_{sn} = SA^\theta \quad (5.S3)$$

where θ is the reference concavity (e.g. Wobus et al., 2006). This approach is often replaced by the χ method, in which the χ is defined as:

$$\chi = \int_{x_b}^x \left(\frac{A_0}{A_x} \right)^\theta dx \quad (5.S4)$$

where x_b is base level, and A_0 is a reference drainage area that is often chosen as 1 km² (Perron and Royden, 2013). Equation 5.S3, where $S = dz/dx$, can be integrated with respect to distance to generate a χ -elevation plot:

$$z = z(x_b) + k_{sn}\chi \quad (5.S5)$$

In this χ -elevation plot, k_{sn} is the gradient. Thus, we smoothed the elevation profile and calculated χ for each pixel (Schwanghart and Scherler, 2014). We then constrained pixel-scale k_{sn} from the elevation gradient in χ space using the TopoToolbox function ‘mchi’ (Equation 5.S5), and used its mean and standard deviation to characterise catchment-scale k_{sn} .

5.7.3 Sensitivity analyses and discharge variability control on channel width

To evaluate the robustness of the relationships between channel geometry and erosion rate, mean annual precipitation (MAP), and aridity index (AI), we conducted sensitivity analyses on the parameters a and b in Equations 5.1 and 5.2, respectively (main text). First, we set $a = 0.4$ and 0.3 to calculate channel steepness (k_{sn}). Following the approach described in the main text, we examined the E - k_{sn} relationship and derived the best-fit functions (P-value and exponents reported in Table 5.S2). We then obtained the derived metric k_{sn-E} and examined its

relationship with MAP and AI (Table 5.S2). As with $a = 0.5$, no statistically significant correlations were found (Table 5.S2). We next set $b = 0.4$ and 0.3 to calculate channel widthness (k_{wn}). Likewise, we evaluated E - k_{wn} relationships and derived best-fit functions (Table 5.S2). Because this correlation is not strong, we analysed both k_{wn} and the derived metric k_{wn-E} against MAP and AI , yielding similar exponents to those for $b = 0.5$ (Table 5.S2). Overall, the sensitivity analyses indicated that the relations reported in the main text are robust to the choice of a and b , as well as to the use of either k_{wn} or the derived metric k_{wn-E} .

To quantitatively interpret the impact of discharge variability on k_{wn} , we exploit the relationship among MAP , mean annual runoff (MAR), and the discharge-variability parameter c . In a comprehensive analysis of gauged rivers across the USA, MAR can be approximated as a linear function of MAP (Rossi et al., 2016). Combining this with the relation $c \sim MAR^{0.21}$ (Rossi et al., 2016) gives $c \sim MAP^{0.21}$. The best-fit function in Figure 5.4B gives $k_{wn} \sim MAP^{-0.30}$, suggesting $k_{wn} \sim c^{-1.43}$. This is stronger, but comparable to the one derived from AI in the main text (exponent -0.6).

5.7.4 Table 5.S1. Tectonic, climatic, lithological, and topographic characteristics of the selected rivers.

Basin ID	Publication	Latitude	Longitude	MAP		AI		E		ksn		kwn		Area (km2)	Lithology	
				Mean	STD	Mean	STD	Mean	STD	Mean	STD	Mean	STD		Plut.%	Meta.%
2249	Dixon, 2016	46.6100	15.1400	1285	130	1.88	0.30	72	15	253	105	1.52	0.67	142.3	0	100
2251	Dixon, 2016	47.1500	14.7800	1279	185	1.98	0.36	67	13	171	74	1.02	0.22	75.0	0	98
2256	Dixon, 2016	47.2700	14.6600	1493	195	2.52	0.47	76	15	221	95	0.94	0.19	71.2	0	99
2257	Dixon, 2016	46.8200	15.1800	1226	129	1.82	0.29	49	10	202	59	1.13	0.20	69.4	0	100
2259	Dixon, 2016	46.9400	14.8200	1289	196	2.02	0.42	92	19	209	79	0.96	0.37	232.0	0	94
2265	Dixon, 2016	46.5300	15.4300	1240	51	1.73	0.10	82	17	273	152	2.52	0.50	10.7	5	95
2579	Wittmann, 2016	45.9558	8.5268	1133	55	1.50	0.07	274	62	267	105	3.08	1.21	121.3	0	100
797	Duxburry, 2009	38.2000	-78.7900	1128	31	1.09	0.04	4	1	69	16	3.21	0.99	12.6	0	100
805	Duxburry, 2009	38.6500	-78.2400	1104	47	1.10	0.07	13	3	120	31	1.83	0.55	24.8	10	83
806	Duxburry, 2009	38.6100	-78.2600	1146	44	1.15	0.07	12	3	150	72	2.83	0.61	14.0	10	90
2791	Linari, 2016	35.3425	-82.1818	1347	48	1.18	0.05	11	3	58	41	2.28	0.33	17.4	0	100
2792	Linari, 2016	35.5419	-82.3828	1397	52	1.31	0.05	12	2	49	21	2.12	0.75	11.2	0	100
2798	Linari, 2016	35.5692	-82.2147	1342	35	1.23	0.03	23	5	73	21	3.20	1.29	10.8	0	100
199	Matmon, 2003	35.5200	-83.2900	1683	134	1.72	0.20	24	5	177	169	1.68	0.44	191.2	0	93
216	Matmon, 2003	35.7400	-83.1300	1654	140	1.68	0.19	31	6	222	84	1.75	0.25	66.3	0	97
222	Matmon, 2003	35.4600	-83.4300	1659	125	1.66	0.18	21	4	156	49	1.33	0.14	105.1	0	94
3429	Portenga, 2019	39.2107	-77.3108	1027	6	0.95	0.01	11	3	18	16	2.84	0.50	12.4	0	100
3430	Portenga, 2019	39.1543	-77.1320	1077	3	0.98	0.01	10	2	14	12	1.99	0.21	13.9	0	100
2836	Reusser, 2015	34.7400	-83.7700	1732	68	1.57	0.08	15	3	100	44	2.18	0.32	49.3	0	100
1565	Bookhagen, 2012	-24.8052	-65.6106	311	84	0.22	0.07	194	36	435	128	7.90	3.59	65.1	0	95
2384	Nóbile, 2016	-28.2200	-65.8800	381	52	0.24	0.03	111	21	491	167	4.22	2.20	140.5	0	100
2385	Nóbile, 2016	-28.0785	-65.9063	382	53	0.24	0.03	83	16	378	145	3.20	0.83	91.7	0	100
2386	Nóbile, 2016	-28.0775	-65.9067	417	54	0.26	0.03	49	9	280	126	2.75	0.74	56.7	0	100
2407	Siame, 2015	-31.5800	-67.8700	305	5	0.18	0.00	34	7	154	63	3.24	0.45	8.7	0	100
3397	Salgado, 2016	-22.6693	-44.5341	1632	110	1.59	0.15	20	5	239	51	2.85	0.35	22.2	7	93
2504	Molliex, 2016	42.6100	9.4400	888	37	0.81	0.16	45	10	200	148	2.59	0.64	67.6	14	85
2515	Molliex, 2016	42.4500	9.4700	810	46	0.76	0.13	110	25	176	77	3.21	0.51	113.7	0	100

