

Tectonically- and climatically-driven mountain-hopping erosion in Central Guatemala from detrital ^{10}Be and river profile analysis

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Abstract. The rise of a mountain range affects moisture circulation in the atmosphere and water runoff across the land surface, modifying the distribution of precipitation and drainage patterns in its vicinity. Water routing in turn affects erosion on hillslopes, and incision in river channels on surrounding mountain ranges. In Central Guatemala, two parallel, closely spaced mountain ranges formed during two consecutive pulses of uplift, the first between 12 and 7 Ma (Sierra de Chuacús-Sierra de las Minas), and the second after 7 Ma (Altos de Cuchumatanes). We explore the climatic and tectonic processes through which the rise of the most recent range drove the slowing of river incision and hillslope erosion over the previously-uplifted range. ^{40}Ar - ^{39}Ar dating of perched volcanic deposits documents the sequential rise and incision of these mountain ranges. Terrestrial cosmogenic ^{10}Be in river sediments indicates that currently, hillslopes in the older range erode more slowly than in the younger range (20-150 vs. 300 $\text{m}\cdot\text{My}^{-1}$). These differences mimic the current distribution of precipitation, with the younger range intercepting the atmospheric moisture before it reaches the older range. River channel steepness and deformation of paleovalleys in the new range further indicate that the younger has risen faster than the older range up to today. We review how atmospheric moisture interception and river long-profile adjustment to the rise of the new range have contributed to the decline of erosion rates over the old range. We also explore the consequences of this decline and of aridification on the topographic evolution of the older range. The older range undergoes a slow topographic decay, dominated by backwearing, by the stacking of slowly-migrating erosion waves along the mountain flanks, and by the formation of

pediments around its base. The morphology of the old range is therefore transitioning from that of a front range to that of a dry interior range.

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1. Introduction

A mountain range affects both the circulation of atmospheric moisture ~~circulation~~ around its relief, and the flow of precipitated water over the land surface on its slope and in the vicinity. Moisture rises and precipitates on its windward side, while rain shadows tend to be cast over its lee-side and over the land surface located downwind (e.g. Meijers et al., 2018; Galewsky, 2009). Once precipitated, the fraction of water that runs off as overland flow drives hillslope erosion and river incision. River drainages are dynamic systems that can transmit forcing along drainage lines in both the downstream and upstream directions. Changes in climate and vegetation modulate hillslope erosion, and these changes are then transmitted to rivers from the headwaters downstream. Conversely, changes affecting downstream river reaches, such as the adjustment of river gradient to the rise of mountain ranges, or to variations in sea level, can be transmitted upstream along river channels (Humphrey and Heller, 1995; Whittaker and Boulton, 2012). From there, they are transmitted uphill along valley slopes (Harvey, 2002; Mudd and Furbish, 2007). Mountain ranges affect the erosion of surrounding reliefs through this combination of top-down and bottom-up processes.

The growth of contractional orogens commonly involves the outward, sequential propagation of contraction, and the formation of successive, in-sequence mountain ranges. Moisture, by contrast, is commonly advected in the opposite direction, from the forelands to the orogen interiors. Precipitated water is then commonly returned to the foreland by river networks that flow from the orogen interiors to the forelands. The rise of front ranges therefore commonly takes place both upwind and downstream of pre-existing ranges, driving the aridification of previously uplifted ranges (Garcia-Castellanos, 2007). The new front ranges will also affect the rivers that drain the previously uplifted ranges toward the forelands. These rivers will adjust to enhanced rock uplift by enhanced incision, through the steepening of their gradient in areas of enhanced uplift (Leland et al., 1998). This steepening is accompanied by a transient decrease in incision rate, upstream of the rising ranges (Champel et al., 2002). In some cases, the topography and underlying tectonic structure will respond fast enough to reach a new dynamic equilibrium between relief, climate, and tectonics (Willett and Brandon, 2002; Whipple and Meade, 2006) without any substantial alteration of the drainage network. In other cases, however, the range-transverse river networks will experience reorganization before the equilibrium is reached (Jackson et al., 2002; van der Beek et al., 2002; Brocard et al., 2012). In a few cases even, the slowing down of landscape response time, as a result of aridification, upstream and downwind or the rising ranges will lead to the disintegration of river drainages, and to the topographic decay of interior ranges. This evolution characterizes the nucleation and growth of orogenic plateaus formed by lateral accretion (Sobel et al.,

2003; Garcia-Castellanos, 2007). These different evolutionary pathways have been explored at the scale of entire orogens. They are, however, seldom documented at the scale of individual mountain ranges, because their manifestation at that scale is harder to separate from more local signals driven by spatial variations in bedrock erodibility, stochastic processes (e.g. landslides), and topographic inheritance.

We document here how the rise of a recent mountain range (the 170 km-long Altos de Cuchumatanes, or AC range) affected the topographic evolution of a nearby, older range (the 220 km-long Sierra de Chuacús-Sierra de las Minas, or SC-SM range) in Guatemala (Fig.1). Sharp topographic, climatic, and tectonic gradients in this region exert themselves over a relatively small (350 x 100 km) area, allowing a detailed investigation of the interactions between the two ranges. The SC-SM range started to rise first in early late Miocene time, when its flanks were deeply eroded (Brocard et al., 2011). The AC range, on the other hand, started to grow in the latest Miocene, next to the SC-SM range. The rise of the AC range sparked widespread drainage rearrangement in front of the SC-SM range during the late Miocene (Brocard et al., 2011). Numerous captures occurred during this event. Interestingly, these captures did not generate upstream-migrating waves of accelerated erosion. On the contrary: river incision stalled upstream of the capture sites (Brocard et al., 2011). We seek to identify the processes that arrested river incision, and retrieve, from the characteristics of river long-profiles, insights on how arrested incision influenced the topographic evolution of the SC range.

New $^{40}\text{Ar}/^{39}\text{Ar}$ ages on volcanic rocks are used first to tighten the chronology of river incision and surface uplift of the SC-SM range. Detrital terrestrial ^{10}Be erosion rates then provide a snapshot of current spatial variations in hillslope erosion rates in the SC-SM and AC ranges. Profile linearization is implemented to study the complex long-profiles of the rivers that drain these ranges. River knickpoints are extracted, and then classified in order to identify river knickpoints that transmit the long-term landscape response of the drainage to the sequential uplift of the two ranges. We then discuss the respective contribution of topographically-controlled climate and of tectonically-controlled river profile adjustment to the decline of incision over the SC range during the rise of the AC range. We finally review topographic characteristics of the SC range directly related to the slowing down of erosion over the SC range.

2. Origin and evolution of the mountain ranges of Central Guatemala

2 .1. Tectonics and orogenesis

Left-lateral motion along the North American-Caribbean plate boundary in Central Guatemala has produced elongate ranges parallel to the plate boundary (Fig.1a). We investigate the growth and erosion of two of these ranges, namely the Sierra de Chuacús - Sierra de las Minas range (SM-SC range), and the Altos de Cuchumatanes range (AC range, Fig.1b). Rocks in the SC-SM range possess a deeply penetrative, sub-vertical tectonic fabric, imparted by 70 My of left-lateral wrenching along

the Caribbean-North American plate boundary (Ratschbacher et al., 2009; Ortega-Gutierrez et al., 2004; Ortega-Obregón et al., 2008). Since Eocene time, left-lateral motion has been accommodated by the Motagua fault, and, to a lesser extent, by
90 the Polochic fault (Fig.2b). The Motagua fault is, with >1,100 km of total cumulative offset, the active subaerial fault with the largest cumulative offset on Earth. The Polochic fault has a total offset of 125 ± 5 km (Burkart, 1978); it probably branches out of the Motagua fault offshore, somewhere in the Caribbean Sea (Fig.1a), before running on land at an average distance of 50 km from the Motagua fault. Strain across the plate boundary is strongly partitioned between almost pure left-lateral slip on the Motagua and Polochic faults, and dip-slip on faults parallel to the Polochic and Motagua faults
95 (Authemayou et al., 2011b; Brocard et al., 2012).

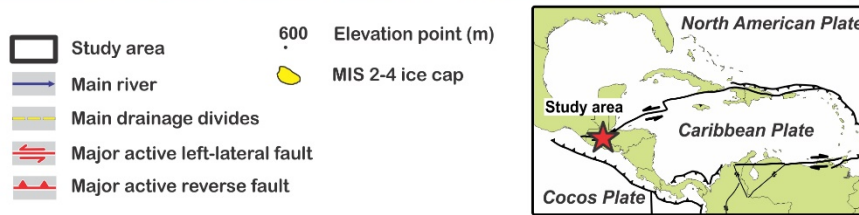
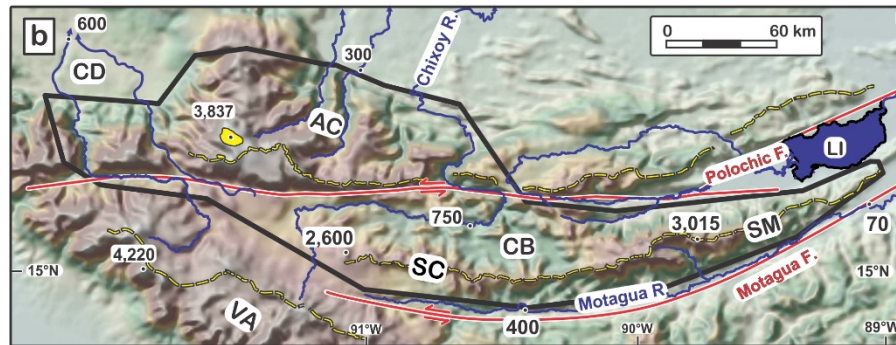
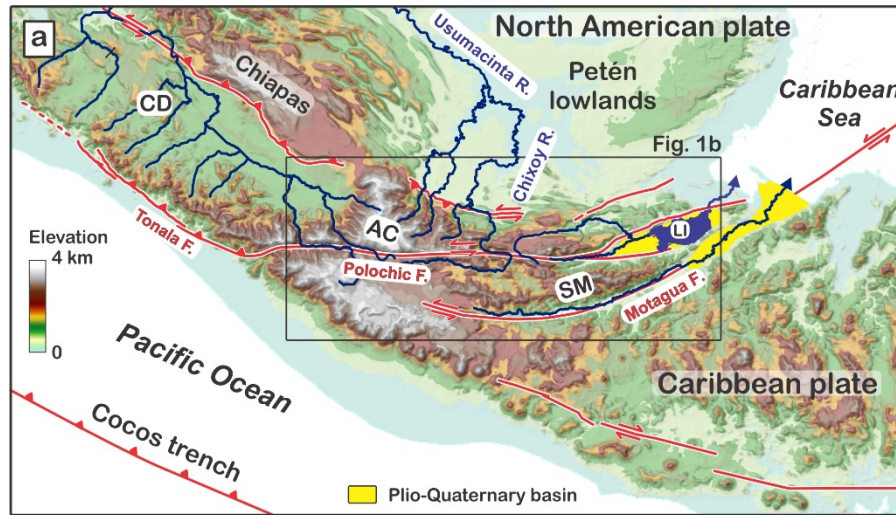


Figure 1. Shaded topography of the study area, showing the tectonic setting of Central Guatemala (a), and of the studied range (b). Topographic features: CB: Chixóy River basin, CD: Central Depression of Chiapas, AC: Altos de Cuchumatanes (AC range), LI: Lake Izabal, SC: Sierra de Chuacús (SC range), SM: Sierra de las Minas (SM range), VA: Central American Volcanic Arc. MIS: $\delta^{18}\text{O}$ Marine Isotopic Stage

Today, Central Guatemala is straddled by 3-4 km-high ranges separated by deep valleys. The floors of these valleys stand at elevations as low as 0.2-0.8 km (Fig.1b). In Middle Miocene times, the topography of Central Guatemala was much more

subdued. Remnants of that past topography (referred to as the Maya surface (Brocard et al., 2011)) still cap numerous mountaintops across the study area (Fig.2). They are separated by regions where the Maya surface has been deeply incised (Fig.3). The Maya surface formed from the topographic decay of Eocene folds (Authemayou et al., 2011b; Brocard et al., 2011). It grades to the east and north into lowlands, near the Caribbean Sea, indicating that it formed near sea level. Its uplift started after the Middle Miocene, affecting both the southern (Simon-Labric et al., 2013), and northern side of the Motagua fault, as far north as the Polochic fault (Brocard et al., 2011). This event saw the rise of the SC-SM range during the late Miocene. Valleys up to 1,000 m deep were incised within the northern flank of the SC-SM range between 12 and 7 Ma (Brocard et al., 2011).

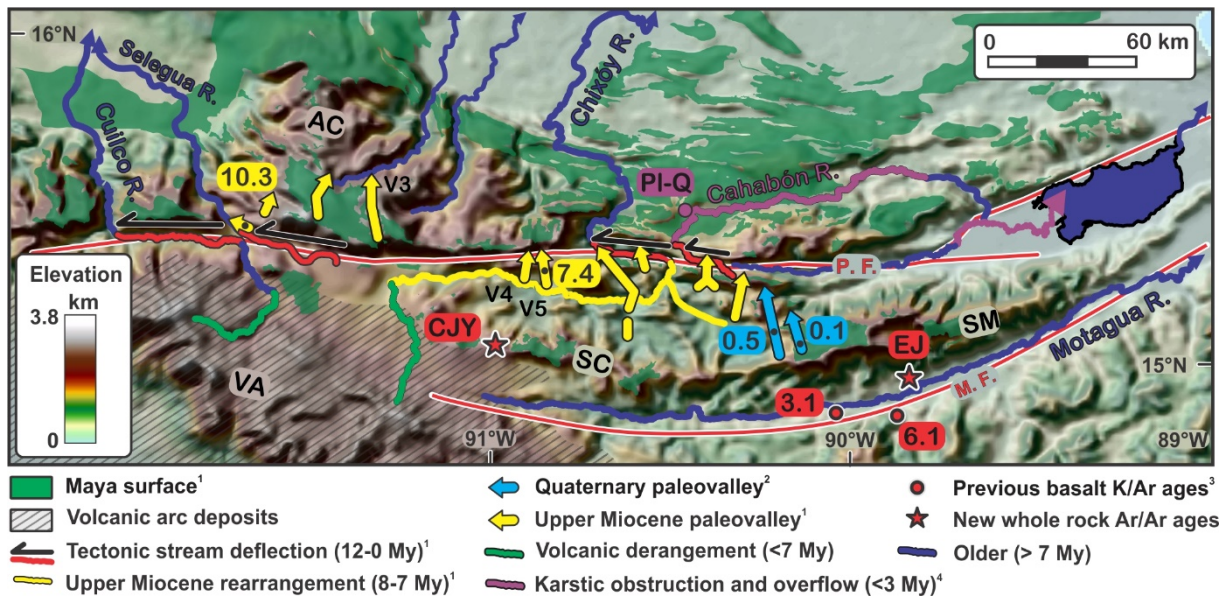


Figure 2. Age of geomorphic markers and drainage lines in Central Guatemala. Ages of Miocene valleys and Quaternary paleovalleys (V1-V12) in My. Data source: 1: (Brocard et al., 2011), 2: (Brocard et al., 2012), 3: (Tobisch, 1986), 4: Plio-Quaternary lacustrine deposits (Brocard et al., 2015a). Newly dated lavas: CJY: Chujuyúb, EJ: El Jute. Range names: AC: Altos de Cuchumatanes, SC: Sierra de Chuacús, SM: Sierra de las Minas, VA: volcanic arc. Faults: MF: Motagua, PF: Polochic. Background: shaded GTOPO 30 DEM.

