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Active hazardous slow faults drive drainage evolution in the Atacama Desert

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ABSTRACT

The Atacama Desert of the Central Andean forearc is the driest non-polar region on Earth, where extremely low erosion rates preserve geomorphic markers over Myr timescales. Between 19°S and 20°S, subtle Quaternary east-vergent reverse faults deform the west-sloping landscape and disrupt drainage development. This setting provides a unique opportunity to investigate fault evolution, uplift thresholds for drainage reorganization, and the seismic significance of slow faulting in hyperarid landscapes. We investigated two faulted interfluvies, Pampa de Tiliviche (~19.5°S) and Pampa de Soga (~19.75°S), using geomorphic mapping and cosmogenic ¹⁰Be dating of displaced alluvial surfaces, channels, and terraces. We constrained fault geometry, reconstructed fault-driven incision histories, and quantified uplift and slip rates. Faulting in interfluvies started after ~2.6 and ~5 Ma, while warped terraces and riverbeds record accelerated faulting since ~330

and ~35 ka, respectively. Simplified fault geometries yield maximum long- and short-term uplift/slip rates of ~0.011 and ~0.3 mm/yr. Numerical landscape-evolution models incorporating measured rates reproduce observed geomorphic patterns and constrain the uplift rate-catchment area ratio governing drainage reorganization. Observations and modelling indicate that tectonic uplift outpaces fluvial incision in hyperarid landscapes when drainage networks with small catchment areas cannot restore channel-bed slope modified by slow uplift. Regional consistency in fault geometry, distribution, and associated microseismicity suggests these counter-slope faults form an active forearc system capable of generating M6.1–6.8 earthquakes. Our results show that extremely slow faulting can substantially reorganize drainages in hyperarid landscapes and that Quaternary faults commonly considered inactive may represent an underestimated seismic hazard.

1 INTRODUCTION

Tectonic forcing of drainage networks by upper-plate faulting is well documented, but general erosion-to-uplift thresholds for drainage deviation, blocking, and beheading, as well as changes in faulting rates over geological timescales, remain difficult to constrain (García & Hérail, 2005; Goldsworthy & Jackson, 2000; Lettis et al., 1997; Medina-Cascales et al., 2021). This is largely due to variable fluvial erosion and deposition, and the degradation of geomorphic markers over tens to hundreds of kyr in most climatic settings (Hanks et al., 1984; Arrowsmith et al., 1998; Marliyani et al., 2016; Sare et al., 2019). Hyperarid landscapes provide an exception, as low rates of discharge make drainage networks particularly sensitive to subtle tectonic forcing and preserves landforms over much longer timescales (Kirk-Lawlor et al., 2013; Ritter et al., 2018b; Binnie et al., 2020).

The Atacama Desert is the driest non-polar region on Earth, lying along the highly active western South America convergent margin and evolving under hyperarid conditions since the

63 Middle Miocene (Dunai et al., 2005; Placzek et al., 2014; Jordan et al., 2014), or earlier (Ritter-
64 Prinz et al., 2026)(Fig. 1a). Its landscape comprises a Late Cenozoic west-sloping alluvial plain
65 across the Central Andean forearc, deformed by faults of different scales and dissected by west-
66 flowing trunk rivers sourced from the wetter volcanic arc (Fig. 1b). This dissection has isolated
67 interfluves with small catchments restricted to the Andean foothills within the hyperarid core of
68 the Atacama (Fig. 1b-c). These interfluves are largely isolated from any major Quaternary,
69 climate-driven changes in Andean runoff (Bobst et al., 2001; de Porras et al., 2017), resulting in
70 exceptionally low erosion and incision rates and the preservation of geomorphic markers over Myr
71 timescales (Dunai et al., 2005; Placzek et al., 2014).

72 Interfluvial drainage networks are locally disrupted by subtle counter-slope scarps
73 associated with east-vergent Plio-Quaternary reverse faults (Vega-Ruiz et al., 2025b). Despite their
74 potential to reorganize drainage networks, the Quaternary uplift rates and geomorphic effects of
75 these faults have been investigated in detail only locally (Binnie et al., 2020). Furthermore, their
76 broader tectonic and seismic significance remains poorly understood, although comparable subtle
77 faults in active margins are increasingly recognized as potentially underestimated seismic hazards
78 (e.g., Bullard & Lettis, 1993; Ishiyama et al., 2013; Vannoli et al., 2015).

79 Here, we combine detailed geomorphic mapping and cosmogenic ^{10}Be exposure dating of
80 geomorphic markers to quantify uplift and slip rates of counter-slope faults. We integrate these
81 results with numerical landscape-evolution modelling to investigate how upper-plate faulting
82 reorganizes drainage networks. We use this exceptional hyperarid setting to constrain erosion-to-
83 uplift thresholds for tectonic forcing of drainage and available seismic data to assess the potential
84 seismic significance of subtle faults in the Andean forearc.