2516	Molliex, 2016	42.2300	9.4400	793	71	0.76	0.12	61	15	235	190	2.86	0.57	65.8	0	100
2903	Vanacker, 2015	-3.0070	-78.5790	1119	118	1.35	0.23	86	17	282	178	2.74	0.65	87.3	0	100
264	Morel, 2003	47.9400	8.2000	1490	93	2.32	0.16	37	8	35	7	1.40	0.76	19.6	0	100
4338	Brocard, 2020	15.0600	-90.6000	1730	174	1.36	0.08	100	21	319	106	2.28	0.34	35.3	0	100
3037	Cyr, 2014	39.1139	16.5054	869	7	1.01	0.00	68	14	41	9	1.88	1.05	14.2	0	100
4243	Kim, 2016	38.0600	128.6100	1307	18	1.48	0.07	108	41	90	43	5.27	3.14	194.4	7	93
998	Heimsath, 2009	-12.4586	133.2891	1426	7	0.77	0.01	35	10	21	6	3.82	1.02	10.1	0	100
1551	Roller, 2012	0.6556	30.1658	1474	12	1.41	0.06	53	11	248	75	1.61	0.41	11.5	0	100
1554	Roller, 2012	0.6400	29.9800	1533	78	1.58	0.24	43	9	492	245	2.30	0.74	89.1	0	96
1560	Roller, 2012	0.1200	29.9500	1652	185	1.91	0.47	99	19	595	316	5.40	2.85	113.9	0	98
1561	Roller, 2012	0.0808	29.7487	1420	99	1.40	0.25	33	7	347	187	4.23	2.11	60.4	0	95
1563	Roller, 2012	0.2100	30.0100	1668	192	1.94	0.44	123	27	632	358	8.93	3.96	95.9	0	100
2855	Acosta, 2015	1.9100	36.8400	459	9	0.29	0.01	4	1	30	12	1.87	0.63	28.5	0	100
2527	Olivetti, 2016	45.2100	4.3800	1072	58	1.30	0.08	36	7	84	35	1.22	0.50	66.8	2	98
2538	Olivetti, 2016	44.9900	4.5300	1009	71	1.14	0.12	53	12	125	27	1.69	0.36	193.4	11	89
3354	Schmidt, 2016	21.9583	100.1804	1515	15	1.25	0.01	35	7	105	37	1.94	0.42	11.8	0	97
2146	Adams, 2016	27.5000	90.6600	908	104	0.91	0.08	71	13	219	65	2.55	0.52	46.5	0	100
2147	Adams, 2016	27.5100	90.6600	918	78	0.91	0.06	43	8	200	62	1.72	0.59	27.1	0	100
2148	Adams, 2016	27.4945	90.6655	954	84	0.96	0.06	105	21	213	76	3.80	1.55	26.4	0	100
2153	Adams, 2016	27.6100	90.6700	805	112	0.82	0.08	72	14	340	159	2.04	0.84	113.3	0	95
2159	Adams, 2016	27.4400	90.9600	841	42	0.87	0.04	38	7	95	26	3.23	0.69	20.5	0	100
2160	Adams, 2016	27.5256	89.8671	1040	156	0.95	0.12	71	14	443	117	1.33	0.32	114.4	0	100
2162	Adams, 2016	27.4560	89.9034	1125	128	1.05	0.11	113	22	473	119	2.63	0.55	73.1	0	94
2166	Adams, 2016	27.7267	91.1479	809	199	0.77	0.15	106	21	703	365	3.46	1.55	71.9	0	100
2178	Adams, 2016	27.5164	91.1712	984	150	0.93	0.14	212	40	600	229	3.27	1.83	273.8	0	92
2179	Adams, 2016	27.6641	91.4376	910	159	0.89	0.13	267	49	508	119	5.56	0.94	51.2	0	100
1979	Mallouf, 2015	27.6340	89.8009	937	187	0.89	0.13	201	40	518	126	1.78	0.36	69.1	0	100
1989	Mallouf, 2015	27.2338	89.4833	1221	220	1.23	0.18	39	7	406	244	1.85	0.10	23.0	0	91
2132	Olen, 2015	27.1200	87.2500	1515	115	1.24	0.16	158	32	402	76	2.63	0.54	84.2	0	100
2040	Portenga, 2015	27.2704	90.0371	1517	150	1.41	0.14	258	50	447	116	1.34	0.00	90.0	1	90
2054	Portenga, 2015	27.8672	89.8862	343	21	0.41	0.02	193	35	207	123	4.71	2.33	18.9	0	100