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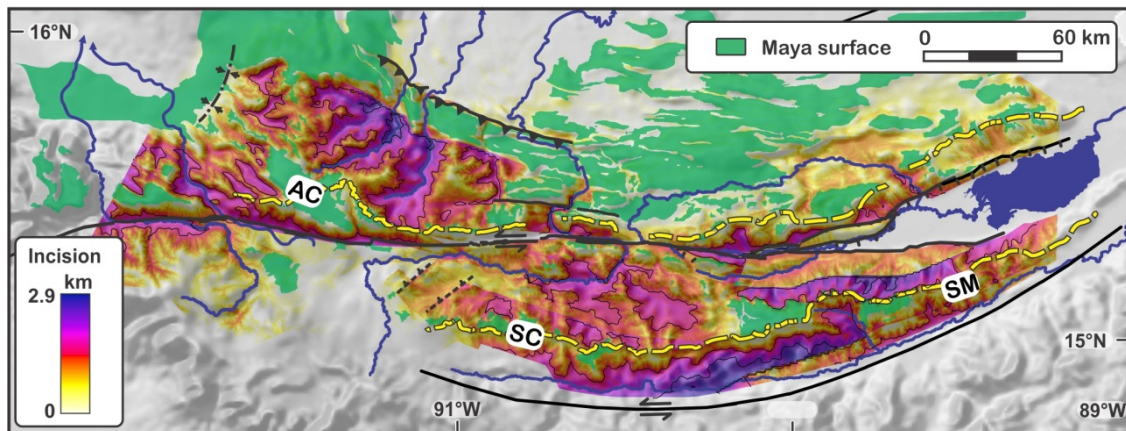


Figure 3. Incision below the Middle Miocene Maya surface, based on the elevation of surface remnants (upland relict surface). Incision
 125 contour line spacing: 1km. Yellow dashed lines: range drainage divides .AC: Altos de Cuchumatanes, SC: Sierra de Chuacús, SM: Sierra
 de las Minas, VA: volcanic arc.

Uplift propagated north of the Polochic fault during the Late Miocene (Brocard et al., 2011) and was marked by the
 rise of the AC range in response to contraction within the North American plate (Authemayou et al., 2011a). The rise of the
 130 AC range drove widespread reorganization among the rivers that drain the northern flank of the SC-SM range (Fig. 2).
 Numerous river valleys were then abandoned and left stranded on the rising AC range. Their deformation indicates that the
 AC range has risen >1-2 km relative to the SC range over the past 7 My (Brocard et al., 2011). Earthquake focal mechanisms
 further indicate that the tectonic structures bordering the AC range to the north still accommodates shortening today
 (Guzmán-Speziale, 2010; Authemayou et al., 2011b).

135 While contraction has defined the evolution of the western part of the study area, transtension has prevailed further
 east since the Late Miocene at least. The dominance of transtension in the east results chiefly from an eastward increase in
 the divergence angle between the strike of the plate boundary and the direction of plate motion (Rogers and Mann, 2007).
 Transtension led to growth of the Lake Izabal basin (Fig.1a), which is has been filled with ~5 km of terrigenous sediments
 since the Middle Miocene (Carballo-Hernandez et al., 1988; Bartole et al., 2019). Another ≥ 1.4 km-deep, elongate (125x15
 140 km) sedimentary basin formed over the same period at the eastern termination of the subaerial trace of the Motagua fault,
 next to the Caribbean Sea (Carballo-Hernandez et al., 1988) (Fig.1a). Transtension spread further to the west during the
 Pliocene, generating normal faults that disrupt the northern flank of the SM range (Authemayou et al., 2011b; Brocard et al.,
 2012).

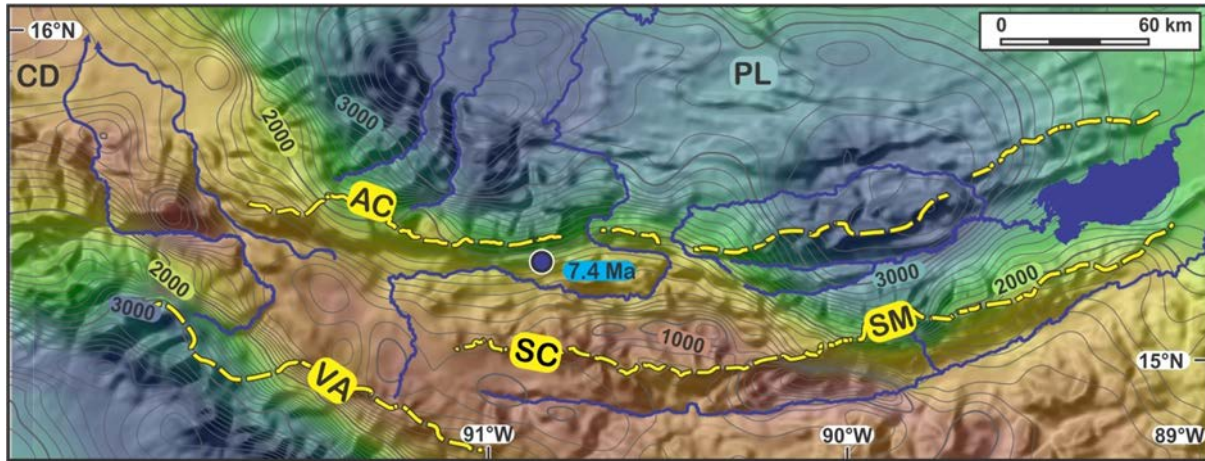
145 2.2. Drainage evolution since the Middle Miocene

The rivers located along the northern flank of the SC range represent the headwaters of a network that, farther downstream, experienced widespread reorganization during the late Miocene (Fig.2). Reorganization led to the formation of range-parallel rivers halfway between the drainage divides of the SC-SM and the AC ranges. These E-W-striking rivers collect the rivers that flow north, down the northern flank of the SC-SM range, and funnel them into the Chixóy River, one of the few rivers, together with the Cahabón, Chixóy, Selegua, and Cuilco Rivers (Fig.2), that still cross the AC range. These later rivers also straddle the trace of left-lateral Polochic fault, before entering the AC range. The Polochic fault has deflected and lengthened their course (Fig.2) since the late Miocene (Brocard et al., 2011). Plio-Quaternary transtensional faulting along the northern flank of the SM range initiated a second (and still ongoing) pulse of drainage reorganization during the Quaternary (Fig.2) (Brocard et al., 2012).

Since the Pliocene, large volcanoclastic aprons have piled up along the NE side of the Central American Volcanic arc. They have buried the western end of the SC range, deranging its river network (Fig.2). This complex area is therefore excluded from the present study. Likewise, the karstic highlands of Central Guatemala, especially those located north of the Polochic fault, are also excluded from the analysis, because their dynamics is strongly influenced by the high-frequency opening and closure of subterranean karstic pathways (Brocard et al., 2015a; Brocard et al., 2016a).

2.3. Current pattern of precipitation

Moisture tracking from the Pacific Ocean and from the Caribbean Sea is intercepted by slopes that face the western and eastern coasts of Guatemala, and the Petén lowlands in the north (Fig. 4). The AC range receives 4-6 m.yr⁻¹ of mean annual precipitation (MAP) along its northern flank in the Zona Reina (Thattai et al., 2003). In the west, the Central American Volcanic Arc intercepts moisture rising from the Pacific Ocean. In the east, moisture from the Caribbean Sea is channeled along the Lake Izabal basin and then rises up along the northern flank of the SM range, where fog interception represents a substantial part of the annual precipitation above 2,000 m (Holder, 2004). The volcanic arc, the SM range, and the AC range cast rain shadows over the Corridor Seco (Machorro, 2014), and in particular over the SC range, which receives little precipitation. Semi-arid climate conditions are met in the valleys that surround the SC range.



175 Figure 4. Mean annual precipitation across the study (MARN, 2017) from dry (red) to wet (blue). Isohyet spacing: 100 mm, draped over the shadowed GTOPO 30 DEM. Yellow dashed lines: range drainage divides. Blue dot: location of the fossil forest of Sicaché, buried below a 7.4 My-old ignimbrite (Brocard et al., 2011). CD: Central Depression of Chiapas, AC: Altos de Cuchumatanes, PL: Petén lowlands, SC: Sierra de Chuacús, SM: Sierra de las Minas, VA: volcanic arc.

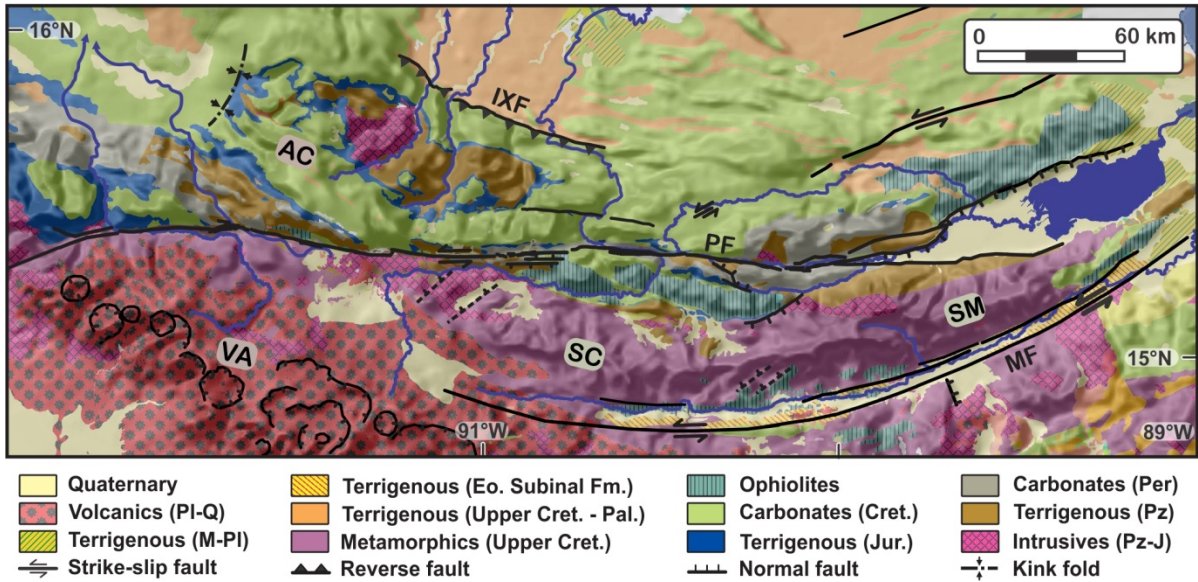
180 2.4. Bedrock lithology

Rock belts in Central Guatemala tend to follow the strike of the mountain ranges (Fig. 5). Late Cretaceous schists and gneisses of the Chuacús Fm. form the core of the SC-SM range. They are flanked by the late Cretaceous migmatites of the San Agustín Fm., and by marbles and amphibolites of the Jones Fm. In the north, this metamorphic core is tectonically
 185 juxtaposed, across the Baja Verapaz shear zone, to the basement of North America, which is covered mostly by a Permian megasequence of terrigenous sediments and carbonates (Sacapulas, Tactic-Esperanza, and Chochal Fms. (Anderson et al., 1973)). The basement and cover are intruded by Paleozoic (e.g. Rabinal), Triassic, and Jurassic (e.g. Matanzas) plutons.

A megasequence of continental terrigenous sediments (the Todos Santos Fm.), Cretaceous carbonates, and Cretaceous evaporites (Cobán Fm., Campur Fm.) covers much of the AC range (Fig. 5). Ultramafic rocks obducted over the
 190 carbonates in Late Cretaceous (Campanian) time (Fourcade et al., 1994) are preserved within weakly metamorphic synformal klippen (Baja Verapaz, Santa Cruz, and Juan de Paz ophiolites). Higher grade serpentine mélanges crop out along the Motagua valley (Flores et al., 2013).

The southern base of the SC-SM range is incised into sediments deposited in narrow transensional basins along the Motagua fault (Ratschbacher et al., 2009). During Eocene times (Newcomb, 1975), one such basin was filled by
 195 continental red beds of the Subinal Formation (Fig.5), which has an exposed thickness of $\geq 1,500$ m (Hirschman,

1962). Some of its detrital sediments are derived from the current basement of the SC-SM range, while others were brought by an axial river that prefigures the current Motagua River (Gutiérrez, 2008). The Subinal Fm. now lies in tectonic contact against the SC-SM basement along high-angle reverse faults (Muller, 1979; Bosc, 1971a).



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Figure 5. Geology and structure of Central Guatemala (Instituto Geográfico Nacional de Guatemala; Instituto Hondureño de Geología y Minas; Instituto Nacional de estadística y geografía de México), draped over the GTOPO 30 DEM. AC: Altos de Cuchumatanes, IXF: Ixcán fault, MF: Motagua fault, PF: Polo chic fault, SC: Sierra de Chuacús, SM: Sierra de las Minas, VA: volcanic arc.

205 3. Methods

3.1. ^{40}Ar - ^{39}Ar radiometric dating of volcanic rocks

The age of the low-relief Maya surface was previously constrained by bracketing age markers, such as 14-15 Ma overlying ignimbrites on the Chortís Block, and a 10 Ma ignimbrite filling a paleovalley, incised into the Maya surface (Brocard et al., 2011). To improve the dating of the Maya surface we resorted to the ^{40}Ar - ^{39}Ar dating of andesite boulders, embedded in a lahar deposit that lies on the Maya surface at the western termination of the SC-range, near the locality of Chujuyúb (Fig. 2). To establish the chronology of incision of the Motagua valley, we used previously dated alkaline basalts that crop out on the floor of the Motagua valley (Tobisch, 1986). We also dated a basalt flow that hangs 500 m above the Motagua River, in the foothills of the Sierra de las Minas, near the town of El Jute (Bosc, 1971b) (Fig. 6). Two ^{40}Ar / ^{39}Ar whole-rock ages were

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retrieved from the basalt of El Jute, and one whole-rock age was obtained on the basaltic andesite of Chujuyúb by the U.S. Geological Survey (USGS) in Denver, CO, USA (see Supplement 1).

3.2. Terrestrial ^{10}Be erosion rates

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We measured the concentration of ^{10}Be in quartz grains extracted from soils and river sediments. We used the ^{10}Be concentration to calculate hillslope erosion rates, integrated over the past 10^3 - 10^4 years (see Supplement 2). Soil and rock samples, collected along ridgelines in the SM range, provide erosion rates restricted to the sites of sampling (Table S2-1, Fig. S2-3c). Three samples consist of quartzose vein fragments exhumed from weathered orthogneiss, and two other samples
225 come from outcrops of highly weathered pegmatite located on the monadnock of Cerro las Palomas, in the Montaña El Imposible (Fig. S2-3b).

The majority of the samples consist of riverborne quartz collected in 30 rivers that drain the SM, SC, and AC ranges (Fig. 6, Table S2-2, Fig. S2-3). They provide catchment-averaged hillslope erosion rates (Brown et al., 1995). Quartz was extracted from the sand grain-size fraction (250-500 μm) of the river sediments. Because ^{10}Be production increases rapidly with
230 elevation, systematic altitudinal variations in the concentration of quartz in the source rocks may skew the calculation of erosion rates. A sensitivity analysis to this effect was conducted in the AC range, where this situation in encountered, by weighting ^{10}Be production according to the quartz concentration in the source rocks (Fig. S2-3a). Besides, the intense weathering of tropical mountains promotes the concentration of quartz in the soils and to the underestimation of erosion rates. An assessment of the effect of quartz enrichment on erosion rates was therefore conducted (Table S2-2), using quartz
235 enrichment values from mountain tropical soils of Puerto Rico (Ferrier et al., 2010). River sampling was conducted in nested catchments (Fig. 6. Fig. S2-3a,b) in order to capture along-stream variations in erosion rates, such as those produced by headward-migrating knickpoints (Willenbring et al., 2013b; Brocard et al., 2015b). Samples were prepared at the ^{10}Be extraction laboratory of the Department of Geology and Geophysics at the University of Minnesota, and at the PennCIL lab of the Earth and Environmental Sciences department at the University of Pennsylvania (see Supplement 2).