2 REGIONAL, CLIMATIC AND GEOMORPHOLOGIC SETTING

2.1 Regional Setting of the Central Andean Forearc

The hyperarid Atacama Desert extends along the Central Andean forearc from southern Peru to northern Chile (16.5°–24.5°S) where modern rainfall is <30 mm/yr and vegetation is nearly absent (Latorre et al., 2002). Here, the Nazca Plate converges with the South American Plate at ~67 mm/yr toward N78°W (Norabuena et al., 1998), and forearc morphology is strongly influenced by subduction-related processes (Melnick, 2016; Bello-González et al., 2018) (Fig. 1a-b). From west to east, the forearc comprises four morphostructural domains: (1) the Coastal Range, formed by Mesozoic rocks (e.g., Scheuber & Reutter, 1992), (2) the Central Depression, a continental basin filled with Cenozoic arc-derived sediments (e.g., Mortimer et al., 1974), (3) the Precordillera, forming the mountain front of the volcanic arc; and (4) the Western Cordillera, the Pliocene–Recent volcanic arc (inset in Fig. 1b). Mean elevations increase from ~1100 m across the Coastal Range–Central Depression to ~3000 m in the Precordillera and >4000 m in the Western Cordillera. Most uplift of the Western Cordillera–Altiplano relative to the Central Depression occurred between the Late Oligocene and Late Miocene through west-vergent reverse faulting along the Precordillera (e.g., Farías et al., 2005; Pinto et al., 1999; Victor et al., 2004). Since ~10 Ma, a further ~1 km of surface uplift has been attributed primarily to mid-crustal processes, with limited contribution from faulting (Armijo et al., 2015; Hoke et al., 2007; Isacks, 1988).

2.2 Late Cenozoic Climatic Change and Fluvial Incision in the Atacama Desert

Since ~14 Ma, persistent hyperaridity has limited erosion of syn-orogenic deposits across the forearc during Andean uplift (Alpers and Brimhall, 1988; Sillitoe and McKee, 1996; Jordan et al., 2014). Although climate fluctuated to wetter phases during the Late Miocene–Pleistocene, conditions remained predominantly dry, with hyperaridity re-established over at least the last ~2 Myr (Zachos et al., 2001; Dunai et al., 2005; Rech et al., 2006; Amundson et al., 2012; Jungers et al., 2013; Wang et al., 2015; Evenstar et al., 2017; Ritter et al., 2018a). Late Quaternary wetter phases were spatially and temporally variable across the Atacama (Bobst et al., 2001; de Porras et al., 2017; Ritter et al., 2019; Rech et al., 2019; Wennrich et al., 2024). Increased Andean monsoon precipitation intensified Altiplano-fed drainage and generated ephemeral downstream lakes, whereas the Coastal Range and upper interfluvies remained drier (Dunai et al., 2005; Quade et al., 2008; de Porras et al., 2017; Ritter et al., 2019). Consequently, Quaternary erosion and deposition rates remained very low (<1.5 m/Myr; Placzek et al., 2014) (Fig. 1b-c).

This climatic and geomorphic stability favored preservation of the smooth, west-sloping Pacific Paleosurface, a Late Cenozoic regional alluvial plain covering the Central Andean forearc (Evenstar et al., 2017) (Fig. 1b). The paleosurface is now incised by two drainage types: (1) trunk channels with large drainage area (>500 km²) that dissect the landscape into discrete, low-relief interfluvies (pampas), and (2) ephemeral interfluvial channels with small drainage areas (<80 km²) disconnected from the trunk systems (Hoke et al., 2004; Binnie et al., 2020; Vega-Ruiz et al., 2025b) (Fig. 1c). Trunk-channel incision across the Central Depression began at ~7 and ~3 Ma north and south of 19.5°S, respectively, reflecting regional uplift and precipitation changes in the Western Cordillera–Altiplano (e.g., Binnie et al., in review; Coudurier-Curveur et al., 2015; García et al., 2011; Hoke et al., 2007). In contrast, interfluvial channels respond primarily to episodic

rainfall in the Precordillera, with past incision preserved as small paleochannels (Binnie et al., 2020; Vega-Ruiz et al., 2025b).

2.3 Interfluves and Faults of Pampa de Tiliviche and Pampa de Soga

Between 18.75° and 19.75°S, Quaternary counter-slope reverse faults deform Central Depression interfluves, producing subtle vertical offsets (Figs. 1a–b, 2). These faults are short (<20 km) and commonly branch into splays within gravel deposits, producing subtle faults and monoclines of <30 m in height (Vega-Ruiz et al., 2025b). Consequently, clear fault planes and geological markers are poorly developed. Limited erosion and incision further restrict exposure and structural constraints. At Pampa de Tana (19.35°S), fault uplift of the Pampa de Tana Anticline (PTA) of ~8–12 m/Myr since prior to ~5.6 Myr has been interpreted to have disrupted drainage by blocking and beheading interfluvial channels (e.g., Binnie et al., 2020; A. Binnie et al., in review; Vega-Ruiz et al., 2025b).

2.3.1 Pampa de Tiliviche

Pampa de Tiliviche is a ~50 × 15 km, west-sloping (~1.8°) interfluve between the Tana and Tiliviche trunk rivers (~19.5°S) (Fig. 1b). Its Cenozoic substrate comprises the Azapa Formation (Oligocene–Lower Miocene fluvial), Oxaya Formation (~26–19.4 Ma ignimbrites), and El Diablo Formation (19.4–8.3 Ma alluvial), overlain by Plio-Quaternary fluvial-alluvial deposits (Evenstar et al., 2017; García et al., 2013; García & Fuentes, 2012; Kirk-Lawlor et al., 2013). A single ³He cosmogenic exposure age dates the alluvial top surface at ~3.2 Ma (Evenstar et al., 2009) (Fig. 2a).