2065	Portenga, 2015	27.5295	89.8771	1041	158	0.95	0.12	125	24	438	119	1.56	0.43	114.7	0	100
4388	Heimsath, 2020	-23.0400	31.2400	492	8	0.32	0.01	4	1	12	4	2.79	1.10	9.9	14	86
4389	Heimsath, 2020	-23.1200	31.2900	549	40	0.37	0.03	4	1	26	14	0.94	0.70	257.3	10	89
4390	Heimsath, 2020	-23.0300	31.2800	564	56	0.37	0.04	5	1	34	26	2.47	0.68	746.5	1	98
2009	Morell, 2015	30.3818	79.3213	1539	246	1.33	0.26	259	62	419	122	4.89	1.08	10.9	0	100
1958	Scherler, 2014	31.2200	77.8300	1179	55	1.10	0.03	261	48	442	68	2.19	0.72	88.3	0	100
1835	Buscher, 2013	38.4182	28.2389	793	57	0.63	0.07	142	29	350	220	3.44	2.21	26.8	0	100
1837	Buscher, 2013	38.3012	28.1693	831	34	0.67	0.04	265	54	351	113	2.28	0.21	37.3	0	100
3704	Heineke, 2017	37.9300	27.7800	797	33	0.58	0.04	157	33	244	103	3.12	1.17	52.2	0	98
4048	Heineke, 2019	38.4300	27.7400	717	10	0.52	0.01	89	19	117	54	3.14	1.08	23.0	0	100
4057	Heineke, 2019	37.9400	27.6900	745	5	0.50	0.01	135	31	86	22	3.37	0.53	21.3	0	100
4062	Heineke, 2019	38.0600	27.9000	774	38	0.57	0.04	231	50	257	68	1.75	0.10	23.3	0	100
4064	Heineke, 2019	38.1551	28.3211	724	24	0.54	0.02	87	19	209	100	5.57	1.33	12.9	0	100
4065	Heineke, 2019	38.1177	28.2724	716	31	0.53	0.03	124	27	210	101	4.19	1.08	17.1	0	100

Due to the size of the dataset, channel-width measurements and drainage area at each location are not included ; the full dataset is available online at <https://doi.org/10.6084/m9.figshare.31991058.v1>

5.7.5 Table 5.S2. P-value and power-law exponents of the relations between channel geometry and tectonic and climatic parameters.

	P-value			exponent		
	E (Erosion rate)	MAP (Mean annual precipitation)	AI (Aridity index)	E (Erosion rate)	MAP (Mean annual precipitation)	AI (Aridity index)
ksn (a = 0.5)	<0.001	0.276	0.482	0.65	\	\
ksn-E (a = 0.5)	\	0.272	0.258	\	\	\
kwn (b = 0.5)	0.020	0.004	0.020	0.12	-0.27	-0.30
kwn-E (b = 0.5)	\	0.001	0.016	\	-0.28	-0.30
ksn (a = 0.4)	<0.001	0.255	0.421	0.62	\	\
ksn-E (a = 0.4)	\	0.237	0.193	\	\	\
kwn (b = 0.4)	0.005	0.006	0.024	0.14	-0.25	-0.29
kwn-E (b = 0.4)	\	0.002	0.017	\	-0.27	-0.29
ksn (a = 0.3)	<0.001	0.267	0.410	0.57	\	\
ksn-E (a = 0.3)	\	0.259	0.189	\	\	\
kwn (b = 0.3)	0.001	0.013	0.036	0.16	-0.23	-0.28
kwn-E (b = 0.3)	\	0.004	0.025	\	-0.25	-0.27

Note, p-value < 0.05 (in red) indicates that the correlation is statistically significant at the 95% confidence level and the exponent of its best-fit line is given.

6 DISCUSSION

Here, I summarise the key findings in relation to the aims and objectives outlined in section 1.3 and then discuss their broader implications for cross-cutting themes. Drawing on these results, I outline potential directions for future research.

6.1 Synthesis

In this thesis, I sought to examine tectonic, lithological, and climatic controls on upland rivers, using field, remote-sensing, and empirical approaches. These rivers are especially important due to their central roles in landscape evolution and source-to-sink systems ([section 1.1](#)). In particular, their channel geometry can help assess long-term seismic hazards linked to fault activity and erosional hazards driven by global warming ([section 1.1](#)). Yet, despite their importance, channel geometric responses to external forcings remain challenging to decipher. This challenge arises because (i) tectonics and lithology often co-shape landscapes, (ii) sediment supply, influenced by lithology, modulates channel responses to external forcing, and (iii) threshold-stochastic processes remain poorly calibrated against field data ([section 1.2](#)). To bridge these research gaps, I then developed four aims and multiple objectives ([section 1.3](#)), addressed in Chapters 2-5.

I first took a landscape-scale perspective in Chapter 2 to isolate tectonic and lithological effects in transient landscapes where spatial variations in climate are limited, using the classic SPIM (**Aim 1**). This constitutes a crucial yet rarely performed test of whether the classic SPIM is effective in landscapes co-shaped by tectonics and lithology, as is often the case across Earth's surface. I leveraged the well-constrained Gulf of Corinth, Greece, as a natural laboratory to demonstrate that lithology-calibrated fluvial metrics, such as knickpoint ages and vertical migration rates, yield interpretations that are quantitatively consistent with independent tectonic constraints. This consistency highlights that to first order, steepness-based lithological erodibility successfully unravels tectonic and lithological imprints encoded in transient river long profiles in Corinth. Building on this result, I further showed that lithology-calibrated metrics can extend local tectonic constraints into broader regions, allowing tectonic history to be refined at very high spatial and temporal resolution. This work demonstrated that sophisticated insights into fault evolution can now be gained from landscapes shaped by both tectonics and heterogeneous lithology.