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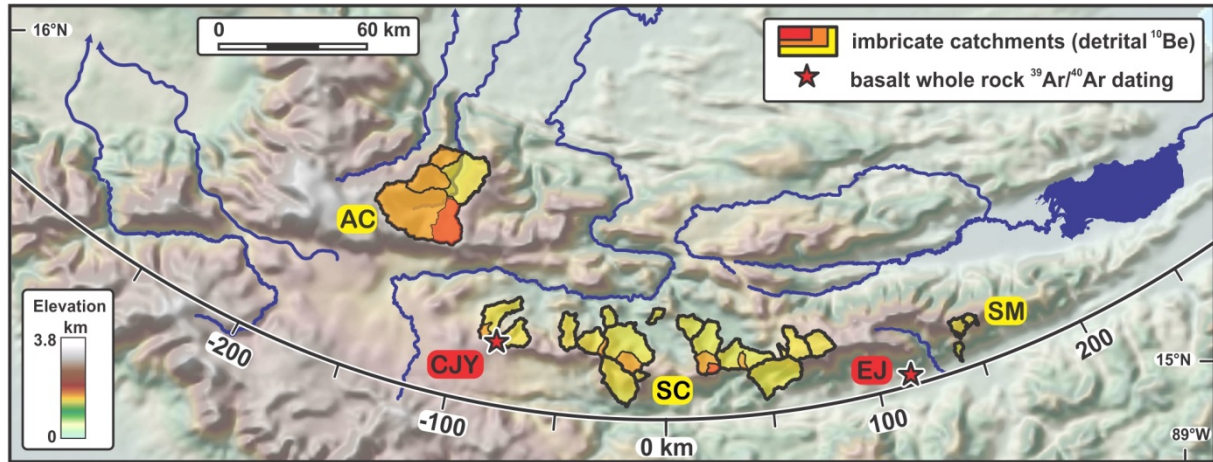


Figure 6. Catchments sampled for the ^{10}Be analysis (AC: Altos de Cuchumatanes, SC: Sierra de Chuacús, SM: Sierra de las Minas), and $^{40}\text{Ar}/^{39}\text{Ar}$ dating (CJK: Chujuyúb, EJ: El Jute). Enlarged maps of the catchments and their lithologies are provided in Fig. S2-1). The arcuate line represents the axis used for plate boundary-parallel projections of Figs. 10 and 12.

3.3. Calculation of an erosion index

To test the influence of hillslope steepness and of precipitation on ^{10}Be -derived erosion rates, we calculated a normalized erosion index (Montgomery and Stolar, 2006; Finnegan et al., 2008) over the study area. We used a formulation in which soil erosion is assumed to be proportional to shear stress: Eq. (1):

$$EI = Q^{1/3} \cdot S^{2/3} \quad (1)$$

where Q is discharge (in $\text{m}^3 \cdot \text{s}^{-1}$) and S the along-slope gradient ($\text{m} \cdot \text{m}^{-1}$). Slope was extracted from the national Guatemalan IGN DEM at a resolution of 20 m, and discharge was calculated using MAP provided in the MARN (2017) report. Rainfall values were corrected for evapotranspiration, using a map of vegetation from the MARN (2017) report, and evapotranspiration values from the Puerto Rico GAP project (Gould et al., 2008). It amounts to 10-82% of the total rainfall, for the different types of vegetation reported in the MARN report. EI values were normalized to the highest obtained EI value within the study area.

3.4. River profile segmentation

We extracted the long-profiles of 220 rivers located in the AC, SC, and SM ranges, using the Guatemala national 20 m-resolution DEM, released by the National Geographic Institute of Guatemala. The beds of these rivers were observed on stereoscopic couples of aerial photographs, provided by the National Geographic Institute of Guatemala (see next section for details). We sieve out rivers along which long-term dynamics is mired by adjustments to shorter-term disturbances, such as interactions with karst conduits, with debris flows, and with deep-seated landslides. A subset of 110 rivers that best capture long-term trends was used in the final analysis (Fig.7).

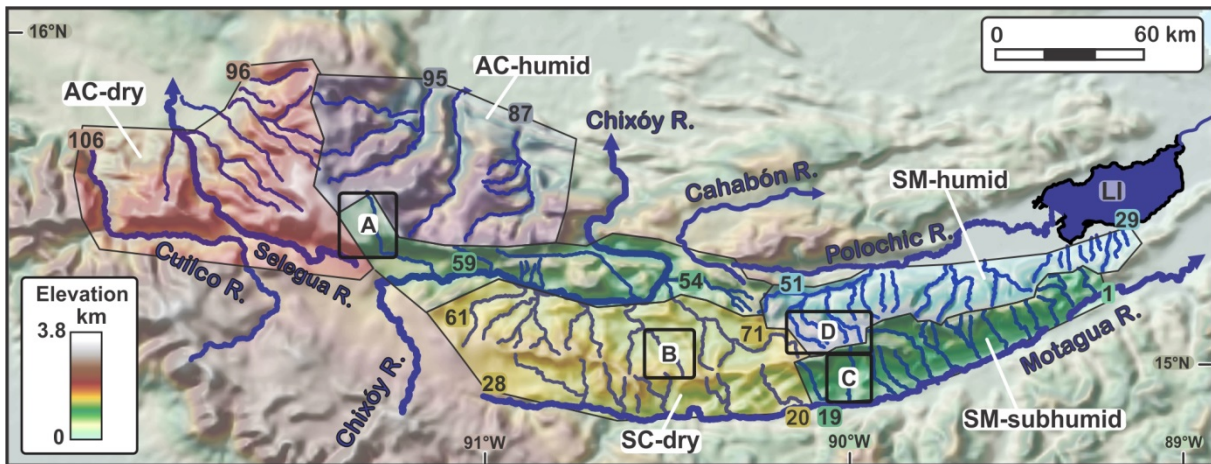


Figure 7. Distribution and grouping of the streams used in the river long-profile analysis, and their grouping by geographic areas. AC: Altos de Cuchumatanes, LI: Lake Izabal, PF: Polochic fault, SC: Sierra de Chuacús, SM: Sierra de las Minas. Numbers (1-106) correspond to the numbers ascribed to the rivers, as listed in table S4-1. The corresponding river profiles are presented on Figs S4-2 to S4-7, and on Fig. 12. Boxes A-D: footprints of the maps displayed on Fig. 8.

We resorted to a linearization method that filters out the downstream increase in stream discharge to identify knickpoints along river profiles. We chose the integral method (Perron and Royden, 2013), in which elevation is plotted (on chi-plots or χ -plots) as a function of chi (or χ), which is an upstream integral of incremental upstream distance, divided by a normalized local drainage area: Eq. (2):

$$\chi = \int_{x_0}^x \frac{A_0}{A(x')} \frac{m}{n} dx' \quad (2)$$

where A and x are the drainage area (in m^2) and upstream distance (in m), respectively; A_0 (in m^2) a reference drainage area and x_0 a reference upstream distance taken at the same point, m and n two exponents that encapsulate the influence of

drainage area and of local slope on river incision rate, respectively. This method divides river profiles into a succession of
285 linear to sublinear segments, separated by break-in-slopes that represent the knickpoints. The method overcomes much of the
scatter that plagues earlier linearization methods (Whipple and Tucker, 2002; Goldrick and Bishop, 1995). It requires,
however, a foreknowledge of the intrinsic concavity θ , defined as m/n . The value of θ can be determined by incrementally
fitting river profiles to a straight line (Mudd et al., 2014). Most of the studied rivers, however, have highly segmented
290 profiles (see section 4.3), preventing the retrieval of well-defined θ values. Besides, θ might vary along-stream as a result of
changes in climate (Murphy et al., 2016), or changes in the dominant erosive processes operating on the streambed (Brocard
and Van der Beek, 2006), such as the alternation of detachment-limited (Howard, 1994) and transport-limited river incision
(Whipple and Tucker, 2002), or such as the alternation of sediment-starved and overfed reaches (Sklar and Dietrich, 2006).
The method is here chiefly used to locate knickpoints. We therefore simply assessed whether the choice of θ would impact
the location and number of break in slopes, when varied within the range (0.4- 0.6) over which the profiles appear well
295 linearized. The segmentation being stable over this range (Supplement 3), a common normalizing concavity θ_n of 0.5 was
applied to all river profiles.

3.5. Classification of stream segments

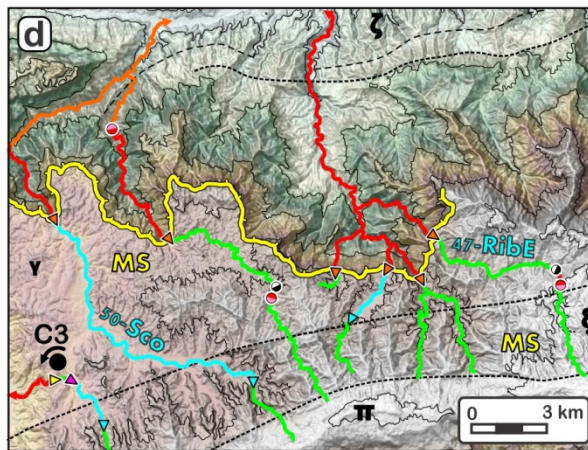
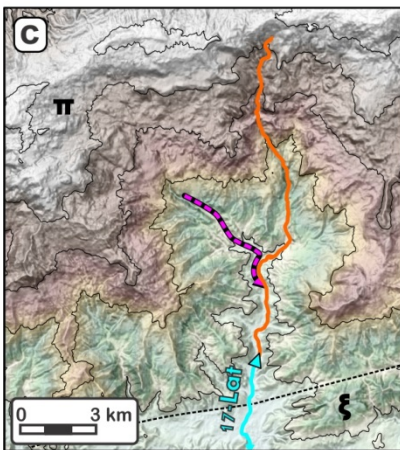
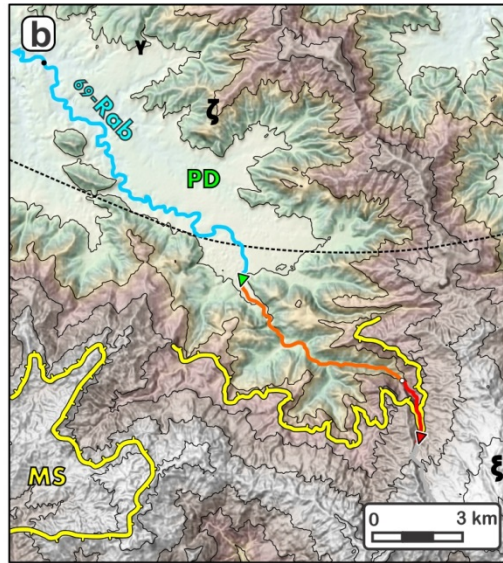
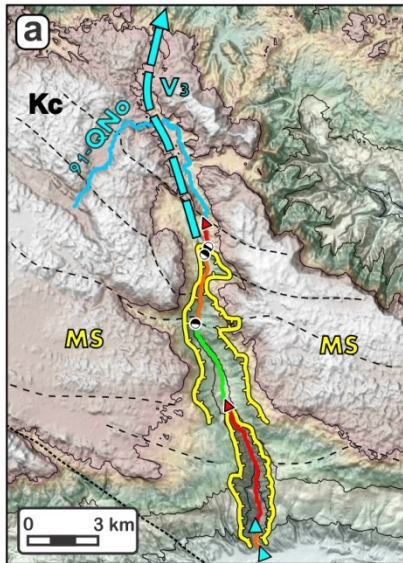
300 Streambed morphology was examined along each linearized segment using stereoscopic black-and-white 0.5-m resolution
aerial photographs shot in 2001, provided by the Guatemala National Institute of Geography (see Supplement 4).
Observations were occasionally ground-proofed during field work campaigns stretching over 6 years. River beds were
grouped into types according to the bed component that dominantly determines river gradient along each segment, namely:
bedrock, bedrock and gravel bars, gravel bars over bedrock strath, gravel bars over thick alluvial fill, colluvium, large
305 immobile boulders, boulders and gravel bars, boulders and bedrock (Table S4-1). Classification failed in many headwater
channels, where beds are masked by overhanging riparian vegetation. The classification roughly reflects differences in the
factors that determine stream incision. River incision is indeed likely detachment-limited along bedload-dominated reaches,
and transport-limited along gravel- and cobble-dominated reaches, lying over bedrock straths (Tucker and Whipple, 2002;
Brocard and Van der Beek, 2006). Boulder-armoured reaches are choked by slowly- to non-moving boulders that act more
310 like bedrock than bedload, and are therefore likely detachment-limited. However, unlike other types of bedrock channels,
boulder channels do not reflect the erodibility of the underlying bedrock, but rather that of surrounding hillslopes, because
the majority of the boulders originate from valley sides. Changes in streambed type from one segment to the next assisted the
classification of river knickpoints (see following section).

315

3.6. Classification of river knickpoints

Convex-up breaks-in-slope along river profiles are commonly referred to as river knickpoints. For convenience, we refer here to all breaks-in-slope, whether convex or concave, as knickpoints. Knickpoints were classified as lithogenic, alluvial, tectonic, migrating and miscellaneous (see supplement 4). Miscellaneous knickpoints represent adaptations of river profiles to local, stochastic disturbances (such as landslides and epigenies), and are usually short-lived. Most knickpoints in the study area reflect adaptations of river gradients to along-stream variations in rock uplift rate, bedrock erodibility, sediment flux, or sediment grain size. These knickpoints can be regarded as steady, inasmuch as their location only changes very slowly along the river profiles, tracking spatial changes in the distribution of rock types, rock uplift, sediment fluxes and bedload grain size. By contrast, knickpoints that spearhead step-increases or step-decreases in river incision rates migrate in the upstream direction along river profiles in the form of waves of accelerated (Rosenbloom and Anderson, 1994; Merritts et al., 1994) or decelerated (Howard, 1997) incision. They are hereafter referred to as migrating knickpoints, for they usually migrate faster than the knickpoints previously described. Concave-up migrating knickpoints commonly mark the transition from detachment-limited to transport-limited river incision (Whipple and Tucker, 2002) are usually found at the apex of alluvial fans (Fig. 8c) and pediments (Fig. 8b).

Theoretical geometric differences between migrating and steady knickpoints in linearized spaces have been used to discriminate unstable, migrating knickpoints from stable, equilibrium knickpoints (Goldrick and Bishop, 1995; Perron and Royden, 2013; Whipple and Tucker, 2002). In χ -space, upstream migrating knickpoints which celerity is controlled by the stream power law, and that propagate along various branches of a single drainage through a homogenous substrate affected by homogeneous rock uplift, should all share the same elevation and the same χ value (Royden and ~~Taylor~~ Perron, 2013). In the real world, however, variations in bedrock erodibility, climate, and rock uplift often scatter these values, challenging interpretations based on these sole geometric properties. This is especially the case in areas where environmental heterogeneities generate steady knickpoints which heights and wavelengths that are similar to that of the migrating knickpoints interspersed among them.