The NNW-striking Pampa de Tiliviche fault (PTF) is a reverse fault with averaged Plio-Quaternary uplift rates of ~5–10 m/Myr, forming a ~20 m high east-facing scarp (Vega-Ruiz et al.,

2025b) (Fig. 2). The PTF has an east-vergent positive-flower fault zone with west-dipping secondary faults and a main reverse plane of $72^{\circ}/291^{\circ}$. West-flowing braided channels of the interfluvial have ponded, forming an elongated claypan on the footwall of the PTF. Two major hanging-wall channels preserve small terraces: a warped northern channel that remains partly connected to the active network, whereas a southern channel is fully beheaded (Figs. 2 & S1, Supplementary Material 1). The PTF links northward with the PTA and both structures show evidence for southward younging of faulting (Binnie et al., 2020; Vega-Ruiz et al., 2025b) (Fig. 2a).

2.3.2 *Pampa de Soga*

Pampa de Soga is a $\sim 35 \times 12$ km, west-sloping ($\sim 1.1^{\circ}$) interfluvial of the Pampa del Tamarugal between the Soga and Aroma trunk rivers ($\sim 19.75^{\circ}\text{S}$) (Fig. 1b). Its Cenozoic substrate is equal to that of Pampa de Tiliviche, and the alluvial top surface is dated at ~ 5.5 Ma using cosmogenic ^3He (Evenstar et al., 2017) (Fig. 2b). The curved, NNW-striking Pampa de Soga Fault (PSF) is a reverse fault with averaged Plio-Quaternary uplift rates of <5 m/Myr forming a ~ 15 m high east-facing scarp displacing the planar pampa surface. Although the main fault zone is not exposed, fault planes of $52\text{--}76^{\circ}/210\text{--}235^{\circ}$ at the surface are reported (Vega-Ruiz et al., 2025b). Two major channels are preserved in the hanging wall of the PSF: a warped northern channel and a southern channel still connected to the inactive drainage network. Both preserve small terraces, and an abandoned meander occurs ~ 2 m above the northern channel floor (Fig. 2). The southward decrease in PSF scarp height and increase in overall hanging-wall incision relative to the northern PTA and PTF suggest a southward younging of Quaternary counter-slope fault activity within the given latitudes (Vega-Ruiz et al., 2025b).

3 METHODS

3.1 Geomorphic Mapping of the Interfluves

We carried out a geomorphic mapping of the Pampa de Tiliviche and Pampa de Soga interfluves, aiming to describe fault scarps and qualitatively evaluate the tectonic forcing of drainages. For this, we combined optical imagery (Bing Aerial, DigitalGlobe, SPOT-6), TerraSAR-X (12 m/px) with SPOT-6-derived (2 m/px), and UAV-derived DEMs (<5 m/px). UAV surveys were performed using a DJI Phantom 4 and Mavic 2 Pro drones flown at ~130 m altitude, acquiring a total of 3152 images with 80% frontal and 70% lateral overlap, covering over 6.74 km². Differential GNSS data were incorporated into photogrammetric processing in Metashape Pro following Carbonneau & Dietrich (2017).

Mapped drainage networks were classified into pre- and syn- faulting networks according to the evidence of fault-driven tectonic forcing in their development. Connected and beheaded channels of both classes were identified along the fault scarps. We also included the mapping and correlation of fluvial terraces between major interfluvial channels within each interfluve. Drainage mapping is summarized in Figure 6a.

Fault scarp segments were characterized by aspect, slope, tectonic crack density and vertical offset. The latter was estimated from 2D swath profiles following the approach of Yang et al. (2015) for counter-slope fault scarps (Fig. 3a). We extracted 85 swath profiles, 200–250 m wide and spaced 250 m along fault strike, from the TerraSAR-X and SPOT-6-derived DEMs using TopoToolbox 2 (Schwanghart and Scherler, 2014). Fault geomorphology is summarized in Figure 6b–e.

3.2 ^{10}Be Cosmogenic Exposure Dating

Our sampling targeted surfaces of abandoned fluvial features, including channels and terraces, and adjacent interfluvial surfaces. Samples consisted of quartz or quartz-rich pebbles and cobbles, we preferentially selected samples with evidence for fluvial rounding where present. Several samples per feature were collected from planar areas within a few tens of meters of each other. To avoid replicate sampling, we collected no more than one fragment from thermally shattered clasts, which are common in the region (McKay et al., 2009). Topographic production rate shielding was measured in the field with a compass and inclinometer.

Samples were prepared and measured at the University of Cologne. After crushing and sieving to 250–500 μm , samples were etched in dilute HF/HNO_3 , following Kohl and Nishiizumi (1992). Sample purity, particularly sufficiently low stable Al contents, was assessed by Inductively Coupled Plasma–Optical Emission Spectroscopy prior to adding a Be carrier and dissolving samples in concentrated HF/HNO_3 . Be was separated by stacked column chromatography; this and other preparation steps are described by (Binnie et al., 2015). Pressed targets were measured on CologneAMS and normalized to the Be standards of Nishiizumi et al. (2007). Chemical blanks were prepared and measured in tandem with the samples.

^{10}Be concentrations were blank-corrected following Binnie et al. (2019), and exposure ages were calculated using the online calculator of Balco et al. (2008). Ages are reported using the LSD production-rate scaling scheme of (Lifton et al., 2014), which is based on first-principles particle physics rather than fits to proxy data. Owing to the lack of proximal calibration datasets at comparable elevations, we use the global production rate of Borchers et al. (2016) with a nominal value of 3.92 atoms $\text{g}^{-1} \text{yr}^{-1}$ at high latitude and sea level. All laboratory and measurement data, along with calculator inputs are provided in Supplementary Material 1.