In the context of topographic evolution, I then turned to its erosional products: sediment. In Chapter 3, I focused on two Corinth rivers in detail to assess the competing influence of tectonics

and lithology in sediment generation (**Aim 2**). I demonstrated that, in soft bedrock such as sandstone-siltstone, although tectonic knickpoints can coarsen hillslope sediment via mass wasting, this coarsening does not extend into channel sediment due to rapid abrasion. I therefore emphasised that tectonic perturbations alone do not guarantee the export of coarser sediments from perturbed catchments. I also suggested that lithology-driven grain size coarsening can extend beyond hillslopes and reach catchment outlets as channels steepen to maintain equilibrium. Taking the two Corinth rivers as an example, I clarified when and how lithology can outweigh tectonics in generating bedload sediment in catchments characterised by basement rocks fringed by sedimentary rocks. Finally, I shed new light on stratigraphy development by illustrating how tectonics and lithology may set the grain size of sediment exported from catchments to adjacent basins.

Based on the finding that lithology dominates bedload sediment generation, I further analysed how bedload sediment supply modulates channel responses to tectonics and lithology, using the two Corinth rivers as an example (**Aim 3**). In Chapter 4, I advanced the flux-dependent dynamics by presenting the first unambiguous field evidence favouring the $f(Q_s/Q_t)$ function proposed by [Turowski et al. \(2007\)](#) over that by [Sklar and Dietrich \(2004\)](#). This matters because the two competing functions predict distinct flux-dependent erosional dynamics and hence have fundamental implications for constraining landscape responses to external forcing. I demonstrated this by showing that the sediment-rich Phoenix river is much more sensitive to tributary input than the sediment-poor Sithas river. Based on this formulation, I further refined the role of flux-dependent dynamics in channel geometry by highlighting that Q_s/Q_t not only determines the excess shear stress required for channels to reach equilibrium with external forcing, but also influences how that excess shear stress is partitioned between channel width and slope adjustments. By doing so, I also underscored that channel slope and width work in concert to accommodate external forcings.

The importance of channel width motivated me to explicitly account for width variations in examining climate control on erosion (**Aim 4**). In Chapter 5, I contrasted channel width and slope responses to mean annual precipitation and aridity index, both of which can be used as effective proxies for discharge variability. This analysis is of considerable societal relevance, given that global warming is expected to increase discharge variability in many rivers in the near future. Notably, I found that channel width, but not slope, correlates with these two discharge variability proxies. This result challenges the traditional slope-centred perspective for examining climate control on erosion. I suggested that the high sensitivity of channel width to discharge variability may reflect the high threshold required to erode the channel banks. Moreover, I

argued that channel widening with discharge variability may have suppressed the expected slope response to discharge variability. I further emphasised that enhanced storminess in a warming world may drive many rivers across the globe to widen, promoting bank failures, intensifying channel mobility, and exacerbating flood risk in the near future.

Overall, I demonstrated how tectonic and lithological effects on landscapes can be isolated using the classic SPIM. In this context, I clarified how flux-dependent dynamics allows lithology-influenced bedload sediment supply to modulate channel geometric responses to external forcing. I also highlighted that the impact of erosional thresholds and stochastic floods on erosion is better captured by variations in channel width and wideness, rather than by changes in slope and steepness.

6.2 Cross-cutting themes

6.2.1 Tectonic versus lithological controls on channel geometry

The Corinth rift exemplifies tectonically active and lithologically heterogeneous landscapes that are common at Earth's surface. In such landscapes, isolating tectonic and lithological controls on channel geometry is challenging, but essential for understanding the relative importance of tectonics and lithology in shaping landscapes and extracting robust tectonic signals from topography (e.g. Gailleton et al., 2021; Gallen, 2018; Bernald et al., 2019). From the perspective of classic SPIM, the common strategy is to account for lithological effects by estimating lithological erodibility and then to derive tectonic signals from lithology-calibrated channel steepness (e.g. Shi and Shao, 2014; Gailleton et al., 2021; Takahashi et al., 2022). In Chapter 3 and 4, I have shown that lithology modulates channel geometry via generating bedload sediment. This provides new, flux-dependent insights into how lithological impact can be accounted for and when tectonic signals can be isolated from topography (cf. Kirby and Whipple, 2012). Below, I compare common methods for estimating lithological erodibility and discuss ways to improve current steepness-centred topographic interpretation.

I have presented several constraints on lithological erodibility in this thesis. First, in Chapter 2, I use channel steepness to calibrate landscape-scale lithological erodibility (K). I show that the relative erodibilities of limestone, conglomerate, sandstone-siltstone, and poorly consolidated sediment— $K_{lime} : K_{cong.} : K_{sand-silt} : K_{p-con sed.} = 1 : 2 : 3 : 4$ —capture the first-order variations in rock erodibility. Second, in Chapter 3, I estimate uniaxial compressive strength (UCS) using Schmidt hammer measurements. The results indicate that the median rock strengths of the limestone, flysch, conglomerate, and sandstone-siltstone are approximately 70, 29, 26, 9 MPa respectively. Based on the correlation between UCS and tensile strength, TS , such that $UCS \sim 10 * TS$, and

the relationship $K \sim TS^{-2}$ (Sklar and Dietrich., 2001), the observed differences in *UCS* can be translated into relative erodibilities of $K_{lime} : K_{flysch} : K_{cong.} : K_{sand-silt} = 1 : 6 : 7 : 60$. Third, in Chapter 4, I calculate excess shear stress using field-based channel-geometry measurements and found a ~ 3 times difference between upstream limestone and flysch reaches and the downstream sandstone-siltstone reaches in the Phoenix catchment. This is comparable to the threefold difference between limestone and conglomerate in the adjacent Vouraikos catchment, reported by Zondervan et al. (2020) using a similar approach.