Knickpoint types:

- ◀ migrating ◀ migrating in soft substrate ● lithogenic ● structural
- ◀ pediment apex ◀ alluvial fan apex

Nature of stream bed:

- bedrock — boulder — gravel-covered strath — gravel fill — debris flow

Landforms:

- front of erosion
- paleovalley # V3

Bedrock structure:

- - - lithological unit boundary
- - - bedding/cleavage trace

Other:

- capture site
- MS Maya surface

Lithology:

- Kc** Cretaceous limestone **π** serpentinite **ζ** phyllite **ξ** gneiss **γ** granite

345 Figure 8. Examples of some knickpoint types presented in their geomorphic setting. Shaded and sloped 30 m resolution ALOS DEM ©
JAXA. Location of maps a-d on Fig. 7. (a) valley of the Quilén Novillo-Chancol river (91-QNo, AC Range), showing the paleovalley V3
(fig.2), shallowly incised into the Maya surface (Brocard et al., 2011), and two imbricate waves of erosion migrating up the reversed
(southward-directed) drainage of the valley. (b) Typical stepped topography of the SC range, in the catchment of the Rabinal river (69-
Rab), showing three imbricate upstream-migrating erosive signals distributed along the mountain slope. The upper one is a wave of
erosion that dissects the Maya surface, a second one is located half-way down the mountain flank, and the basal and final wave is located
350 at the base of the range, at the apex of pediments (PD). (c) Diffusive erosion in serpentinite mélanges, in the catchment of Río Hato (17-
Lat, SM Range). (d) dissection of the Maya surface by prominent migrating knickpoints along the northern flank of the SM Range, from
the Ribaco to the Chilasco Rivers (47-Rib to 50Sco). C3: 200 ky-old avulsion site (Brocard et al., 2012).

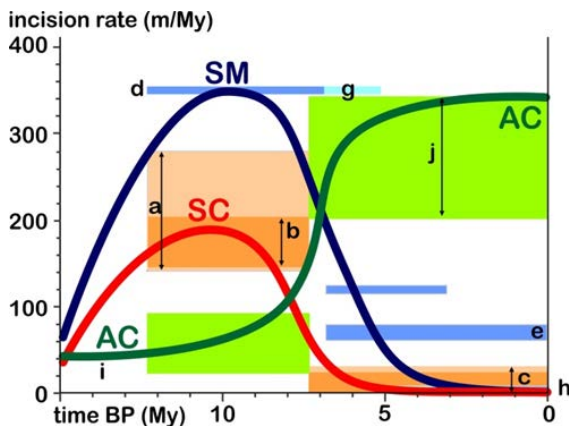
355 Additional discriminating elements must be used. A first screening consisted in checking whether the knickpoints coincide
with marked variations in bedrock erodibility, rock uplift rates, or local anomalies, in which case they were classified as
steady. To assess the effect of lithological variations we used 1:50,000 and 1:250,000 geological quadrangles of Guatemala,
and topical geologic maps from published papers (e.g. (Brocard et al., 2011; Bosc, 1971a). Stereoscopic black-and white 0.5-
m aerial photographs of the Guatemala National Institute of Geography were used to refine the location of lithological
360 contacts, and to assess the effects of bedrock fabric, fault damage zones, active faults, deep-seated landslides, and large
debris flows on the location of knickpoints. We used our foreknowledge of the tectonics of the area (Authemayou et al.,
2011a; Authemayou et al., 2012; Brocard et al., 2012) and of the Quaternary drainage reorganization (Brocard et al., 2012)
to assist the identification of tectonic knickpoints and of migrating knickpoints related to this recent reorganization. The
remaining knickpoints were then considered as potentially migrating, carrying the signal of the long-term evolution of the
365 studied mountains. We looked for supporting evidence of knickpoint migration, in particular break-in-slopes running along
valley flanks, tied to specific knickpoints (Fig. 8a,b) after verifying that such break-in-slopes were not produced by
lithological variations along valley sides. Changes in erosion rates along hillslopes associated to the passage of migrating
knickpoints can affect saprolite thickness (Brocard et al., 2015b) and associated drainage density. In only a few cases were
abandoned river terraces and hanging pediments found in the wake of migrating knickpoints.

370 The method has some limitations: first, local variations in bedrock erodibility maybe not be systematically detected, as a
result of the imprecision of geologic maps, especially in the least accessible parts of the SM and AC ranges. Second, large
intraformational changes in facies can generate variations in bedrock resistance as sharp as, or even sharper than erodibility
differences between mapped geological units. These two effects may lead to the interpretation of stable knickpoints as
migrating knickpoints. Conversely, some migrating knickpoints may be pinned to lithological contacts (Crosby and Whipple,
375 2006), and filtered out by the analysis. Nonetheless, we consider that, given the large number of analysed knickpoints, the
analysis captures the most import aspects of the evolution of the landscape within the study area.

4. Results

380 4.1. Rock uplift and stream incision chronology from $^{40}\text{Ar}/^{39}\text{Ar}$ dating

The Maya surface (Fig. 2,3) likely formed close to sea level, because it can be traced to the coast of the Caribbean Sea (Brocard et al., 2011). It was once covered by extensive fluvial deposits, especially south of the Motagua fault, where the fluvial deposits are preserved below extensive ignimbrites (Williams and McBirney, 1969). The lahar deposit of Chujuyúb rests directly onto a thick saprolite that blankets the Maya surface. Lahar emplacement predates the incision of a 385 450 m-deep valley. The lahar yielded a plateau age of 12.54 ± 0.04 Ma (Fig. S1-1, Table S1-2). It indicates that incision at Chujuyúb started after 12 Ma. This is consistent with the proposed 12 Ma entrenchment of the Cuilco River valley into the Maya surface (Fig.7), 70 km to the NW of Chujuyúb (Brocard et al., 2011), as well as with the 10.3 Ma emplacement of an ignimbrite within the Colotenango valley (Fig.2), 35 km to the NNW of Chujuyúb (Authemayou et al., 2012), which is 390 shallowly-incised into the Maya surface. It is also consistent with the 7.4 Ma deposition of an ignimbrite paleovalley, incised 1 km into the Maya surface (Fig.2), 10-30 km to the NE of Chujuyúb (Brocard et al., 2011). The depth reached by the late Miocene valleys prior to their abandonment implies that incision proceeded at $>140\text{-}280$ $\text{m}\cdot\text{My}^{-1}$ from 12 to 7 Ma, assuming that the dissection of the Maya surface started at 12 Ma (a, Fig. 9). Incision rates averaged over the length of the paleovalleys, between the SC and AC ranges, range from 145 to 205 $\text{m}\cdot\text{My}^{-1}$ (b, Fig. 9). Subsequent incision, from the base 395 of the late Miocene valley fills, down to modern valley floors, only amounts to a few tens of meters, at rates of <30 $\text{m}\cdot\text{My}^{-1}$ (c, Fig. 9).



400 Figure 9. Evolution of incision rates in the studied ranges.

Letters a-j correspond to river incision rates inferred from $^{40}\text{Ar}/^{39}\text{Ar}$ dating (see text). Black double arrowed lines: range of calculated value for each of incision rate. Rectangles: range of calculated values and time span of each of incision rate. Curves: overall evolution of incision rates within each range, inferred from these discrete estimates. Shades of blue: Sierra de las Minas (SM), red: Sierra de Chuacús (SC), and green: Altos de Cuchumatanes (AC).

405

The chronology of incision along the southern side of the SC range is documented by remnants of basalt flows scattered along the floor of the Motagua valley. These flows track from vents located south of the valley, on the Caribbean plate (Tobisch, 1986). The outcrop of El Jute represents the distal end of a lava flow which abutted the base of the SM range, backfilling the Huijo River valley with ≥ 70 m of basalt. The base of the flow lies >400 m above the Huijo River. Using the
410 modern gradient of the transport-limited Huijo River as a proxy for its 6 My-old gradient, we find that the basalt flow crossed the Motagua River 360 m above the current elevation of the Motagua River. The basalt yields a plateau age of 6.88 ± 0.03 Ma, and a slightly less constrained total age of 6.46 ± 0.09 Ma (Fig. S1-1, Table S1-2). Assuming that incision of the Maya surface started 12 My ago, then the 2.6 km-deep Motagua valley would have been incised at ~ 350 m \cdot My $^{-1}$ between 12 and 7 Ma (d, Fig. 9). Incision would have continued until today at an average rate 79 ± 4 m \cdot My $^{-1}$. If the basalt dam was
415 removed rapidly however, then incision would have instead proceeded more slowly, at 59 ± 9 m \cdot My $^{-1}$.

The chronology of incision can be refined by incorporating the previously dated basalts (Tobisch, 1986). The closest occurrence, located 6 km upstream along the Motagua River, is the 6.1 Ma Cerro lo de China flow. The flow was actually emplaced 120 km farther west at current plate-boundary slip rates, because it lies on the southern side of the Motagua fault. Conversely, the 3.1 ± 0.7 Ma Cerro Onanopa was emplaced on the same side of the plate boundary, 16 km
420 upstream of El Jute. Its high vesicularity implies an emplacement at, or near the ground surface, rather than as a sill, deep within the Subinal Fm., followed by exhumation. Its base lies <10 m above the Motagua River. Strath terraces of the Motagua River have been cut in its flanks (Tobisch, 1986), indicating that the flow underwent some minor burial and exhumation. The accordance in elevation between its basal contact and the Motagua River suggests that the Motagua River has oscillated tightly around its current vertical position over the past 3 My.

425 Incision of the Motagua valley, from the elevation of the basalt of El Jute, down to the current valley floor, would thus have occurred between 6.1 and 3.1 Ma, at $> 110 \pm 40$ m \cdot My $^{-1}$ (f, Fig. 9). If, after the emplacement of the basalt of El Jute, incision continued at the same ~ 350 m \cdot My $^{-1}$ rate as before (g, Fig. 9), then incision would have reached the current valley floor at ~ 5 Ma, no incision taking place afterwards (h, Fig. 9). The evolution of incision during the rise of the SC-SM range therefore looks similar on either side of the range: it is dominated by a single step of rapid incision, at 140-350 m \cdot My $^{-1}$
430 1, between 12 and 7-5 Ma, followed by an almost complete cessation of incision along the main trunk streams (the Motagua and Chixóy Rivers), which act as base levels of the streams located in the SC range.

Large steeply-dipping faults bound the Eocene fill of the Motagua valley. Dip-slip on these faults could be responsible, in part or in whole, for the deepening of the Motagua valley, a possibility contemplated by Tobisch (1986). Various traits of the valley, however, rule out any substantial contribution of these faults. First, fluvial sediments have
435 bypassed the Motagua valley since Eocene time, feeding a transtensional basin at the lowest eastern end of the Motagua valley (Fig.1b). Second, the alluvial fans that have grown astride these faults show no evidence of faulting, nor any anomaly in their catchment/fan surface ratios (Tobisch, 1986). Third, the faults encountered along the base of the SC-SM range exhibit only ancient, ductile to ductile-brittle left-lateral deformation (Bosc, 1971a; Roper, 1978). Last, the middle Miocene low-relief surfaces lie at about the same elevation north and south of the Motagua fault (Simon-Labric et al., 2013).
440 Extension on antithetic boundary faults would need to remain well-balanced, despite hundreds of kilometers of left-lateral displacement along the Motagua fault since the middle Miocene, to avoid the development of significant offsets of these surfaces. The deepening of the Motagua valley therefore appears to have been achieved by erosion, through the removal of the erodible Eocene sediments that filled the Eocene fault basin, giving the valley the appearance of a recently active graben.

445 The incision chronology of the AC range is constrained by transverse paleovalleys that are shallowly incised into the Maya surface (e.g. Fig. 8a; i, Fig. 9). Uplift of the AC range since their abandonment provided space for the incision of 1,500- 2,600 m deep valleys along the northern flank of the range (Fig. 3) at 200-350 m·My⁻¹ (j, Fig. 9). River incision of the AC range therefore started and developed while river incision in the SC-SM range was stalling.

450 **4.2. Spatial variations in ¹⁰Be-derived erosion rates**

Catchment-averaged detrital ¹⁰Be erosion rates range from 11 m·My⁻¹ within the catchments that drain the Maya surface on the SM range, up to 330 m·My⁻¹ along the wet and steep northern flank of the AC range (Fig. 10). Most slowly-eroding catchments are located within the SC range. Weighting erosion rates by the relative concentration of quartz in
455 quartz-feeding lithologies only marginally affects the calculated rates (by 3.4% on average in the SC range, 4.8% in the SM range, and < 7% in the AC range). Quartz enrichment corrections, on the other hand, increase erosion rates by up to 40% (Fig. 10, Table S2-2), but the amplitude of this effect remains speculative, in the absence of field measurements. However, because quartz enrichment increases with weathering intensity, its effects are probably smaller in the AC range, where soils erode the fastest. Quartz enrichment could therefore act such as to reduce contrasts in erosion rates between the SC-SM
460 range and the AC range.

In the AC range, erosion rates (arrows, Fig. 10) show a marked increase from the drier, and less steep highlands, to the wet and steep frontal slopes (from CATA to CHEL to XAC). The SM range displays a similar trend of increasing erosion down the mountain flank, as entrenchment in the Maya surface increases (from COL to FRI to RAN), with one outlier

(SLO). The magnitude of increase in the SM range intermediate between the ones measured in the SC and CA ranges. In the
 465 SC range indeed, a downstream increase would be expected to occur first, in the downstream direction, between the drainage
 divide and the mountain flanks, as a result of the decreasing contribution of slowly-eroding low-relief uplands with
 downstream distance (Willenbring et al., 2013b). Increase should be followed by a decrease in erosion rate as rivers start
 draining pediments that floor the Chixóy River catchment. An increase in erosion rate, downstream of the paleosurface, is
 measured (from PAS to PAE), but it is much less pronounced than in the AC range. The decrease in erosion rate is also very
 470 subdued (from XEU to CUB). In one case (from SMS to SMM to SMI), no increase nor decrease are observed.