3.3 Quantification and Evolution of Fault Uplift and Slip Rates

Vertical offsets measured from swath profiles were converted to 2D fault throws accounting for fault dip, dip-slip ratio (S_l/S_d), scarp slope, and aspect (Benavente et al., 2022) (Table 1; Fig. 3). Because fault rakes are unavailable, expected rake and S_l/S_d ranges were considered for reverse (70–110°) and right-lateral reverse (110–180°) kinematics (Davis et al., 2011) (Table 1). Measurement uncertainties were propagated using a Monte Carlo approach (Mackenzie & Elliott, 2017; Fig. 3b). Although 3D throws were also calculated, the low slopes ($<3^\circ$), small slope differences ($<1^\circ$), and similar surface aspects across scarps indicate a negligible horizontal-slip component (Fig. 3c). Both 2D and 3D throws were projected onto the fault planes to derive slip values, which were combined with surface-exposure ages to estimate slip rates (Fig. 6; Table 1; Supplementary Material 2).

With the growing amount of instrumental data, it is known that faults slip rates are not constant but vary over time (Carena and Friedrich, 2018; Dolan et al., 2024). To resolve temporal variations in fault slip, we used fluvial terraces and incision as a proxy for tectonic uplift (Fig. 4a). Interfluvial channels are disconnected from trunk rivers and their terraces are confined to fault scarps. Given the extremely low Quaternary denudation and fluvial activity on the interfluvial scarps (Plaszcek, 2014), localized incision and terrace formation are therefore most likely primarily fault-driven, allowing incision rates to record temporal variations in tectonic uplift (e.g., Lavé & Avouac, 2001).

For each channel, terrace elevations were projected onto longitudinal profiles and compared with reconstructed linear, no-uplift profiles to estimate tectonic uplift since incision (Binnie et al., 2020) (Fig. 4b). At both sites, terraces are warped across the fault-scarp hinge, recording deformation by complex anticlinal folding and precluding direct slip-rate estimates from

vertical offsets. Assuming that temporal variations in dip-slip rates are reflected in uplift histories, we reconstructed the evolution of uplift and slip rates by combining ^{10}Be exposure ages of terraces and riverbeds with the maximum elevation difference between these warped markers and the riverbeds relative to their reconstructed undeformed longitudinal profiles (Fig. 4c-d).

3.4 Evaluation of Faulting Rates and Tectonic Forcing of Drainages by an Active Fault using Landscape Evolution Modelling

We used long-term geomorphically derived uplift rates as rock-uplift inputs to a landscape-evolution model to evaluate fault-driven reorganization of forearc drainage (Fig. 5). We compared a reference scenario without faulting (M_{ref}) with 11 scenarios (M_{1-11}) incorporating a counter-slope active fault with linearly increasing uplift, spanning the full range of uplift rates observed in nature. For each model, we quantified the number of incising hanging-wall channels (N_{ch}), ponded drainage area (A_p , km²), channel depth and width (D and W , m), and the channel deflection index (T_p) (Fig. 4). We used FastScape (Bovy & Lange, 2023), a Python-based model that efficiently solves the stream-power law through space and time (Braun & Willett, 2013), including erosion and sediment deposition (Yuan et al., 2019) (Equation 1):

$$\frac{\partial h}{\partial t} = U - K_f A^m S^n + K_d \nabla^2 h + \frac{G}{A} \int_A \left(U - \frac{\partial h}{\partial t} \right) dA \quad (1)$$

where h is the topographic elevation, t is time, U is the uplift rate, K_f is the fluvial erodibility, A is the upstream catchment area, S the local slope, K_d the hillslope diffusion coefficient, G the dimensionless deposition coefficient and m and n the stream power exponents.

Model domains were 200×200 km, with initial topography, uplift, and precipitation patterns based on a representative forearc profile and extended along strike, following Coudurier-

Curveur et al. (2015) (Fig. 4). The models evolved for 10 Myr, simulating ~1 km of Altiplano monocline growth relative to the Central Depression, including a 4 Myr pre-faulting phase that allowed drainage incision to develop before fault initiation, consistent with geological constraints (Binnie et al., 2020; Hoke et al., 2007; Jordan et al., 2010; Kirk-Lawlor et al., 2013). Coastal uplift rates and precipitation gradients were prescribed from Melnick (2016) and Karger et al. (2017, 2018), respectively (Fig. 1b). We used a m and n of 0.5 and 1 (Stock and Montgomery, 1999; Vega-Ruiz et al., 2025b) and K_d of 5×10^{-2} m²/yr for bedrock and sediments (Coudurier-Curveur et al., 2015; Neuharth et al., 2022). K_f integrates parameters such as precipitation, lithology, and vegetation, and therefore, has a wide variation from 10^{-7} to 10^{-2} m/yr (Stock and Montgomery, 1999; Whipple and Tucker, 1999). Following Coudurier-Curveur et al. (2015) and Yuan et al. (2019), we constrained $K_f = 5 \times 10^{-7}$ m/yr and G equal to the minimum precipitation rate (1 mm/yr), which best reproduced fluvial incision across the models.