In general, the three approaches give qualitatively consistent results, but the rock-strength approach yields much greater contrasts in erodibility among lithologies. This discrepancy may arise because different rock types have varying degrees of large-scale heterogeneity (e.g. fracturing) that rock-strength measurements fail to capture. In fractured limestone (e.g. Wicker et al., 2024), for example, the effective bulk erodibility of the rock mass may be markedly underestimated by rock-strength measurements at local scale. This effect may be less pronounced in sandstone-siltstone, likely exaggerating the erodibility contrast with limestone in rock-strength-based estimates. Interestingly, relative erodibility estimated from steepness variations is comparable to that derived from excess shear stress or unit stream power (Zondervan et al., 2020), both of which incorporate not only channel slope but also width. This similarity implies that, at a landscape scale, Corinth rivers respond to external forcings mostly by adjusting channel slope, rather than width. Combined with the conceptual model in Chapter 4, this dynamic occurs when both channel banks and bed lie in zone 3, where erosion efficacy $f(Q_s/Q_t)$ approaches a constant. As a result, I suggest that excess shear stress or unit stream power is the recommended approach to estimate lithological erodibility. In rivers mantled by sediment, steepness-based erodibility may serve as a simple yet effective means. The rock-strength-based erodibility, however, appears to be effective in some cases (e.g. Haag et al., 2025) but less so in Corinth and in other areas (e.g. Gallen et al., 2015), likely depending on the length scales and magnitudes of heterogeneities in the rock types examined.

To date, most studies have used channel steepness or its associated metrics, such as knickpoints and relief, to decode tectonics from topography (e.g. Wobus et al., 2006; Miller et al., 2012). This approach is built upon the assumption that channels mostly increase their slope to raise excess shear stress to a level that balances tectonic uplift (e.g. Kirby and Whipple, 2012). In lithologically heterogeneous landscapes, however, lithology-influenced bedload sediment supply can determine the excess shear stress required to maintain equilibrium and influence how that excess shear stress is partitioned between width and slope adjustments, as demonstrated in Chapter 4. For example, I suggest that zone 2 rivers mostly adjust channel width,

rather than slope, to balance tectonic uplift (e.g. downstream Phoenix river). These rivers are therefore, in nature, less likely to steepen channels and develop knickpoints (e.g. Lavé and Avouac, 2001; Yanites, 2018). In turn, a lack of knickpoints or significant relief does not necessarily indicate that a river is tectonically less active than others (compare, for instance, the Phoenix and Sithas rivers presented in this study, which experience similar tectonic uplift rates, but show markedly different landscape responses; see Table 3.1). This is the same with excess shear stress: rivers with low excess shear stress (wide and gently-sloping channels) may be tectonically more active than rivers with high excess shear stress, when their erosion efficacy is much higher (e.g. Cowie et al., 2008). Thus, I conclude that, in lithologically heterogeneous landscapes, excess shear stress estimates that incorporate both width and slope are a better metric than steepness alone for mapping tectonic activity. Moreover, I suggest neither performs particularly well where the impact of flux-dependent erosion efficacy is important, highlighting the need to identify suitable rivers for analysis; the criteria for selecting such rivers have been outlined in Chapter 4.

6.2.2 Emerging perspectives on channel-width dynamics

A prominent theme in this thesis is the importance of channel width in accommodating external forcings, such as tectonics and climate. In Chapter 4, I show that two Corinth rivers differ markedly in their channel geometrical responses to comparable gradients in bedrock erodibility and uplift rate: the sediment-rich Phoenix river mostly adjusts width, except in its upstream zone; in contrast, the sediment-poor Sithas river predominantly adjusts slope. This result suggests that channel width responds to both bedrock and tectonic uplift, especially when sediment supply is high. In addition, channel-width response to discharge variability has been revealed clearly in Chapter 5. Hence, I demonstrate that channel width is not simply a power-law function of drainage area or discharge, as commonly assumed (e.g. Whipple and Tucker, 1999; Lague et al., 2005), but instead adjusts dynamically to sediment flux and discharge variability.

First, by comparing the two well-constrained Corinth rivers with distinct relative sediment flux, I highlighted the fundamental role of flux-dependent dynamics in regulating channel width. This finding explains why numerical models based on the classic SPIM, in which flux-dependent dynamics and stochastic floods are not explicitly incorporated, have not always succeeded in capturing channel-width response to tectonic uplift in real rivers (Finnegan et al., 2005; Wobus et al., 2006; Turowski et al., 2007). This is evidenced by the inconsistency between the modelled power-law scaling between channel width and uplift rate, with a simple exponent of ~ 0.23 , and observed complex relations ranging from strong negative power-law correlations to little or no

relation (e.g. Lavé and Avouac, 2001; Snyder et al., 2003; Duvall et al., 2004; Whittaker et al., 2007). This complex range of relationships is reproduced, to my knowledge, only by the model of Yanites and Tucker (2010); however, that model is based on two assumptions with limited physical justification: (i) overly simplified flux-dependent incision dynamics, in which sediment only inhibits channel-bed erosion, and (ii) channels reach a steady-state width at which slope is minimised. In Chapter 4, I bridged this gap by proposing a conceptual model based on the constrained relative sediment-flux formulation $f(Q_s/Q_t)$ of Turowski et al. (2007) and the equilibrium principle that steady-state channel width is reached when bank erosion balances bed erosion. In Chapter 5, I further show that stochastic floods can also shape channel width via discharge variability. This relationship has been rarely explored in either nature or numerical models, and the empirical relation derived here can guide future modelling efforts. Yet, how flux-dependent dynamics and stochastic floods combine to control channel width remains elusive, and I aim to explore this topic in 6.3.1.

In Chapters 4 and 5, I suggest that width dynamics modulate slope adjustment to external forcing, likely leading to landscape behaviours distinct from those modelled by simple width-discharge scaling (e.g. Whipple and Tucker, 1999; Lague et al., 2005). This effect has been well illustrated in model space by Attal et al. (2008), who compared footwall catchment behaviours under two types of width dynamics: a static width dependent solely on discharge and a dynamic width dependent on both discharge and slope. Their results showed that when dynamic width adjustment is allowed, transient landscapes exhibit gentler slopes, shorter response times, and a lower likelihood of drainage reversal by fault-induced back-tilting. In addition, channel narrowing with increasing slope markedly reduces the threshold discharge required for erosion in steep reaches. This led Lague (2014) to suggest that the impact of threshold-stochastic processes is negligible in reaches steepened by knickpoints but important elsewhere. This spatial variability is essential for reconciling two features commonly observed in natural rivers: the linear migration of knickpoints (i.e. slope exponent $n=1$) and non-linear relation between channel steepness and erosion rate (i.e. $n > 1$) (Lague, 2014). In general, both modelling studies illustrate that the full range of landscape behaviours in nature may only be captured when channel-width dynamics are incorporated. I note, however, that width dynamics in both models were either empirical or based on the classic SPIM. The improved understanding of channel-width controls demonstrated above thus paves the way for more accurate modelling of complex landscape behaviours. For example, threshold-dependent width dynamics suggest that climate change may enhance bank erosion via increased discharge variability, leading to enhanced sediment and nutrient export by accelerated mobility, rather than by enhanced discharge.