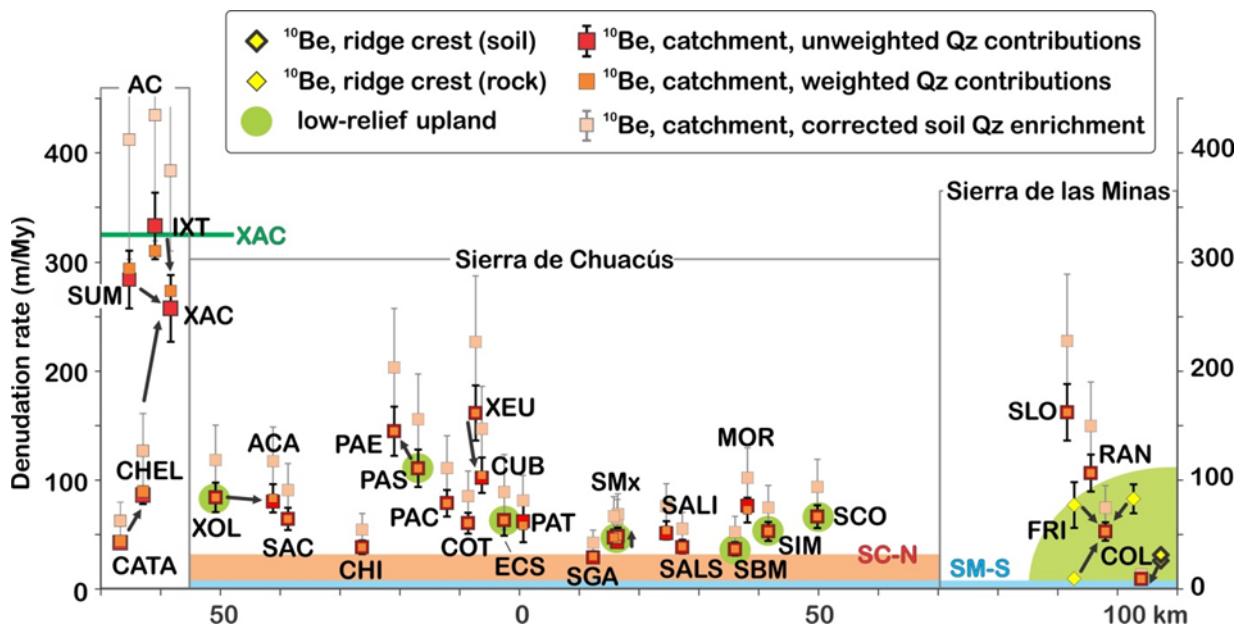
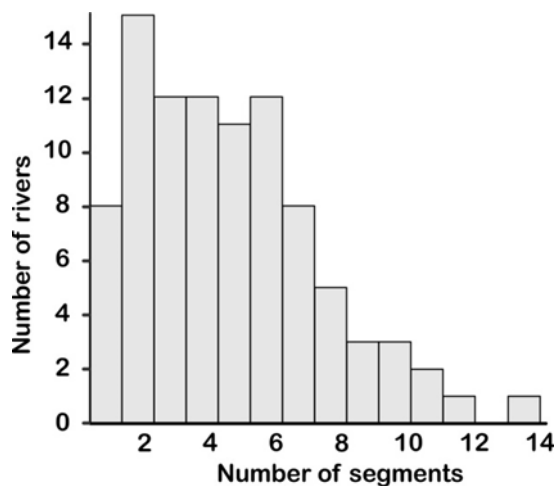


Figure 10. Variations in detrital ^{10}Be hillslope erosion rates along the strike of the plate boundary, from the AC range in the west, to the
 475 SM range in the east. Data are projected along the strike of the plate boundary, on an arc displayed on Fig. 6. Arrows show feeding
 relationship, between nested hillslope sites and catchments. Some uplands drain remnants of the Middle Miocene Maya surface (green
 shaded areas). Detrital ^{10}Be hillslope erosion rates are compared to river incision rates along the northern (salmon-colored shaded area,
 SC-N) and southern (blue-shaded area, SM-S) base of the SC-SM range, over the past 7 My. XAC: peak incision rates in the AC range
 along Río Xacbal, below the Maya surface (Fig. 3).

480

4.3. Distribution of streambed types

9% of the rivers retained in the analysis do not host any knickpoint, 16% host one knickpoint, 51% host 2 to 5 knickpoints, and 25% host 5-12 knickpoints (Fig. 11), reflecting the high degree of segmentation of many rivers, and high density of
 485 knickpoints. The knickpoints delimitate 452 river segments, of which 92% are well linearized, 6% are concave, and 2% are convex, for an applied intrinsic concavity $\theta = 0.5$. River segments were grouped according to streambed types. The distribution of streambed types according to elevation is projected along the strike of the studied ranges on Figure 12 and Figure S4-7, and their distribution across the strike of the ranges on Figures S4-2 to S4-6.



490 Figure 11. Distribution of rivers according to the number of segments identified in each river

The distribution of alluvial reaches is bimodal in the SC-SM range (Fig. 12a1-b2): alluvial reaches tend to be found either at the base of the mountains, or at high elevation, over the remnants of the Maya surface (e.g. Fig. 8 a,d). High-
 495 elevation alluvial reaches tend to transport a rather fine-grained bedload, composed of sand derived from the weathering of micaschist, gneiss and granite, and of gravel derived from quartzose veins and silicified pegmatites (Brocard et al., 2012). Intermediate-elevation alluvial reaches occur upstream of obstructions, most notably landslides in the SM-SC range, over extremely erodible fault damage zones, and within localized areas of tectonic subsidence (especially along the Polochic fault corridor, on the southern flank of the AC range).

500 Boulder reaches are found mostly on crystalline rocks. There, they are more frequent on the wet slopes SM range than on the dry slopes of the SC range. In the SM range, many boulder-strewn reaches form after the winnowing of the fine-grained matrix of debris flows. The SM range is the first range hit by tropical depressions tracking from the Caribbean Sea. They trigger numerous landslides along the wettest slopes of the SM range (Ramos Scharrón et al., 2012; Bucknam et al., 2001). Because SM range soils are more frequently close to water-saturation, they are more likely to be affected by

505 landslides when earthquakes strike the range (Harp et al., 1981). In the SM range, boulder armoring is common on the serpentinite mélanges that crop out up to high elevations along its southern flank, owing to the presence of knockers in the mélanges (e.g. Fig. 8c). Boulder-strewn reaches in the AC range form over phyllites. There, boulders are made of the most resistant beds of Pennsylvanian phyllites, and of sandstone and limestone blocks derived from overlying formations that slid along valley flanks down to the streambeds.

510 Bedrock river reaches are most commonly found downstream of convex migrating knickpoints, the distribution of which is presented in the following section.

4.5. Distribution of steady and migrating knickpoints

515 Among the 350 identified knickpoints, 40% can be tied to variations in bedrock erodibility, 6% to temporary obstructions, 8% to active tectonics, 21% to upstream-migrating waves of accelerated erosion, and 14% to upstream-migrating waves of decelerated incision. 11% are composite and result from some combination of the above.

Details about the significance of the distribution of steady knickpoints are provided in Suppl.4, as well as a more systematic review of the origin of all identified clusters of migrating knickpoints. Some migrating knickpoints can be tied to well-identified and well-dated river Quaternary diversions (e.g. S3-1 to S3-3, Fig. 12a2, (Brocard et al., 2012)). Most migrating knickpoints dot the rim of upland low-relief surface remnants (Fig. 12 a1 and a2; Fig. 8d). They may have therefore initiated when the Maya surface started being incised, at ~12 Ma, on the sides of the rising SM-SC range. Other clusters of migrating knickpoints are found halfway down mountain flanks. The most conspicuous of these clusters is restricted to the northern flank of the SC range, within the watershed of the Chixóy River. It hangs above a cluster of concave-up migrating knickpoints located at the apexes of pediments restricted to the drainage of the Chixóy River. The significance of these concave-up knickpoints and their genetic relationship with the clusters of convex knickpoints located upstream is discussed in section 5.3.2.

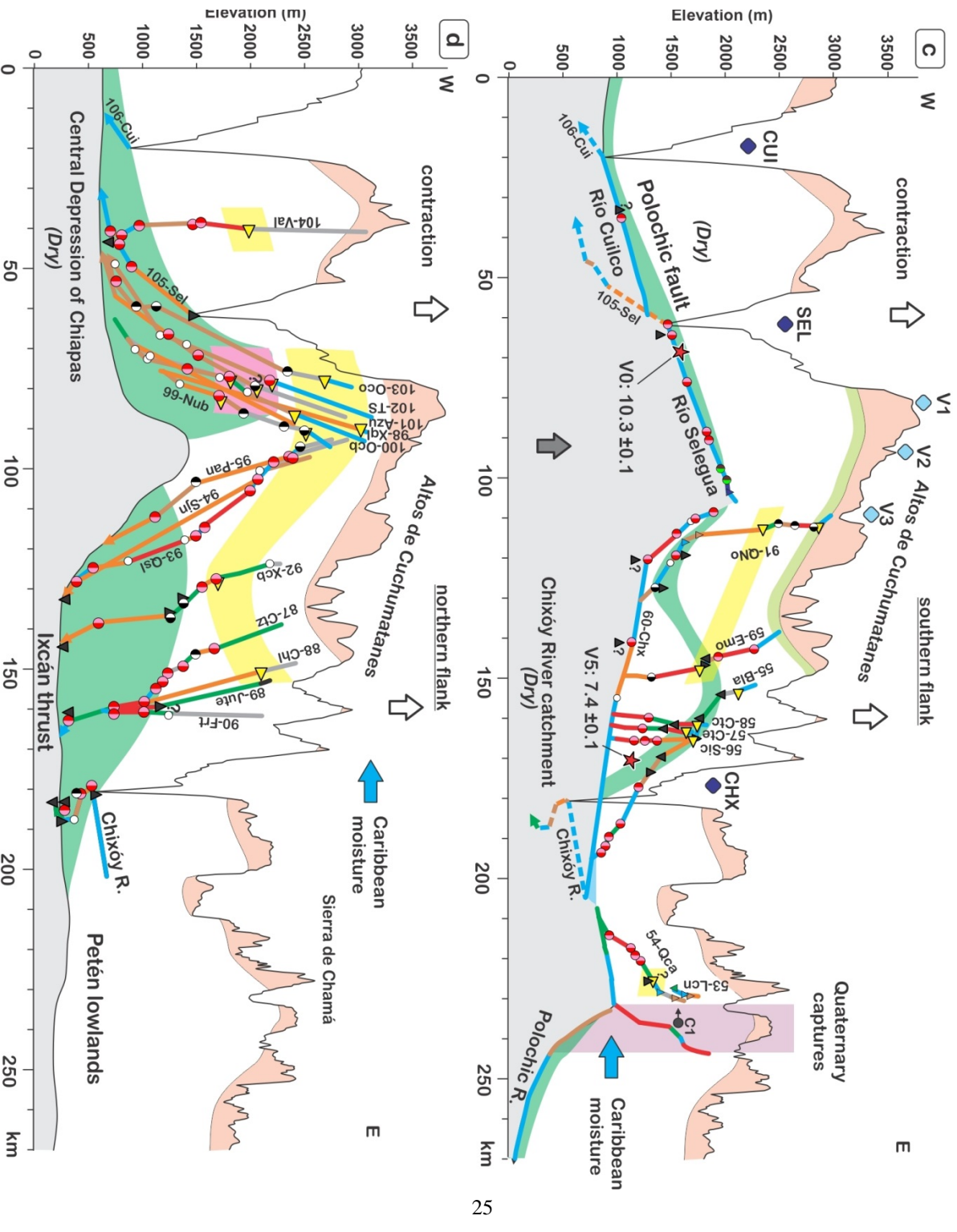


Figure 12.(c) and (d): distribution of linearized stream segments and knickpoints along the AC ranges.

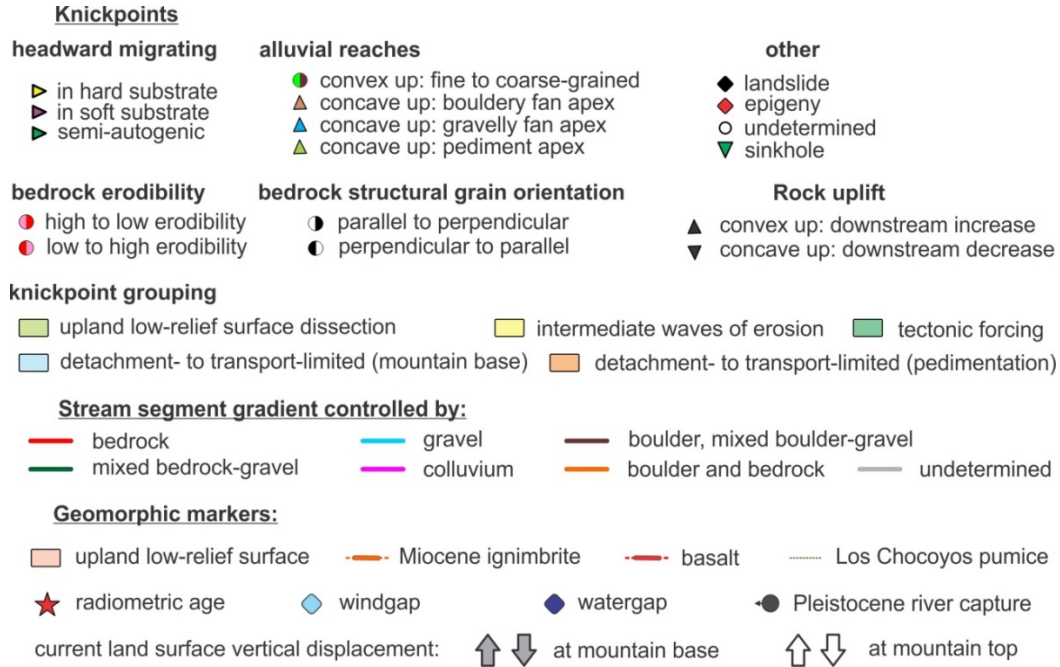


Figure 12. Distribution of linearized stream segments and knickpoints along the SC-SM and AC ranges. Mountain ranges are projected on the plate boundary, according to a small circle defined on Fig. 6. (a) and (b) southern and northern flanks of the SC-SM range, (c) and (d): southern and northern flanks of the AC range. Key to abbreviated stream names is provided in Table S4-1 and in the captions of figures S4-2 to S4-7. Water gap names: CUI: Cuilco, SEL: Selegua, CHX: Chixóy. Paleovalley numbering from Brocard et al. (2011), river capture numbering from Brocard et al. (2012). LA: city of Los Amates.

540

5. Discussion

The decline of river incision rates in the SC-SM range was coeval to the rise of in the AC range (Fig.9). This can reflect a complete transfer of rock uplift from the SC-SM range to the AC range, but it remains surprising that river incision rates declined so sharply, considering the range had not undergone any substantial topographic decay. Likewise, very low

current hillslope erosion rates are maintained on steep slopes within the SC-SM range today. A genetic relationship can therefore exist in the rise of the AC range and the decline of river incision and hillslope erosion in the SC-SMA range. The rise of the AC range may have affected incision rates in the SC-SM range in two ways. First, by decreasing moisture
550 delivery to the SC-SM range, it may have reduced hillslope erosion rates and the delivery of water and sediment to the streams, thereby decreasing river incision rates. Second, by forcing the drainage of the northern side of the SC range to adjust to rock uplift in the AC range, it promoted a decrease in river incision rates, upstream of the AC, among the rivers of the SC range that still cross the AC range. After reviewing the respective contributions of these top-down and bottom-up processes, we analyse how they combined to affect the present-day morphological evolution of the SC-SM range.