The channel deflection index (T_p) was originally derived by Tomkin and Braun (1999) from the surface-process modelling framework of Kooi and Beaumont (1996) to describe the conditions under which streams are deflected by actively uplifting shutter ridges in the Otago Plains, New Zealand (Jackson et al., 1996). Within the stream-power framework, topographic evolution reflects the competition between tectonic uplift (U) and fluvial erosion (E), with E proportional to drainage area A^m and channel slope S^n (Equation 2):

$$\frac{\partial h}{\partial t} = U - E = U - K_f A^m S^n \quad (2)$$

For a fault producing a linear uplift gradient over distance L , with $m = 0.5$ and $n = 1$, the governing equation can be expressed in terms of slope change (Equation 3):

$$\frac{1}{L} \frac{\partial h}{\partial t} = \frac{U_f}{L} - \frac{K_f \sqrt{A} S}{L} \quad (3)$$

and solved using the erosion timescale $\tau = \frac{1}{K_f}$ (Whipple and Tucker, 1999) and the ratio r

$$= \frac{\sqrt{A}}{L} \text{ (Equation 4):}$$

$$\frac{\partial S}{\partial t} = \frac{U_f}{L} - \frac{rS}{\tau} \quad (4)$$

which has the following solution (Equation 5):

$$S = \left(S_0 - \frac{U_f \tau}{S_0 r L} \right) e^{-rt/\tau} + \frac{U_f \tau}{S_0 r L} \quad (5)$$

where S_0 is the initial slope, i.e., before the fault starts moving. We can render this equation dimensionless by dividing all terms by S_0 to obtain (Equation 6):

$$\frac{S}{S_0} = \left(1 - \frac{U_f \tau}{S_0 r L} \right) e^{-rt/\tau} + \frac{U_f \tau}{S_0 r L} \quad (6)$$

Nondimensionalization shows that the solution depends on how the ratio $T_p = \frac{U_f \tau}{S_0 r L}$ which compares to 1 (Tomkin and Braun, 1999), where U_f is the maximum fault uplift rate and S_0 is the pre-fault slope. T_p therefore represents the relative timescales of fault uplift and fluvial adjustment. At $T_p = 1$, uplift-induced slope changes are balanced by enhanced erosion and the channels remain effectively unperturbed. For $T_p > 1$, uplift exceeds the capacity of fluvial adjustment, producing slope increase and channel deflection, whereas $T_p < 1$ indicates that erosion outpaces uplift, and no significant channel perturbation is expected.

296 A critical catchment area (A_c), i.e., the required upstream area of channels crossing the
297 modelled fault to avoid deflection ($T_p = 1$), can be extracted as the dimensionless term is rendered
298 as $\frac{U_f}{K_f S_0 \sqrt{A}}$ (Equation 7-8):

$$299 \quad T_p = 1 \Leftrightarrow \frac{U_f}{K_f S_0 \sqrt{A_c}} = 1 \quad (7)$$

300 Hence:

$$301 \quad A_c = \left(\frac{U_f}{K_f S_0} \right)^2 \quad (8)$$

302

4 RESULTS

4.1 Tectonic Geomorphology of the Pampa de Tiliviche and Pampa de Soga Sites

4.1.1 Drainage Networks

Pre-faulting drainages form dense networks of narrow (5–10 m wide), shallow (<1 m deep) channels parallel to the interfluvial that are mostly connected and preserved drainage pattern across the fault trace, showing their formation predates faulting. They show no evidence of fault-enhanced incision and, where deformed, are typically beheaded without major changes in their patterns. These channels are crosscut and captured by syn-faulting drainages.

Syn-faulting drainages also form dense upstream networks but become fewer and larger when approaching fault scarps. Several of these channels are ponded along the footwall and feed breakthrough channels in the hanging wall. Breakthrough channels show evidence of progressive fault-driven incision, forming narrow (<20 m wide), deeply incised (>7 m deep) and sinuous channels near the scarp hinge that become broad (>200 m wide), shallow (<3 m deep) and braided channels downstream. Some of these channels are subsequently warped and beheaded. Both PTF and PSF have a pair of large syn-faulting channels across their hanging wall, preserving five and four terrace generations at Pampa de Tiliviche and Pampa de Soga, respectively (Fig. S2). At Pampa de Tiliviche, ephemeral drainages transporting fine sediments are diverted northward into one of these channels (Fig. S1).

4.1.2 Pampa de Tiliviche Fault

The northern segment comprises a NNW-striking scarp with a secondary, broad (<2° slope) scarp in front. Scarp height and vertical offset range from 1.6–9.57 m and 5.7–11.5 m, respectively. Despite its low height, the main scarp produces an abrupt change in interfluvial aspect and enhanced

hanging-wall incision in syn-faulting channels. Small secondary faults laterally offset channels, while en-echelon cracks oriented $\sim 18^\circ$ off the fault strike indicate minor oblique slip.

The central segment is characterized by a prominent NNW-striking scarp ($>2^\circ$), tight hanging-wall anticlines, and subparallel splays. Warped and beheaded syn-faulting channels occur throughout the segment. Cracks oriented $\sim 9^\circ$ off the fault strike cluster along the anticlines, indicating predominantly reverse slip. Scarp height and vertical offset range from 6.3–20.5 m and 19.4–27.5 m, respectively, with westward tilting of the hanging wall and folding producing offsets substantially greater than scarp heights. The southern segment comprises a prominent NNW-striking scarp ($>2^\circ$) with tight hanging-wall anticlines. Most channels in this segment correspond to beheaded pre-faulting drainages. Subparallel cracks oriented $<5^\circ$ off the fault strike cluster along the anticlines, indicating predominantly reverse slip. Scarp height and vertical offset range from 9.7–19.7 m and 11.6–27.1 m, respectively.