6.2.3 Rethinking stream-power incision models

In this thesis, I have employed three SPIM variants to investigate tectonic, lithological, and climatic controls on channel geometry. The first is the classic SPIM, in which erosion scales simply with erodibility, drainage area, and channel slope (see section 1.2.1). In contrast, the second variant explicitly incorporates flux-dependent dynamics, whereas the third emphasises threshold-stochastic processes (see section 1.2.1). The three SPIM variants and other incision models have been examined against well-constrained rivers to test which model works best in upland rivers (e.g. van der Beek and Bishop, 2003; Tomkin et al., 2003; Valla et al., 2010; Ferrier et al., 2013; Lague, 2014). The emerging consensus is that no single model can provide a universal solution. This may not be surprising, as each model emphasises a particular aspect of the general SPIM and so its success depends on the relative importance of that aspect in a given setting (e.g. Lague, 2014). Hence, I discuss below the possible mechanisms underlying the effectiveness of the three SPIM variants in this thesis.

In Chapter 5, I demonstrated that channel width is more sensitive to discharge variability than slope, and that channels widen with increasing variability. Both patterns are consistent with the threshold-stochastic model, in which a given increase in discharge variability can cause greater erosion enhancement when thresholds become higher. This implies that the impact of theoretically expected flux-dependent dynamics is limited, and I argue that this result may arise because all surveyed rivers are dominantly incising metamorphic rock and therefore have a similar potential to generate bedload sediment. In contrast, in Chapter 4 I highlighted that flux-dependent dynamics shape channel geometry, suggesting a limited role for threshold-stochastic processes in the adjacent two Corinth rivers. I suggest that this limited effect may arise from the highly intermittent nature of the Corinth rivers: their geomorphic work is done during extreme storms, when potential variations in erosion thresholds caused by sediment grain size and flux differences (Chapter 3) are likely to be negligible (McLeod et al., 2024). This likely also contributes to the success of the classic SPIM in disentangling tectonic and lithological effects on Corinth rivers, as shown in Chapter 2. A second condition favouring this success is high sediment supply, or zone 3 conditions, under which channel erosional efficacy becomes insensitive to Q_s/Q_t variations and excess shear stress is accommodated mostly by slope adjustments (e.g. Lavé and Avouac, 2001; Yanites et al., 2010).

Overall, I argue that the effectiveness of SPIM variants is case-dependent, and so more emphasis should be placed on clarifying the conditions under which a model can be employed. This was pursued in Chapter 4, where I suggested that the width-to-depth ratio and the sensitivity of

excess shear stress to tributary input could be used to identify the Q_s/Q_t zone in which a river lies; under zone 3 conditions, the classic SPIM performs reasonably well despite not explicitly incorporating flux-dependent dynamics. However, future work still needs to establish tangible criteria to assess when threshold-stochastic processes can be considered negligible. An encouraging step was taken by [Lague \(2014\)](#), who suggested that the role of threshold-stochastic processes can be markedly reduced by channel narrowing in knickpoint-steepened reaches, allowing the classic SPIM to be employed in modelling knickpoint propagation.

6.3 Future work

6.3.1 Proposal 1: how may lithology modulate climate control on erosion?

6.3.1.1 Motivation

In Chapter 5, I focus on rivers incising >80% metamorphic rock and found that channel width, but not slope, correlates with mean annual precipitation (*MAP*) and aridity index (*AI*). This is consistent with the higher erosion thresholds for the channel banks than the bed; however, the potential role of flux-dependent dynamics remains elusive. In Chapter 4, sediment supply has been suggested to control channel geometry by determining the shear stress required to maintain equilibrium and by influencing how that shear stress is partitioned between width and slope adjustments. Combined with the finding in Chapter 3 that bedload sediment supply is influenced by lithology, lithology may modulate climate control on erosion via influencing bedload sediment supply.

In recent years, climate control on erosion has been demonstrated with respect to *MAP*. This is well-illustrated by [Leonard et al. \(2023a\)](#), who show that channel steepness, when weighted by *MAP*, correlates more strongly with erosion rates and performs better in resolving lithologic variations. In cases where similar conclusions are made (cf. [Burbank et al., 2003](#)), the reported power-law exponents between *MAP* and erosion rates range from ~ 0.5 (e.g. [Bookhagen and Strecker, 2012](#); [Ferrier et al., 2013](#); [D'Arcy and Whittaker, 2014](#)) to ~ 1 (e.g. [Moon et al., 2011](#); [Adams et al., 2020](#); [Leonard et al., 2023a](#); [Leonard et al., 2023b](#)). In general, the lack of a consistent correlation appears to support the idea that climate control on erosion is modulated by other factors. In a comprehensive analysis of steady-state rivers across the Himalaya, [Shao et al. \(2026\)](#) found that the tendency for rivers to become significantly steeper under lower *MAP* and fewer extreme rainfall events is evident in those incising sedimentary and metasedimentary rocks, but less so in those incising crystalline rocks. This pattern points to a lithological control, but the underlying mechanism is yet to be clarified. Moreover, the possible climate control on channel width and its impact on erosion have not been examined.

6.3.1.2 Methods and analysis

To tackle these research questions, I plan to build upon Chapter 5, where well-constrained rivers incising > 80% metamorphic rocks have been examined. I propose to contrast these rivers with those incising > 80% fine-grained siliciclastic sedimentary rocks, given that the two lithologies are likely to differ in bedload sediment supply, as shown in Chapter 3. As in Chapter 5, I will limit my analysis to fluvial-dominated, steady-state rivers with well-constrained erosion rates.