555

5.1. Effect of the rise of the AC range on climate-driven hillslope erosion

5.1.1. Climate and hillslope erosion in modern times

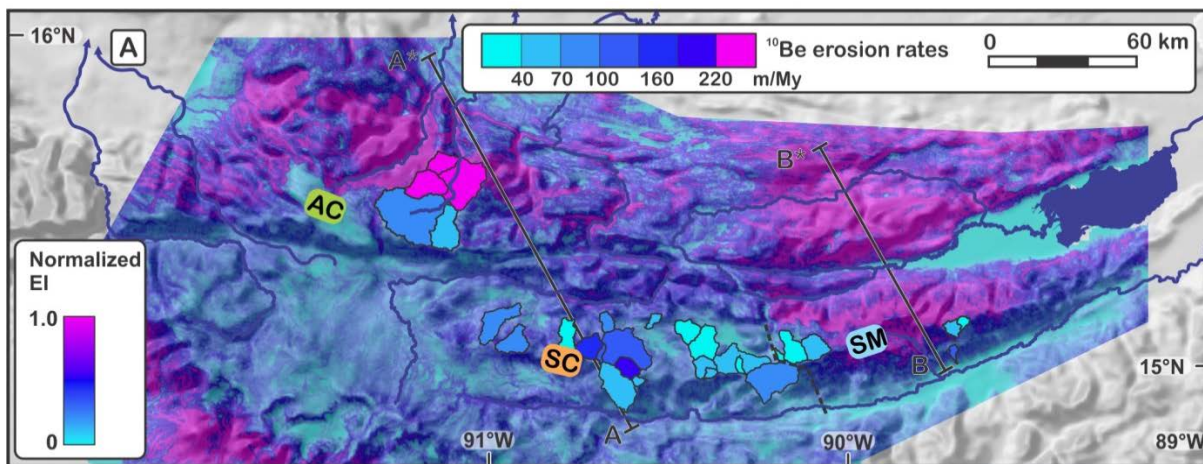
560 Silicate weathering is three times faster on the wet (1,800-3,000 mm·y⁻¹, Fig.4) side of the SM range than along its drier (1,000-2,400-mm·y⁻¹) side (McAdams et al., 2015). Using detrital cosmogenic ¹⁰Be we find that the wet (1,900-3,700 mm·y⁻¹) side of the the AC range erodes on average distinctively faster (with a 92 % probability, based on a Welch's t-test, with $t = 2.246$, $p = 0.08$) than the drier (900-1,300 mm·y⁻¹) SC range. Individual catchments document a sixfold increase in erosion (50 to 300 m·My⁻¹), from the SC range to the AC range (Fig.10, Fig. 13b,c). The spatial distribution of erosion rates
565 predicted by the Erosion Index (Fig.13a), which combines the effects of slope and precipitation on erosion, is consistent with the spatial distribution of measured detrital ¹⁰Be erosion rates. ¹⁰Be-feeding slopes in the AC range receive three times more precipitation (MAP = 3.2 ± 0.7 m·y⁻¹) than their counterparts in the SC and SM ranges (MAP= 0.9 ± 0.4 and 1.2 ± 0.3 m·y⁻¹, respectively, Fig.13b). Comparatively, ¹⁰Be-feeding slopes in the AC range are also marginally (1.2 times) steeper ($26 \pm 4^\circ$) than in the SC and SM ranges (22 ± 3 and $22 \pm 2^\circ$, respectively, Fig.13c). In the AC range, a linear relationship is observed
570 between erosion rate, precipitation ($r^2 = 0.83$), and slope ($r^2 = 0.95$). When all ranges are considered together, however, the correlation is weaker ($r^2 = 0.61$ for precipitation and 0.52 for slope), and very weak if the SC and SM range are considered individually ($-0.1 < r^2 < 0.4$). The strong correlation in the AC range stems foremost from the greater homogeneity in bedrock erodibility among the measured catchments. In the AC range, slope gradient and MAP are good predictors of hillslope erosion rates at precipitation > 2 m·y⁻¹ (Fig. 13b). However, this linear relationship predicts that erosion ceases for MAP < 2
575 m·y⁻¹ and slopes $< 15^\circ$. By contrast, if the entire dataset is considered, the linear regression predicts that erosion tends toward zero as MAP decline toward zero, while hillslope erosion ceases on slopes that are still steep ($\geq 19^\circ$, Fig.13c). Globally, erosion rates < 50 m·My⁻¹ are generally observed at slope values $< 10^\circ$ (Willenbring et al., 2013a). They are here measured in many catchments of the the SC and SM ranges that still maintain average slope values of 18-25°, pointing for a significant

580 contribution of aridification to the limitation of erosion. The overall decrease in the steepness in relationship between erosion and MAP curve from the AC range to the overall dataset may imply that the relationship is nonlinear, and that the curve flattens for $\text{MAP} < 2 \text{ m}\cdot\text{y}^{-1}$, a trend observed elsewhere (for example, at $\text{MAP} < 2 \text{ m}\cdot\text{y}^{-1}$ on Kauai (Ferrier et al., 2013)).

585 For $\text{MAP} < 2 \text{ m}\cdot\text{y}^{-1}$, MAP is a poor predictor of erosion rates, likely because storminess and rainfall intensity become better predictors of hillslope erosion (Liang et al., 2019). The correlation between slope and erosion rates in the AC range implies that slopes have not reached the critical threshold for slope stability (Clarke and Burbank, 2010) in the AC range. In the SC range, slope steepness only faintly captures variations in erosion rates ($r^2 = 0.41$) under low MAP: comparison between nested catchments indeed shows that, in 4 out of 6 instances, decreases and increases in EI values are not associated with significant changes in erosion rates.

It seems, therefore, that the most arid parts of these ranges erode slowly to very slowly, despite maintaining steep slopes. Besides, slope steepness and MAP become poor predictors of short-term (10^3 - 10^4 y) hillslope erosion rates at low MAP. The pattern of catchment-averaged ^{10}Be erosion rates therefore that aridification limits the erosion on steep slopes. Fluctuations of climate, over millions of years, can therefore be expected to impact the long-term evolution of these ranges. This aspect is explored in the following section, and in particular the interplay of mountain building and climate evolution in Central Guatemala.

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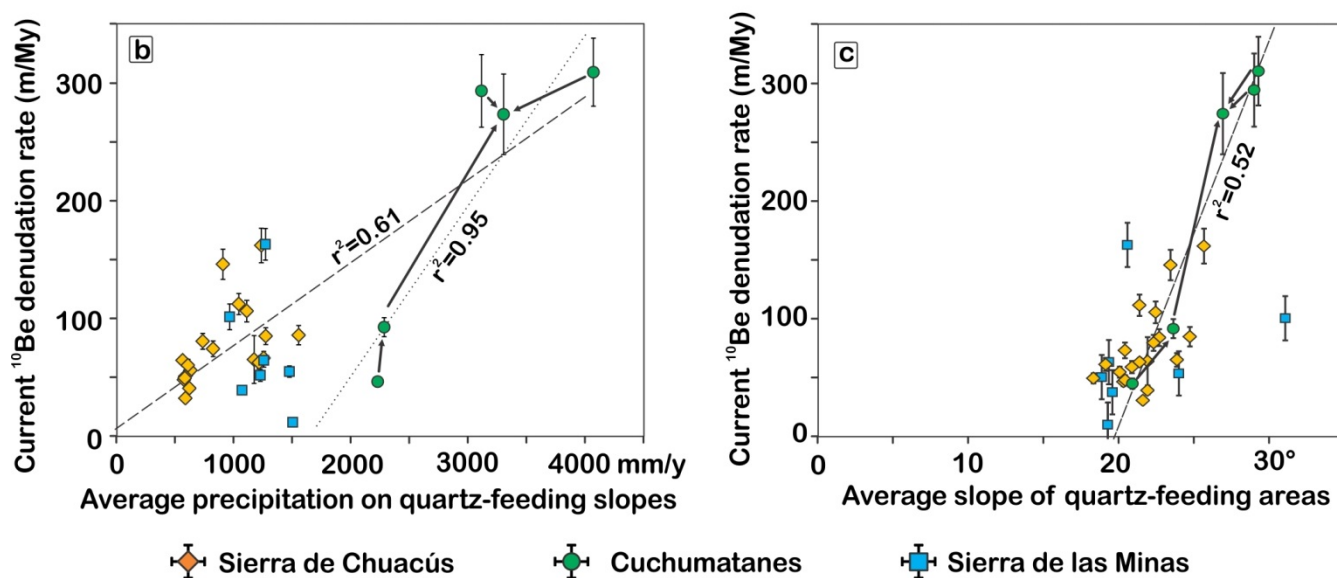


Figure 13. (a) Map showing the spatial distribution of normalized catchment-wide ^{10}Be erosion rates superposed to the predicted spatial distribution of erosion according to the normalized erosion index. (b) catchment-wide ^{10}Be erosion rates as a function of catchment-averaged precipitation. (c) catchment-wide ^{10}Be erosion rates as a function of catchment-averaged slope (from quartz-contributing slopes)

600 Arrows: downstream connections between nested catchment within the AC range. Dashed line: linear correlation within AC range data, with correlation coefficient (r^2) reported next to the line.

The sequential rise of the SC-SM and AC range generated an evolving pattern of hillslope steepness and of precipitation which may have contributed to the decline of incision in the SC range since the Middle Miocene. We review hereafter evidence for changes in climate and tectonics, susceptible to have impacted hillslope steepness and erosion over time.

5.1.2. Evolution of climate over the SC range since the rise of the AC range

610 The distribution of precipitation (Fig.2) shows that the AC range currently prevents the ingress of moisture tracking from the Yucatán and Petén lowlands toward the SC range. Moisture was therefore likely able to reach the SC range before the rise of the AC range. Such a deeper penetration of moisture is supported by paleo-precipitation estimates obtained from the analysis of tree species and of paleosols in the 7.4 Ma forest of Sicaché (Fig. 2). This subtropical forest was growing on the floor of a late Miocene paleovalley. The geochemical characteristics of its paleosols suggest mean annual precipitation

615 between 950 and 1,300 mm·yr⁻¹ (Brocard et al., 2011). Today, the fossil forest is located in a drier area covered by
xerophytic vegetation. The SC range therefore likely received higher precipitation between 12 and 7 Ma, and its aridification
was coeval to the rise of the AC range. Aridification likely led to the decrease in hillslope erosion rates, down to the values
that are measured today.

620 **5.2. Effect of the rise of the AC range on tectonically-driven river incision in the SC range**

The rise of the AC range sparked widespread reorganization of the range-transverse drainage (Brocard et al., 2011),
tectonically defeating many rivers that used to cross the AC range. Some rivers maintained a course transverse to the rising
structure by steepening their gradient. We review hereafter the evidence for faster rock uplift in the AC range than in the SC
625 range, for river course lengthening along the Polochic fault, and for decreased rock uplift rates in the SC range. We review
the contribution of each of these tectonic processes to the decline of river incision in the SC range.

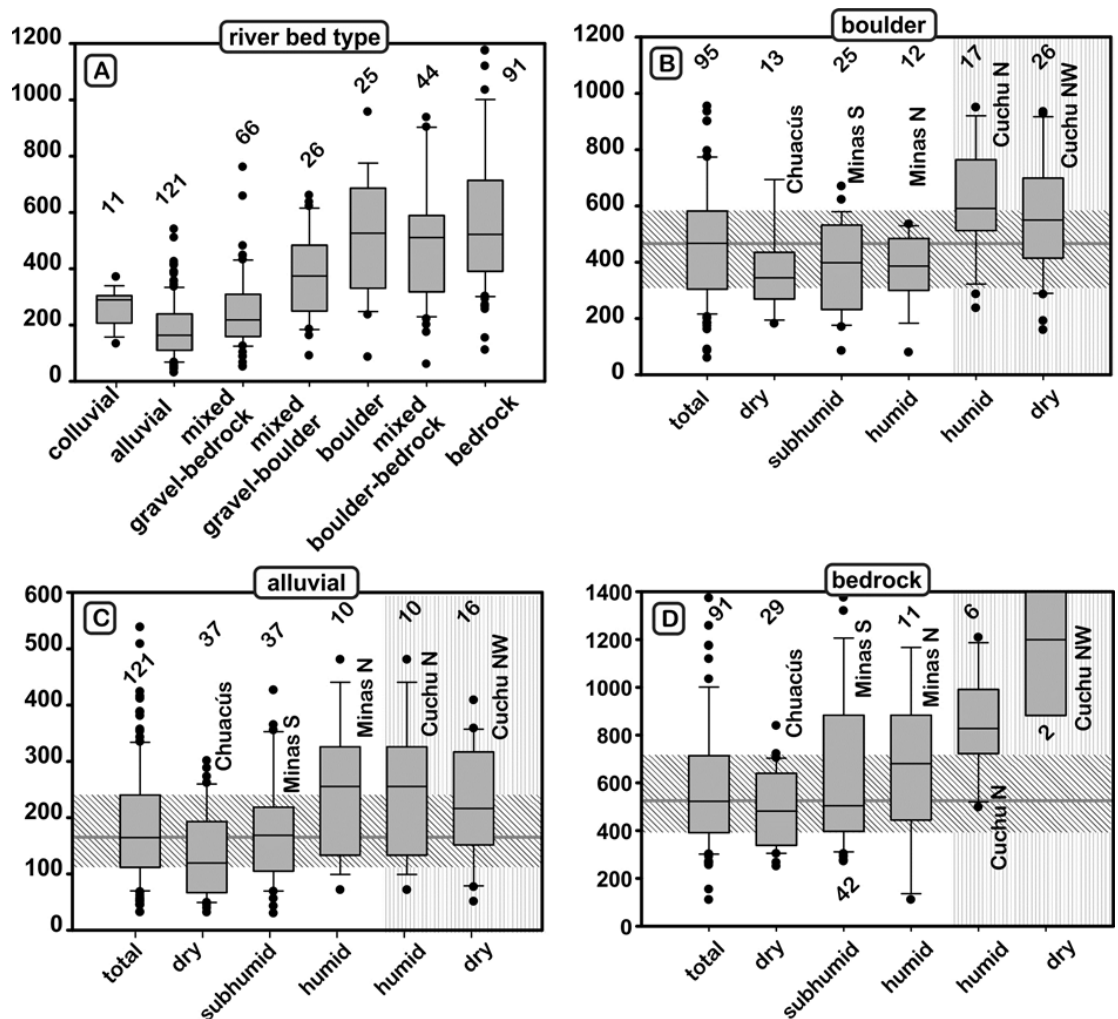
5.2.1. Fast, ongoing rise of the AC range from river channel steepness

630 The deformation of paleovalleys abandoned during the uplift of the AC range documents >1-2 km of rock uplift in
the AC range relative to the SC range since the late Miocene (Brocard et al., 2011), implying on average faster rock uplift in
the AC range than in the SC range over 7 My. Whether such difference continues today is important to establish before
analysing river profile dynamics. Glacial and fluvial landforms in the AC range suggest that the AC range still rises faster
than the SC range today. A 20 x 30 km ice cap was established on the summit plateau of the AC range (Fig. 1) during the last
635 glaciation (Anderson et al., 1973; Lachniet and Vazquez-Selem, 2005). On other neotropical mountains, last glaciation
moraine arcs are set in moraine arcs deposited by more extensive glaciers during earlier glaciations (Lachniet and Vazquez-
Selem, 2005). In the AC range, however, older moraine arcs are not observed. Their absence implies that earlier ice caps
were smaller than the one that occupied the plateau during the last glaciation. Their deposits were eroded away by the last
glaciation ice cap. There is no particular reason why climatic forcing would have affected Central Guatemala in a different
640 manner than other neotropical areas. The most straightforward explanation therefore is that the fraction of the range located
above the equilibrium line of accumulation (ELA) has increased steadily from one glacial cycle to the next, as a result of
sustained surface uplift. Uplift was fast enough to overcome differences in the intensity of glacial cycles, from one cycle to
the next. The effect of surface uplift on ELA was enhanced by the fact that the top of the range is a plateau, such that small
increments of uplift bring large areas of the range above the ELA. The current driver of surface uplift is still contraction
645 (Guzmán-Speziale, 2010; Authemayou et al., 2011b), just as it was during the early stages of mountain growth. Erosional

unloading driven by deep dissection along the northern flank of the AC range may now contribute substantially to surface uplift of the summit plateau.