4.1.2 Pampa de Soga Fault

The northern segment is NE-striking, with a sharp scarp ($>2^\circ$ slope) that strongly disrupts the interfluvial drainage. Drainage consists of beheaded channels that pre-date faulting. Scarp height and vertical offset reach 10.4 and 16.8 m, respectively, but decrease where the strike rotates to NNE. The central segment is NW-striking, with a smoother scarp and deeply incised syn-faulting channels. Parallel cracks concentrate in the central portion, indicating predominantly reverse slip. Scarp height and vertical offset decrease southward from 7.2 to 1.2 m and 14.7 to 2.7 m, respectively. The southern segment forms a gentle NW-striking fold scarp. Pre-faulting drainages remain connected, while interfluvial disturbance and crack density are low. Scarp height and vertical offset reach 6.1 and 4.1 m, respectively.

4.2 Cosmogenic Exposure Ages Interpretation

Maximum blank subtractions amount to $< 5 \%$ of the total ^{10}Be measured in the samples. The ages derived from the background corrected concentrations show notable scatter from single surfaces but are consistent with the geomorphological setting in that the samples from the interfluvial top surface are consistently older than the ages from the channels and terraces (Figs. 2, 5 & Table 3). Below we report ages with one standard deviation external uncertainties, which include analytical as well as production rate uncertainties.

4.2.1 Interfluvial Surface Age

Interfluvial ages show relatively low scatter, although each site includes one notably younger sample (Table 3). Given the Late Cenozoic age of the Pampa de Tiliviche and Pampa de Soga interfluvial, sediment reworking is considered more likely to introduce scatter than significant ^{10}Be inheritance. We therefore use the oldest ^{10}Be ages, $2.61 \pm 0.37 \text{ Ma}$ and $5.00 \pm 1.52 \text{ Ma}$, as maximum abandonment ages for the Pampa de Tiliviche and Pampa de Soga interfluvial, respectively. Because the onset of surface displacement could post-date the age of the surface, the resulting long-term slip rates should be regarded as minimum estimates.

4.2.2 Riverbeds and Fluvial Terraces Ages

Ages from riverbeds and terraces are more scattered (Table 3). The terraces are straths but both the terraces and the channel bed surfaces may be mantled with cobbles reworked from upstream portions of the Late Cenozoic interfluvial, potentially carrying inherited ^{10}Be . We therefore use the youngest ages as the most conservative estimates of abandonment, following the approach used by Binnie et al (2020). At Pampa de Tiliviche, abandonment ages are $0.328 \pm 0.024 \text{ Ma}$ for the northern channel, $0.262 \pm 0.019 \text{ Ma}$ for terrace T3, and $0.182 \pm 0.013 \text{ Ma}$ for the

southern channel. At Pampa de Soga, the northern channel riverbed and terrace T2 yield abandonment ages of 25.4 ± 1.9 ka and 32.9 ± 2.4 kyr, respectively.

4.3 Patterns and Evolution of Fault Uplift and Slip Rates

4.3.1 Pampa de Tiliviche Fault

Using a maximum top-surface abandonment age of ~ 2.59 Ma, long-term minimum uplift and slip rates have primary modes of ~ 8.3 m/Myr, with maxima of ~ 10.3 and ~ 11.1 m/Myr, respectively (Fig. 7a–b). Uplift rates at the northern and southern channel sites are ~ 7.5 and ~ 9.0 m/Myr, respectively. Channels remain connected across the fault only where long-term uplift rates are < 1.7 m/Myr (Figs. 6a and 7). Considering that pre-faulting incision is minimum (< 1 m), terrace and riverbed warping indicates that only $\sim 20\%$ of cumulative uplift occurred between the top-surface abandonment and T3, whereas $\sim 70\%$ postdates abandonment of the riverbed (Fig. 8). Uplift rates therefore apparently increased to ~ 41.6 m/Myr at the northern site over the last ~ 330 kyr and ~ 131 m/Myr at the southern site over the last ~ 180 kyr (Fig. S2).

4.3.2 Pampa de Soga Fault

Using a maximum top-surface abandonment age of ~ 5 Ma, minimum long-term uplift and slip rates have primary modes of ~ 2.8 m/Myr, with maxima of ~ 3.4 and ~ 3.6 m/Myr, respectively (Fig. 7c–d). Uplift rates at the northern and southern channel sites are ~ 2.1 and ~ 0.6 m/Myr, respectively. Similar to the PTF, channels remain connected only where long-term slip rates are < 1.5 m/Myr (Figs. 6a and 7). The lower quality and resolution of the digital elevation model and smaller deformation produce noisy terrace profiles, allowing a reliable uplift history only at the northern channel (Fig. S2). There, terrace and riverbed warping indicate that only $\sim 20\%$ of cumulative uplift occurred between the top-surface and T2 abandonment, whereas $\sim 75\%$ postdates

riverbed abandonment. Uplift rates therefore apparently increased to ~300 m/Myr over the last ~25 kyr.