To this end, I will first draw on the existing global dataset of standardised ^{10}Be -based erosion rates E (OCTOPUS; Codilean et al., 2022) and choose unglaciated rivers draining >10 km² to ensure fluvial dominance. I will then restrict the analysis further to rivers in a morphologically steady state, based on criteria: (i) absence of major knickpoints, and (ii) river profiles in each catchment collapsing into a straight line in χ plots ($R^2 \geq 0.9$) (e.g. Marder and Gallen, 2023). In this selection, I will also ensure that rivers cover multiple climate zones with distinct MAP (WorldClim 2; Fick and Hijmans, 2017) and AI (Global_AI_PET; Trabucco and Zomer, 2009). Finally, I will derive the catchment lithology (GLiM; Hartmann and Moosdorf, 2012) and focus on those dominantly incising fine-grained siliciclastic sedimentary rocks (> 80%).

To characterise climate control on erosion, I will follow the approach developed in Chapter 5 to measure channel width from high-resolution Google Earth imagery (e.g. Maxar and Airbus imagery, with pixel sizes of ~30–50 cm) and slope from DEM. I will calculate channel steepness ($k_{sn} = SA^a$, where a is a constant) and channel wideness ($k_{wn} = SA^b$, where b is a constant). As in Chapter 5, I will also examine whether and to what extent k_{wn} and k_{sn} correlate with MAP and AI in rivers incising > 80% fine-grained siliciclastic sedimentary rocks. I will then compare these results with those of Chapter 5 to test the hypothesis that lithology modulates climate control on erosion.

To further examine the mechanism underlying this lithological modulation, I will quantify the controls on erosional efficiency (K). Erosional efficiency can be calculated as:

$$E = KA^m S^n = K k_{sn}^n, \quad (6.1)$$

Here, the erosional efficiency is often considered as a function of bedrock erodibility and precipitation ($K = K_p P^m$) (e.g. Adams et al., 2020; Ferrier et al., 2013). I hypothesise that the extent to which precipitation increases K depends on bedload sediment availability, which is regulated by hillslope landslides. Following Moon et al. (2011), I will quantify the fractions of the catchment area susceptible to shallow landsliding, using the model developed by Montgomery and Dietrich (1994). This hypothesis can then be tested by determining the relationship among

precipitation, landslide fractions, and K . I will also examine whether narrower channels correspond to higher K via increased stream power, all else being equal.

6.3.2 Proposal 2: can erosional dynamics be discerned in stratigraphy?

6.3.2.1 Motivation

Upland rivers convert intact bedrock into sediment cascades (e.g. [Allen, 2008](#); [Whittaker et al., 2010](#)). The resulting sediment flux, grain size, and composition determine the material available for deposition in downstream basins (e.g. [Armitage et al., 2011](#); [Sharman et al., 2019](#)). This rationale motivates the work in Chapter 3, where I provide the upstream erosional evidence to support the hypothesis that bedrock unroofing in upland catchments has the potential to shape stratigraphy (e.g. [Carroll et al., 2006](#); [McArther et al., 2013](#)). I also show that footwall uplift acceleration can coarsen sediment grain size via knickpoints (e.g. [Attal et al., 2015](#)) and hypothesise that such tectonically coarsened sediment may impact stratigraphy once knickpoints propagate into hard bedrock (e.g. [Armitage et al., 2011](#)). Yet, the hypothesised link between upland erosional dynamics and downstream stratigraphy has rarely been substantiated in the same source-to-sink systems.

Evaluating this source-to-sink link in natural systems is a fundamental challenge in Earth science. Numerical models with different set-ups and assumptions yield distinct outcomes (e.g. active, buffered, shredded), highlighting the need for calibration against field observations ([Watkins et al., 2019](#)). For the fault-bounded system in Baja California Sur, Mexico, [Mortimer and Carrapa \(2007\)](#) related clast compositional changes in hangingwall stratigraphy to inferred footwall bedrock unroofing and argue for a short ~ 30 -40 k.y. lag time; however, their work does not provide direct constraints on erosional catchments. If this short lag time does hold, then much can be learnt from stratigraphy. For example, [Litty et al. \(2017\)](#) used detrital zircons in preserved stratigraphy to infer past climate changes from temporal variations in erosion patterns. This also opens up opportunities to refine current tectonic-stratigraphic models, where the characteristics of sediment supply are poorly incorporated (cf. [Gawthorpe and Leeder, 2000](#); [Elliott et al., 2012](#)).

6.3.3.2 Methods

I suggest testing this source-to-sink link in footwall-bounded small source-to-sink systems, for which the theoretically rapid response (e.g. [Helland-Hansen et al., 2016](#)) puts a conservative estimate on source-to-sink signal propagation time. Moreover, I plan to focus on knickpoint-associated erosional dynamics, because knickpoints initiated by fault uplift acceleration are

expected to be ubiquitous (e.g. Roda-Boluda and Whittaker, 2018; Wrona et al., 2022), and their erosional dynamics are tractable, as demonstrated by knickpoint propagation models and mass-wasting distribution in Chapters 2 and 3. These insights enable us to predict how sediment composition and grain size may evolve over time, as knickpoints propagate upstream, rejuvenating topography and shaping erosion processes (e.g. Whittaker et al., 2010). This prediction can then be compared with hangingwall stratigraphy to answer whether and to what extent erosional processes and stratigraphy are coupled.