The steepness of river profiles in the AC range further support the hypothesis that the range still rises faster than the SC range. The projection of river profiles in χ space (Figs. S4-2 to S4-7) shows that, in each of the analysed areas (AC, SC, and SM ranges), linearized segments with similar streambed conditions (alluvial, boulder-armoured, bedrock) tend to share similar θ_n -normalized steepness (Fig. 14a). Rivers that flow over bedrock are steeper than rivers that flow over immobile boulders, which in turn are steeper than rivers that flow over alternations of gravel bars, or bedrock and boulder, which in turn are steeper than alluvial rivers. Each category, however, exhibit steepness values that change from one range to the next (Fig. 14 b,c,d). It is expected that bedrock rivers conform to the predictions of the stream power law, because their incision is detachment-limited. At dynamic equilibrium, their steepness should therefore scale with rock uplift, bedrock erodibility and precipitation (Whipple and Tucker, 1999). The progressive increase in steepness from the SC range to the AC range does not result from an increase in bedrock erodibility, because erodibility is higher in the AC range than in the SC and SM ranges. It does not result either from the observed increase in precipitation from the SC to the AC range, as increased precipitation instead decreases river gradient. The steeper reaches of the AC range are therefore best explained by faster incision in the AC range, driven by faster rock uplift. This is consistent with the higher ^{10}Be hillslope erosion rates measured in the AC range, assuming that slopes and channels are well coupled (Callahan et al., 2019).

Boulder-armoured and alluvial channels are also steeper in the AC range (Fig. 14b,c). Boulders act as bedrock, and boulder-armoured channels can therefore be expected to behave like detachment-limited channels. Alluvial channels are likely transport-limited, and therefore their gradient is less sensitive to rock uplift (Whipple and Tucker, 2002; Cowie et al., 2008). The observed increase in alluvial channel gradient therefore most likely reflects an increase in bedload grain size with increasing erosion rate, resulting from shorter residence time and limited comminution of bedrock blocks in hillslope soils (Riebe et al., 2015; Neely and DiBiase, 2020).



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Figure 14. Box plots of stream segment normalized steepness as a function of streambed type and location.

(a) Comparison of steepness by streambed environment; (b), (c), and (d): spatial variations in channel steepness across the study area, according to regions defined on Fig.7, for the three main types of streambed environment: (b): boulder, (c): alluvial, (d): bedrock. Numbers above box plots: number of segments in each category. Oblique hatches: average over the total number of segments. Vertical hatches: rapidly uplifting AC range.

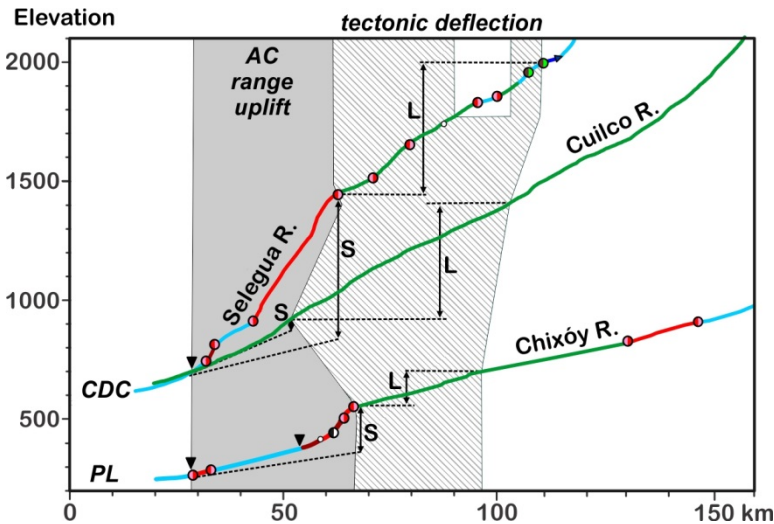
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5.2.2. Tectonic steepening of rivers transverse the AC range

The four rivers that still cross the AC range are the Cuilco, Selegua, Chixóy, and Cahabón Rivers (Fig. 7). Of these, the Cahabón River is the smallest. It is affected by an ongoing pulse of drainage rearrangement which over the past ~1 My has

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drastically reduced the size of the Cahabón River catchment south of the Polochic fault (Brocard et al., 2012). To analyse the impact of the AC range on transverse rivers, we therefore focus on the three other rivers. Two of these, the Chixóy and Selegua Rivers, exhibit steeper profiles as they cross the AC range (Fig. 15). Along these steepened reaches, the rivers incise rock formations that they incise farther upstream without displaying similar steepening. The steepening is therefore not caused by more resistant rocks but by faster uplift in the AC range (Leland et al., 1998; Kirby, 2003), consistent with the steepening of rivers in the AC range described in the previous section. The limited steepening of the Cuilco River could imply that, in the area crossed by the Cuilco River, the AC range does not rise faster than areas located farther upstream. Such an interpretation, however, is at odds with documented ongoing contractional deformation in the area (Authemayou et al., 2012). Instead, the lack of steepening is better explained by the fact that the Cuilco River is transport-limited over its entire length, as evidenced by a continuous cover of alluvium. The Selegua and Chixóy display alternations of bedrock and alluvial reaches typical of detachment-limited rivers. Transport-limited conditions tend to generate less steepening than detachment-limited conditions in response to enhanced rock uplift (Cowie et al., 2008; Whipple and Tucker, 2002). Transport-limited conditions along the Cuilco River result from the profuse delivery of volcanic gravel in its headwaters, where river drains the Central American volcanic arc. The Chixóy River also drains the volcanic arc, but volcanic gravel sources in the Chixóy River catchment are located farther away from the AC range, such that, by the time the Chixóy River reaches the AC range, most of this gravel has been trapped or comminuted and its bedload is dominated by proximal gravel sources (Deaton and Burkart, 1984).



700 Figure 15. Long profiles of rivers transverse to the AC range: Chixóy, Selegua, and Cuilco, with indications of reaches affected by vertical rock uplift (grey area), and those affected by horizontal lengthening along the Polochic fault (hatched area). S and L: contributions of steepening and lengthening to the uplift of river profiles, upstream of the AC range. River knickpoint nomenclature: see Fig. 12.

5.2.3. Transient slowing of incision in the SC range in response to river steepening in the AC range

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The steepening of river profiles across the AC range occurs in response to the rise of the AC range, in an area previously characterized by a foreland trough which flowed shallow-gradient rivers. The phase of steepening that converted these shallow-gradient rivers to steep-gradient transverse rivers should have ended when the gradient of the transverse rivers became steep enough for river incision to counterbalance rock uplift in the AC range. Steepening requires the headwaters to rise with respect to the foreland. Such rise requires a transient imbalance, upstream of the AC range, during which incision rates are lower than rock uplift rates. The sharp drop in incision rates in the SC range from 145-205 $\text{m}\cdot\text{My}^{-1}$ to $<30 \text{ m}\cdot\text{My}^{-1}$ after $\sim 7 \text{ Ma}$ (see section 4.1) could have been driven by the steepening of the Chixóy River in response to the rise of the AC range, because the Chixóy River is the outlet of most rivers draining the northern flank of the SC range.

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Using the river profile of the Chixóy River (Fig.15), it can be assessed that, $\sim 190 \text{ m}$ of surface uplift along the Chixóy River can be ascribed to river steepening, at $\sim 27 \text{ m}\cdot\text{My}^{-1}$. If, before 7 Ma , rock uplift rates in the SC range matched its incision rates of 145-205 $\text{m}\cdot\text{My}^{-1}$ (Fig.9), then, if such uplift rates were sustained after 7 Ma , 1.5 to 2 My would be necessary to lift up the Chixóy River to its current elevation, upstream of the AC range. Incision would resume afterwards. This estimate has limitations, in particular because the Chixóy and Selegua Rivers experienced large changes in drainage area upstream of the AC range during the 7 Ma reorganization event (Brocard et al., 2011), which affected their gradient across the AC range. The Chixóy River was initially smaller, and therefore probably steeper. Conversely, the Río Selegua lost some of its headwaters, and this loss may have contributed to its steepening. The calculation, however, suggest that the timescale of equilibration should be much shorter than the 7 My that have elapsed since the AC range started to rise. Ever increasing rock uplift rate in the AC range would promote continuous steepening, and therefore could prevent the return of incision, upstream of the AC range. Although plausible, such an evolution remains speculative, because available data lack the resolution necessary to test it. We identified, however, other processes that can delay the resumption of incision along the northern flank of the SC range.

5.2.4. Slowing of incision in the SC range in response to river lengthening along the Polochic fault

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River courses are progressively offset by left-lateral slip on the Polochic fault, leading to the development of ever-lengthening tectonic deflections above the fault. Maintaining gradients along these deflections sufficiently steep to allow the downstream transport of the bedload requires uplift of river channels upstream of the tectonic deflections, commensurate to the amount of river lengthening. Before the rise of the AC range, left-lateral deflections formed at the contact between the SC range and the northern foreland. Streambed uplift along the deflections happened in an area where streambeds were

735 shallowly incised, allowing frequent avulsions toward the foreland that annealed deflections (Sieh and Jahns, 1984) and
limited river lengthening. The rise of the AC range promoted deeper entrenchment of the tectonic deflections along the
Polochic fault, preventing the annealing of deflections and converting slip on the Polochic fault into permanent, cumulative
river lengthening. The courses of the Chixóy, Selegua, and Chilco Rivers have thus been lengthened by 25 to 40 km (Fig.7).
The westward increase in the length of these deflections, in a context of westward decrease in total slip on the Polochic fault
740 (Authemayou et al., 2011b) was interpreted as the result of an earlier entrenchment of the rivers in the west (Brocard et al.,
2011). River lengthening drove 500, 600 m, and 150 m of surface uplift along the Cuilco, Selegua, and Chixóy Rivers,
respectively (Fig.15). Because left-lateral slip on the Polochic fault has been fairly steady over the past 7 Ma, at 2.9 ± 0.4
 $\text{mm}\cdot\text{y}^{-1}$ along the Chixóy River deflection (Authemayou et al., 2012; Bartole et al., 2019), river lengthening and uplift,
likewise, must have steadily contributed to surface uplift in the upstream parts of the Cuilco, Selegua, and Chixóy River, at
745 $15 \pm 2 \text{ m}\cdot\text{My}^{-1}$ along the Chixóy River. In contrast to river steepening, which generated 1.5 to 2 My of transient uplift, river
lengthening has generated a constant uplift roughly proportional to slip rate on the Polochic fault.

5.2.5. Slowing of incision in the SC range in response to decreased in rock uplift rates in the SC range

750 With estimated surface uplift rates of $\sim 27 \text{ m}\cdot\text{My}^{-1}$ and $15 \pm 2 \text{ m}\cdot\text{My}^{-1}$, tectonic steepening and river lengthening are
the two bottom-up processes, driven by the rise of the AC range, that contributed to the decline in incision rates in the SC
range. They only represent a fraction of the measured decline, which went from 145-205 $\text{m}\cdot\text{My}^{-1}$ before 7 Ma, down to <30
 $\text{m}\cdot\text{My}^{-1}$ after ~ 7 Ma (Fig.9). The decrease may also result from a substantial decrease in rock uplift rate in the SC range.

A decrease in rock uplift rates in the SC range is documented by the difference in elevation between the
755 paleovalleys located in the AC range, such as V3 (Fig. 2, Fig.8a), which underwent 2.8 km of uplift since ~ 7 Ma (assuming
an initial foreland elevation of 0.3 km), and the Chixóy River, which underwent only 0.35 km of uplift in the meantime. If
the SC range had continued to rise after 7 Ma at the rate at which river incision was taking place in the SC range before 7 Ma
(145-205 $\text{m}\cdot\text{My}^{-1}$), then the resulting difference in elevation between V3 and the Chixóy River (0.9 km) would be only $1/3^{\text{rd}}$
of the actual difference. It would be even smaller than the 1 km of back-tilting measured along paleovalleys V4-V5, south of
760 the Polochic fault (Fig.2 in Brocard et al. (2011)). The observed difference therefore requires a substantial decrease in rock
uplift in the SC range after 7 Ma.

Between 50 and 300 m of fluvial and volcanoclastic aggradation occurred along the rivers that used to drain the
northern flank of the SC range prior to 7 Ma (Brocard et al., 2011). This phase of aggradation can be viewed as a transient
response (van der Beek et al., 2002; Attal et al., 2008) to the rise of AC range (Brocard et al., 2011). It suggests that the AC
765 range was already undergoing faster rock uplift than the SC range early during its growth. Rivers have since then incised
down to pre-aggradation levels. If river incision in the SC range is now in dynamic equilibrium, then the lack of incision

over the past 7 My implies that the SC range has not undergone any rock uplift relative to the foreland since 7 Ma. More precisely, rock uplift in the SC range would exactly match the 350 m of surface uplift necessary to accommodate river steepening in the AC range and river lengthening above the Polochic fault. We could not identify any obvious tectonic process whereby such a fine balance would be achieved.

5.3. Contribution of aridification to the decline of river incision in the SC range

The absence of incision in the SC range could also be viewed also as a lack of response to tectonic forcing, in a landscape subjected to aridification since 7 Ma. By decreasing water discharge, aridification would reduce stream power. Besides, aridification could alter the balance between water and sediment discharges such as to reduce sediment transport capacity and river incision (Beaumont et al., 1992). It could also dampen the delivery of tools to erode the bedrock (Sklar and Dietrich, 2006). A contribution of climate to the slowing down of incision is further supported by the evolution of the southern flank of the SC and SM ranges. These flanks, albeit exposed to a different tectonic forcing than the northern side of the SC range, underwent a similar decline in incision. The only possible contribution of bottom-up processes to the slowing down of incision along these flanks is the lengthening of the Motagua River at its downstream end, into the Caribbean Sea. Such lengthening would promote surface uplift farther upstream, in order to maintain the downstream dispersal of sediments. Some lengthening probably occurred during the emergence, at the downstream end of the river, of a transtensional basin (Fig. 1b) that has been filled with terrigenous sediments since the Mio-Pliocene (Carballo-Hernandez et al., 1988). This lengthening of >180 km involved <75 m of surface uplift (elevation of the river at the upstream end of the basin). This amount of uplift does not match the amount of surface uplift affecting the northern side of the range. It seems highly unlikely that independent tectonic phenomena such as the uplift of the AC range and the lengthening of river courses along the Polochic fault to the north, the decrease in rock uplift rates within the SC range, and the lengthening of the Motagua River to the west, could produce a similar decline of incision on either side of the SC range. The aridification of the range, which is well documented on either side (Machorro, 2014), appears therefore to significantly contribute to the slowing down of river incision on both sides.