4.4 Landscape Evolution Modelling Patterns and Evolution of Fault Uplift Rates

The reference (M_{ref} ; Fig. 8a; Video 1; Supplementary Material 2) and synthetic-fault models (M_{1-12} ; Fig. 8b–d; Videos 2–4; Supplementary Material 2) reproduce channels that flow on the surface of the interfluvies with Precordillera-confined catchments, and trunk channels sourced from the Western Cordillera–Altiplano. They also capture the resulting fluvial and tectonic features observed at comparable uplift rates of studied faults, although modelled channels are wider, however, owing to differences in the semi-automatic detection of modelled valley shoulders (Fig. S3–6; Supplementary Material 1). Overall, depth (D) and width (W) increase with fault uplift, reaching maximum depths comparable to equivalent segments of the Tana, Tiliviche, and Aroma rivers. Similarly, erosion rates in modelled interfluvies and canyons increases with U_f reaching maximums of ~1.5 m/Myr and ~20 m/Myr, respectively.

At long-term uplift rates $U_f \leq 2.2$ m/Myr (M_{1-2}), (N_{ch}) of a number of incising hanging-wall channels remains constant, indicating largely undisturbed drainage networks. These rates are comparable to those of the northern PTF and southern PSF, where interfluvie channels remain connected. At $U_f \sim 3.3$ m/Myr (M_3), N_{ch} decreases, and ponded drainage area (A_p) increases as channels begin to be blocked and concentrate discharge into fewer breakthrough channels, consistent with geomorphic observations along the central segment of the PSF. From $U_f \sim 6.5$ m/Myr (M_6), a syn-faulting drainage network develops in the hanging wall alongside breakthrough channels and footwall ponds, as observed in the central segment of the PTF. With increasing uplift (M_{7-8}), N_{ch} and A_p progressively decreases and increases, respectively, reflecting enhanced channel beheading, blocking, and diversion toward trunk channels. Under this parametrization, beheading

and blocking are complete at $U_f \geq 9.8$ m/Myr (M_{9-10}). Small trunk channels begin to be deviated by the fault at $U_f \geq 12$ m/Myr (M_{11}) (Fig. 8d; Fig. S6; Supplementary Material 1). These last two configurations match the southern segment of the PTF, where interfluvial channels are completely blocked.

The median of the channel deflection index (T_p) across the modelled fault zone rapidly increases with uplift rate, from $T_p \sim 7$ in (M_I) to $T_p \sim 77.3$ in (M_{11}), indicating that channels that form on the surface of the interfluvial are very susceptible to faulting even at very low U_f . These values translate to a median critical catchment area ranging (A_c) from ~ 3 km² in M_I to ~ 361 km² in M_{12} (Fig. 8b–d).

5 DISCUSSION

5.1 Reliability of Uplift and Slip Rates Estimates

Our geomorphic mapping shows limited secondary faulting, few cross-cutting structures, and predominantly fault-parallel surface cracks. These observations indicate broadly consistent fault geometry and predominantly dip-slip kinematics along strike despite poor structural exposure (Davis et al., 2005; González et al., 2008; Vázquez et al., 2017), supporting the use of constant fault geometry for uplift and slip-rate estimates (Table 1). Channel classification at both faults further shows that drainage patterns reflect relative uplift magnitude and provide a qualitative proxy for uplift along individual structures (Fig. 6a). Furthermore, Monte Carlo simulations of Mackenzie and Elliott (2017) approach incorporating displacement and age uncertainties further improves the slip-rate estimates (Liu et al., 2021).

Scatter in cosmogenic exposure ages from individual sites may reflect post-depositional processes (Table 3). Post-depositional burial, exhumation of clasts from depth, or erosion of clast surfaces

can reduce ^{10}Be concentrations and produce ages younger than surface formation, whereas inherited ^{10}Be results in anomalously old ages (Evenstar et al., 2009, 2017). For the Pampa de Tiliviche and Pampa de Soga interfluves, we consider the oldest ages most reliable. This is consistent with studies of sediment reworking on marine terraces under similar climatic conditions, where old surfaces are particularly susceptible to reworking biases (Binnie et al., 2016). Transported clasts may retain inherited ^{10}Be (Anderson et al., 1996), but the antiquity of Atacama surfaces (Placzek et al., 2010; Amundson et al., 2012; Evenstar et al., 2017; Binnie et al., 2020; Ritter-Prinz et al., 2026) means that much of this inherited component may have decayed since deposition. Although this approach best fits the geomorphic setting, averaging fault slip over the full age of the interfluve could underestimate rates during younger periods of enhanced activity if the faults are substantially younger than the displaced surfaces (Salomon et al., 2013). However, stream beheading and, locally, antecedence indicate that fluvial activity continued while parts of the pampa were being uplifted and isolated from channels, consistent with fault activity around the same time as surface abandonment. Conversely, for abandoned channels and fluvial terraces crossing the interfluves, we consider the youngest ages most reliable. These deposits were likely derived partly from reworked pampa sediments upstream and, given their younger ages relative to the interfluve surfaces, inheritance provides a plausible explanation for older anomalous ages. This interpretation is consistent with Binnie et al. (2020), who showed that ^{10}Be concentrations in apparently active paleochannel sediments may reflect reworking of upstream material with complex exposure histories.