In detail, I propose to test the source-to-sink link in the Gulf of Corinth, where upland topographic evolution can be reconstructed (e.g. Zhou et al., 2024) and Pleistocene Gilbert deltas with relatively good time constraints are currently well-exposed (e.g. Rezwan et al., 2025). For example, the Kerenitis delta has been divided into 11 stratal units, representing ~100 kyr of glacial-interglacial cycles (Backert et al., 2010). I plan to first reconstruct the past locations of knickpoints at 100 kyr intervals and focus on the periods surrounding their traversal of lithological boundaries. I hypothesise that these transitions markedly alter sediment composition and grain size, likely propagating this signal downstream into the stratigraphy. For the time steps before and after these transitions, I will constrain their sediment characteristics. Based on the finding that the slope exponent n is ~1 in the Corinth rift, I will estimate erosion rates upstream of knickpoints (E_{up}) from the steepness ratio across knickpoints (k_{sn-up}/k_{sn-dn}) and downstream erosion rates (E_{dn}) ($E_{up} = E_{dn} * (k_{sn-up}/k_{sn-dn})$; e.g. Whittaker and Walker, 2015). I will also use the average elevation of the knickpoints as a contour to reconstruct the areas above and below the knickpoints (A_{up} and A_{dn}). Based on erosion rates and areas, I will predict lithological composition and grain size at each time step by mixing sediment generated by different rock types at varying erosion rates. For example, the limestone flux Q_{lime} can be estimated as $Q_{lime} = A_{up-lime} * E_{up} + A_{dn-lime} * E_{dn}$. Finally, I plan to collect both sediment grain size and composition data for different stratal units in the field. In this process, I will distinguish different grain size fractions (sand versus gravel) to evaluate the impact of grain size on signal propagation.

6.3.3.3 Analysis

To evaluate the extent to which the transient erosional dynamics are transmitted into stratigraphy, I plan to compare predicted sediment grain size and composition with measured values. In addition, I can compare the sensitivity of grain size and composition to external forcings. This work will further our understanding of signal propagation in source-to-sink systems in rift basins.

7 CONCLUSIONS

Upland rivers are central to landscape evolution and source-to-sink systems, but their dynamics remain challenging to quantify, limiting long-term understanding of seismic and erosion hazards. I address this challenge using the family of stream power incision models (SPIMs) and an empirical approach that combines DEM, field, and remote sensing analyses. In this thesis, I first focus on rivers draining into the Gulf of Corinth, Greece as a natural laboratory to unravel tectonic versus lithological effects on long-term topography, and to examine the sediment-mediated lithological control on channel geometry. I then use carefully selected rivers across the world to investigate the response of river erosion to discharge variability. The main conclusions of this thesis are:

1. Landscape records both tectonic and lithological effects; lithology-calibrated fluvial metrics yield robust constraints on fault growth and linkage processes in the southern margin of the Gulf of Corinth, Greece, when lithological effects are accounted for using steepness-based lithological erodibility.
 - In general, segment-scale metrics are more sensitive to lithological differences than their catchment-scale counterparts. Combined with a well-constrained, ca. 700 ± 100 ka knickpoint in the Vouraikos catchment, segment-scale steepness variations yield lithology-dependent erodibilities.
 - To first order, the lithology-dependent erodibilities account for lithological impacts, as supported by two observations: (i) river long profiles, only when calibrated by lithological erodibility, obtain optimal concavities that fall within the theoretical range; and (ii) lithology-calibrated knickpoint metrics yield interpretations consistent with independent tectonic constraints.
 - Lithology-calibrated metrics provide geomorphic constraints on tectonic processes, allowing the existing tectonic models to be refined at high spatial and temporal resolution. For example, along-strike variations in knickpoint age appear to reflect the eastward growth of the Eliki fault and the linkage of the Derveni and Lykoporia faults.
2. In two Corinth rivers, the Phoenix and Sithas, lithology outweighs tectonics in generating bedload sediment due to abrasion. This finding may apply to many catchments in which basement rocks are fringed by sedimentary rocks, with important implications for stratigraphic development in such settings.

- Tectonic perturbation alone does not guarantee the export of coarser sediments from tectonically perturbed catchments, unless knickpoints are initiated and migrate into abrasion-resistant bedrock.
 - Lithology-driven grain size coarsening can extend beyond hillslopes and reach catchment outlets, because selective transport in channels approaching equilibrium likely only attenuates, but not obliterates, coarsening signals.
 - Fault acceleration and bedrock unroofing likely generate characteristic stratigraphic architectures by influencing the grain size of sediment exported to neighbouring depocenters.
3. Bedload sediment supply modulates channel morphological responses to external forcings by determining the excess shear stress required for equilibrium and influencing how that shear stress is partitioned between width and slope adjustments.
- The finding that the sediment-rich Phoenix river, with a high relative sediment flux ($Q_s/Q_t \sim 0.8$), is characterised by a nearly constant erosional efficacy provides the first unambiguous field evidence favouring the asymmetric tools-and-cover formulation of [Turowski et al. \(2007\)](#) over the model of [Sklar and Dietrich \(2004\)](#).
 - The flux-dependent importance of channel width and slope adjustments in accommodating external forcing is captured by the proposed conceptual model. It indicates that certain rivers are, in nature, less likely to steepen and develop knickpoints (e.g. downstream Phoenix); as a result, a lack of knickpoints or significant relief does not necessarily indicate that a river is tectonically less active than others.
 - Flux-dependent incisions can be approximated by the classic stream power incision model (SPIM), where Q_s/Q_t remains constant or Q_s/Q_t variations does not markedly change erosion efficacy, as in the Phoenix river. The latter can be identified by examining the width-depth ratio and excess shear stress response to sediment input.
4. Based on a global dataset of fluvial-dominated, steady-state catchments with similar lithology and with erosion rates and precipitation typical of most rivers worldwide, channel width is shown to be more sensitive to discharge variability than slope, likely due to erosion thresholds are higher for channel banks than for channel beds. The result provides a new, width-centred perspective to characterise climate control on erosion.

- For 72 rivers selected across the globe, their channel width, but not slope, correlates with mean annual precipitation and aridity index, both of which can be used as effective proxies for discharge variability.
- The contrasting sensitivities of width and slope suggest that width adjustment may suppress the expected slope response to discharge variability, a dynamic not yet fully incorporated in many river incision and landscape evolution models.
- As ongoing climate change is projected to increase discharge variability in many rivers, channel widening through bank erosion is expected in the near future, leading to promoted bank failures, intensified channel mobility, and exacerbated flood risk.

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