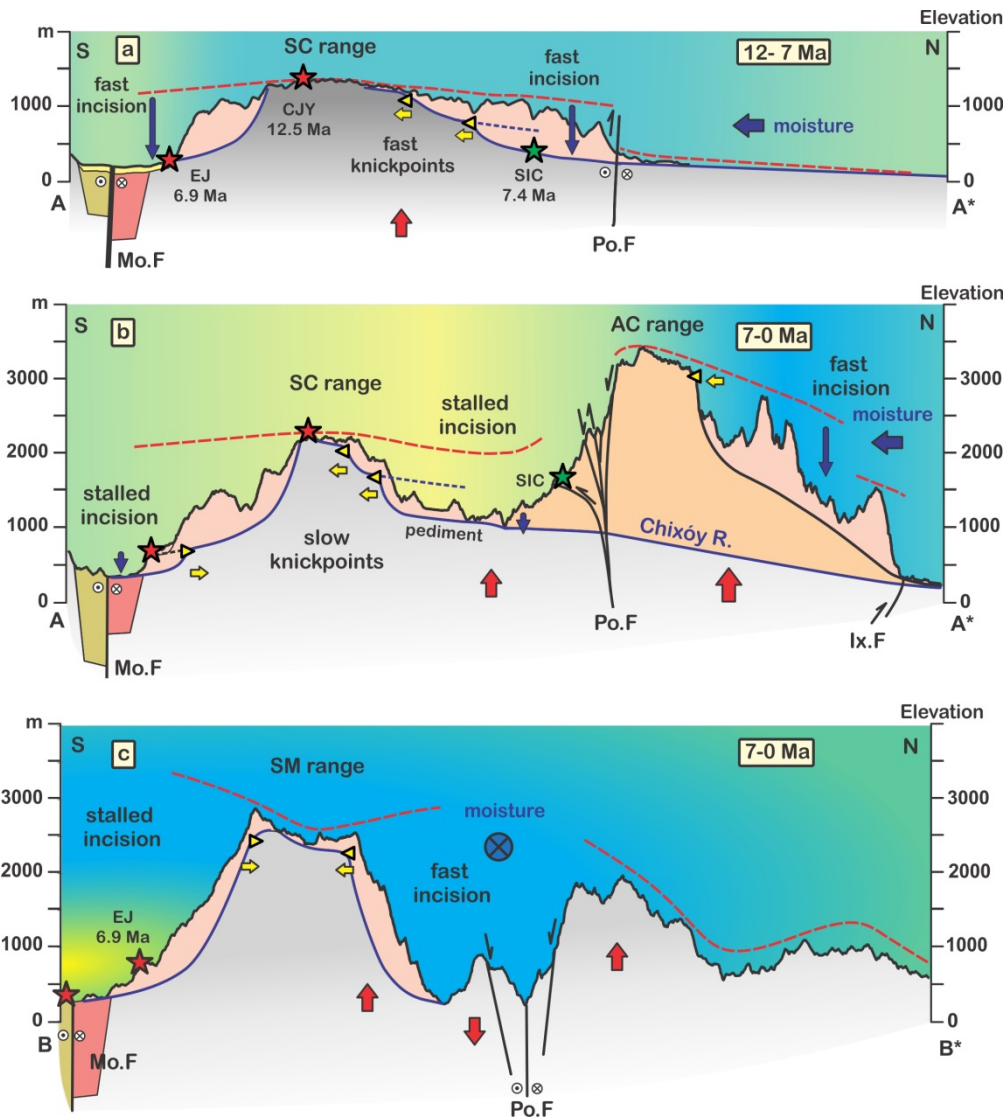


Figure 16. Development of the central ranges of Guatemala. Along profiles A-A* and B-B* (see Fig. 13a for location). Red arrows: rock uplift/subsidence (arrow thickness proportional to rate). Thick blue arrows: moisture fluxes. Thin blue arrows: river incision (length proportional to rate). Yellow triangles and arrows: migrating knickpoints with direction of migration. Atmosphere depicted as blue to yellow from wet to dry. Red stars: dated lavas, green star: fossil forest of Sicaché (SIC). Solid ground surface lines: black: topography along transects, blue: nearby river profiles. Dashed lines: blue: former river profile, red: Maya surface. Mo.F: Motagua fault, Po.F: Polo chic fault. Pink, brown, and yellow sediment fills next to the Mo.F: Eocene Subinal Fm., Middle Miocene Padre Miguel Fm., and younger alluvium.

5.3. Topographic evolution of the SC range in response to the uplift of the AC range

5.3.1. Slowing down of hillslope erosion, topographic decay, and backwearing

805 Modern ^{10}Be -derived hillslope erosion rates in the SC are higher ($> 50 \text{ m}\cdot\text{My}^{-1}$) than the $<30 \text{ m}\cdot\text{My}^{-1}$ long-term river incision rates measured around the SC range (Fig.10), suggesting that the relief of the SC range has been slowing decaying since 7 Ma. Hillslope degradation during the decay could account for the lower steepness of the quartz-feeding slopes in the SC range ($22\pm 3^\circ$) compared to those in the AC range ($26 \pm 4^\circ$). Hillslope degradation should ultimately lead to the decrease of the local relief of the SC range. The preservation of the Maya surface on many summits, and the absence of incision at the
810 base of the SC range however imply that the height of the SC range has not been significantly reduced over the past 7 My. Instead, decay appears to have proceeded by backwearing of the mountain flanks, from the base of the range inward, toward its divide. This is manifested by the development of pediments at the base of the range, and by the presence of upstream-migrating river knickpoints on the backwearing slopes.

815 5.3.2. Origin of migrating river knickpoints within the SC range

Three clusters of migrating knickpoints are found at different elevations along the northern flank of the SC range (Fig. 8b and 12a2). The uppermost cluster consists of convex knickpoints that dissect the middle Miocene Maya surface. They represent the front of an erosion wave that formed in response to the initial uplift of the SC range, at 12 Ma. These
820 knickpoints may have nucleated near the base of the range, along the Polochic fault, which, back then, represented the boundary between the SC range and the northern foreland (Fig. 16a). The second cluster of knickpoints consists of convex knickpoints located 500 m farther downslope, along the northern flank of the SC range (Fig. 8b and 12a2). These knickpoints have heights of 200-300 m (Fig. S4-3). If their height represents the amount of incision associated to their passage, then their passage must predate 7 Ma, because these rivers have only incised 20-90 m since 7 Ma between the knickpoints and the
825 Polochic fault (Brocard et al., 2011). Like the knickpoints of the upper cluster, the middle cluster knickpoints could have nucleated above the Polochic fault, but at a later time.

^{10}Be hillslope erosion rates suggest that today, these two generations of migrating knickpoints are almost immobile. Indeed, hillslope erosion rates are expected to increase in the downstream direction across actively migrating knickpoints. Yet, only marginal (e.g. catchments XEU to Cub) to negligible (e.g. SMS to SMM) increases in erosion rate are documented
830 across the knickpoints (see section 4.2). Knickpoint celerity is influenced by top-down processes, such as the amount of water runoff and sediment discharge delivered from upslope (Crosby and Whipple, 2006; Brocard et al., 2016b), and, under certain circumstances, by bottom-up processes, such as the rate of base level fall (Whittaker and Boulton, 2012). Therefore,

both the stability of the base level around the SC range over the past 7 My and the reduction in rainfall resulting from the rise of the AC range may have contributed, together with decreasing upstream drainage area (Crosby and Whipple, 2006), to the
835 slowing down of migrating knickpoints within the SC range (Fig. 16a,b).

The third cluster is located at the base of the mountain. It consists of concave-up knickpoints dotting the apex of pediments that have formed along the range (Fig. 8b and 12a2). The pediments are currently extensively buried under pumice, deposited during a large, late Pleistocene eruption (Brocard and Morán, 2014; Rose et al., 1987). Pediments usually
840 form under semi-arid climates, along drainages with stable base levels. They grow by extending their apex into the range (Pelletier, 2010; Strudley et al., 2006; Thomas, 1989). The pediments of the SC range have developed in an area that has been incising slowly over the past 7 My (<30 m-My⁻¹), and where climate is among the driest of Guatemala (Machorro, 2014). The concave-up knickpoints can be regarded as knickpoints spearheading a wave of decreased incision (Baldwin, 2003). Streambed analysis further shows that these knickpoints are located at the transition between detachment-limited and
845 transport-limited reaches downstream (Fig. S4-3), as predicted by theory (Whipple and Tucker, 2002). Lateral planation tends to dominate over vertical incision, downstream of that transition (Brocard and Van der Beek, 2006), which is consistent with pediment development.

The steepest reaches along the northern flank of the SC range are located between the intermediate cluster of convex-up knickpoints, and basal cluster of concave-up knickpoints. The close association of the middle and lower clusters
850 may not be coincidental. An alternate hypothesis for the formation of the intermediate cluster is that it did not nucleate, as hypothesized above, at the front of the range, but on site, by faster backwearing of the lower slopes than downwearing of the upper slopes, immediately ahead of the lower cluster. They would have since grown in height and steepness as the pediments extend into the range. The intermediate cluster and basal cluster can be viewed as the lips and toes of large knickzones (or knickpoint faces (Gardner, 1983)). Models predict that, if water discharge has a larger influence than river gradient on stream
855 incision, such knickpoint faces tend to steepen and amplify over time, during backwearing (Weissel and Seidl, 1998; Tucker and Whipple, 2002), which is consistent with the observed topography. The intermediate and basal clusters are not present farther east, along the wet flank of the SM range (Fig. 12d, 16c). Only one cluster of large migrating knickpoints separates the very flat uplands of the SM range from its deeply incised, wet lower flank (Fig. 8d). Aridification, therefore, appears to be responsible for the development of the highly stepped topography of the SC range.

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5.3.3. Range-hopping erosion and the development of dry orogen interiors

The evolution of the study area can be summarized as follows. Orogenesis started with the rise and coeval incision
865 of the SC-SM range, from 12 to 7 Ma (Fig.16a). After 7 Ma, rock uplift decreased in the SC range and picked up in the AC
range. Fast uplift in the AC range led to the tectonic defeat of many rivers which, upon exiting the SC range, flowed across
the foreland. Most of the defeated rivers were rerouted into the drainage of the Chixóy River, one the four rivers that
maintained a course across the AC range. Transient steepening of rivers in response to the fast rise of the AC range and the
lengthening of these same rivers by the Polochic fault promoted surface uplift within the SC range. The complete cessation
870 of river incision along the northern side of the range may therefore result from a combination of river steepening, river
lengthening, and decrease in rock uplift in the SC range. However, the fact the stalling of incision is as complete along the
southern side of the range as along the northern side suggests that tectonics was not the only cause, but that the aridification
of the SC range, resulting from the rise of the AC range, was also instrumental in the stalling of river incision. The decrease
in precipitation over the SC range reduced erosion on its hillslopes, decreased river discharge, and therefore contributed to
875 the stalling of river incision on both sides of the range. Precipitation and erosion concentrated on the northern flanks of the
AC and SM ranges, while the SC range became almost passively uplifted.

The high erosion rates and wet slopes that characterize the AC range today are reminiscent of the SC range between
12 and 7 Ma. The AC range became the front range, intercepting the moisture that tracks from the foreland, while the SC
became an inner range, upstream of the front range, and within its rain shadow. A decrease in rock uplift rate alone cannot
880 account for the sharp decline of erosion in the SC range after 7 Ma, because the relief and steepness of the range did
suddenly decrease. Aridification, within the rain shadow of the AC range, contributed to the overall decline in erosion in
the SC range. The SC range displays characteristics of dry orogen interior ranges and of orogenic plateaus: it has entered a
stage of progressive topographic decay, marked by the development of pediments. Pediment formation occurs in a context of
aridity and very low incision rates. Such low incision rates, driven by aridification in the rain shadow of front ranges, and by
885 river steepening across the front ranges, characterize the evolution orogenic plateaus formed by tectonic accretion (Sobel et
al., 2003; Garcia-Castellanos, 2007). On orogenic plateaus, this evolution ultimately leads to the disintegration of river
drainages, isolating dry interior drainages from the forelands. Along interior drainages incision then further decreases, the
rivers being graded to high-elevation base levels. Reduction in local relief then receives the added contribution of the
sediment infilling of the now closed catchments (Sobel et al., 2003). If the drying of the SC range was more severe it could
890 ultimately lead to drainage disintegration.

Continued growth of the pediment could lead to the formation of an intramontane pediplain (Baulig, 1957), at
elevations of 0.9-1.2 km, halfway between the high-standing remnants of the middle Miocene Maya surface, and the Petén-
Yucatán lowlands. The Central American Dry Corridor, which straddle the SC range, continues to the NW over the Central
Depression of Chiapas (Fig.1a), where extensive pediments have developed over the basement rocks of the Sierra Madre de
895 Chiapas (Authemayou et al., 2011a), in an area isolated from the Pacific moisture by the Sierra Madre de Chiapas to the SE,

and from Caribbean moisture by the Sierra de Chiapas in the NW (Fig.1a). These pediments are in a more advanced stage of coalescence and pediplanation. Pediplanation in a context of active orogenesis may explain the great abundance of low-relief perched surfaces that appear to form quickly in many orogens without significant pauses in mountain building (Calvet et al., 2015; Pain and Ollier, 1995; Babault et al., 2005).

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6. Conclusions

- Radiometric ^{39}Ar - ^{40}Ar dating on volcanic rocks confirms our earlier finding that the mountain range the closest to the plate boundary (the SM-SC range) was incised during the late Miocene, from 12 to 7 Ma. Incision almost completely stalled afterwards, during the formation of the Altos de Cuchumatanes (AC range) farther north.

- The deformation of paleovalleys indicates that the AC range experienced > 1,000 m of uplift relative to the SC range over the past 7 My. Today the range is highly dissected. High river profile steepness and ever-expanding ice-caps indicate that the AC range still undergo fast rock and surface uplift.

910 - The concentration of detrital terrestrial ^{10}Be in the sediments of rivers that drain these ranges show that hillslope erosion reaches $300 \text{ m}\cdot\text{My}^{-1}$ in the AC range, but is commonly $<100 \text{ m}\cdot\text{My}^{-1}$ in the SC-SM range. The patterns of hillslope erosion rates mimic spatial patterns of precipitation and stream incision.

- Precipitation is strongly controlled by topographic obstructions resulting from the rise of the AC range, which intercepts Caribbean moisture. Precipitation is high along the northern flanks of the AC range, but low over the SC range, which lies within rain shadows. Fossil vegetation preserved at the base of the SC range indicates a wetter climate at 7 Ma, when the AC range started to grow.

915 - In this context, the slow current hillslope erosion rates in the SC range appear to be contributed in part by the rise of the AC and by the development of rain shadows. It can be hypothesized, therefore, that hillslope erosion rates in the SC range were higher before the uplift of the AC range.

920 - The rise of the AC range led to the steepening of river profiles along the rivers that maintained a course across the rising range. The steepening of river profiles triggered a transient decrease in river incision rates upstream of the AC range. The lack of resumption of river incision upstream of the AC range implies either that these rivers have not re-established equilibrium profiles, or that other factors prevent the return of incision.

925 - The rise of the AC range led to the entrenchment of range-transverse rivers along the Polochic strike-slip fault. Fault slip has driven continuous lengthening of river courses on top of the fault over the past 7 My. Lengthening contributes to the rise of river profiles upstream of the Polochic fault, and therefore to the slowing down of river incision in the SC range.

- The difference in elevation between paleovalley in the AC range and the SC range imply rock uplift rates have declined in the SC range. This decrease in rock uplift rates may have also contributed to the decline in incision in the SC range

-The fact that incision decreased equally on either sides of the SC range, despite widely different tectonic forcing, implies however that aridification of the SC range contributed significantly to the decrease of river incision over the SC range.

- In the SC range hillslope erosion slightly outpaces base level lowering, implying an overall trend to slow topographic decay, despite continuing surface uplift. The persistence of Middle Miocene low-relief surfaces on mountain tops in the SC range however implies no net reduction in the range height, and a decay that proceeds by backwearing rather than downwearing.

- The slowing down of erosion rates over the SC range has resulted in a slowing down and stacking of upstream-migrating knickpoints and erosion waves over the SC range and of the development of pediments at its base.

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Author contribution

Gilles Brocard: project design, river ^{10}Be sampling (AC and SM ranges), ^{10}Be sample processing (AC and SM ranges), river segment analysis, manuscript preparation, with contributions from all authors. Jane Willenbring: ^{10}Be sample processing (SM-SC range). Tristan Salles: river profile segmentation. Mickael Cosca: $^{40}\text{Ar}/^{39}\text{Ar}$ dating. Axel Gutiérrez-Orrego and Noé Cacao-Chiquín: ^{10}Be stream sampling (SM-SC range). Sergio Morán-Ical: field work coordination. Christian Teyssier: project design and coordination.

950 Competing interests : the authors declare that they have no conflict of interest.

Code/Data availability : all in supplements

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