The channel-bed and T3 ages from the northern and southern PTF are scattered (Table 2). Although additional sampling would better constrain biases such as inheritance, their age ranges do not alter the conclusion that substantial uplift occurred during a more recent period than earlier

offset of the interfluvial surfaces. This interpretation is less straightforward for the PSF, where the pampa surface preserves less evidence of channel abandonment or persistence during uplift. Instead, the similar ages of the slightly elevated meander and adjacent channel suggest that recent incision affected both surfaces contemporaneously, consistent with recent tectonic activity.

Overall, the substantially older interfluvial ages versus the ages for the terraces and channel beds result in lower long-term average uplift rates than those inferred from younger geomorphic markers. This difference may reflect incomplete temporal sampling of the geological record (Gingerich, 2021) or genuine acceleration of slip during the Pleistocene. Distinguishing between these alternatives requires additional age constraints and is important for assessing the recent tectonic activity and seismic potential of the region.

5.2 Climatic Controls on Landforms Development and Preservation

The studied interfluvial channels have small catchments restricted to the Precordillera (Fig. 1c), making them largely insensitive to High Andes precipitation and reliant mainly on Pacific moisture (Placzek et al., 2009; Vuille & Keimig, 2004). Our geomorphic mapping indicates that drainage incision and network reorganization are primarily tectonically driven across the study area with the main changes occurring across fault scarps. This is particularly evident in the development of narrow, deeply incised syn-faulting channels and terraces across fault hinges, which rapidly widen downstream into braided systems lacking terraces (Fig. S2). Despite the evidence of recent sediment transport in the northern PTF channel, it is highly likely that this is restricted to short distances, fine-grained material, and unwarped channel sections (Fig. S1), as riverbed exposure ages indicate that sustained transport of coarser material has not occurred along these channels for hundreds of kyr.

The fault scarps and associated landforms documented here, together with examples from the Atacama Fault System (Masana et al., 2005; Carrizo et al., 2008; Gonzalez et al., 2025; Rivera et al., 2025), demonstrate how hyperaridity favors the preservation of Late Quaternary tectonic landforms. Under slightly wetter conditions, comparable features would likely degrade rapidly and remain identifiable for only tens to hundreds of kyr (Hanks et al., 1984; Arrowsmith et al., 1998; Marliyani et al., 2016; Sare et al., 2019; Rivera et al., 2025).

5.3 Thresholds for Tectonic Forcing of Drainages

The progressive increase in modelled ponded drainage area (A_p) and reduction in the number of breakthrough channels in the hanging wall (N_{ch}) with increasing (U_f) demonstrate how tectonic uplift reorganises drainage networks. The high values of the channel slope index (T_p) further indicate that even relatively low fault uplift rates can impose substantial slope perturbations on small channels. This results from the low-relief landscape and the small upstream catchment areas of channels crossing the fault zones. Where drainage areas remain below the critical area (A_c) required for fluvial incision to compensate for tectonic uplift, channels cannot fully restore their pre-existing slope following fault displacement. The magnitude of this response is sensitive to the prescribed precipitation rates and fluvial erodibility (K_f) (Coudurier-Curveur et al., 2015). However, the models reproduce the observed deformation and drainage deflection patterns at their respective U_f , while modelled erosion rates across both interfluves and canyons are consistent with estimates from the Central Depression (Placzek et al., 2010, 2014) and incision rates of the Lluta, Tiliviche, and Tana rivers (García et al., 2011; Placzek et al., 2010, 2014) (Fig. S3–6, Supplementary Material 1). This agreement suggests that the prescribed parameter range is consistent with the long-term geomorphic behaviour of the study area.

In nature, most of the hyperarid core of the Atacama lies within the low-relief region between the Coastal Cordillera and Precordillera foothills. Modelling and field observations therefore suggest that drainage networks across the absolute Atacama are particularly susceptible to tectonic forcing by upper-plate structures. There, planar interfluvial channels have limited capacity to compensate for fault scarps and folds through enhanced incision. This interpretation is consistent with field evidence for tectonic control on drainage evolution across the absolute Atacama (Riquelme et al., 2003; Ritter et al., 2018b; Binnie et al., 2020; Aránguiz-Rago et al., 2025a, 2025b). Fluvial incision may be further inhibited by widespread gypsisol development, which indurates alluvial deposits and can increase their resistance to surface erosion (Rech et al., 2019; Ritter et al., 2022). In principle, channels flowing in the interfluvial surfaces could compensate for tectonic relief through sediment aggradation during sporadic high-magnitude rainfall events (Humphrey and Konrad, 2000). However, evidence for recent activity in these channels is largely restricted to the transport of fine-grained material, and the claypan developed on the PTF footwall remains well below the level required to overtop the fault scarp.

In contrast, large trunk channels in the Central Depression are capable of maintaining their drainage pathways across the studied faults because their catchment areas substantially exceed the critical drainage area. The Tiliviche, Tana and Aroma rivers have present-day upstream catchment areas of ~500, ~800 and 1500 km², respectively. For the adopted K_f and representative slope, these drainage areas correspond to critical uplift rates of approximately ~240, ~300 and ~415 m/Myr, respectively, above both the modelled uplift rates and short-term uplift estimated for the studied faults. Thus, their present-day catchment areas provide sufficient erosional capacity to maintain cross-fault connectivity under the uplift rates considered here. However, these rivers have subsequently captured drainage from surrounding interfluvial surfaces (Binnie et al., 2020), so their present-

day catchment areas likely overestimate those available during earlier stages of their development. The forcing of relatively small trunk channels in scenario M_{II} therefore indicates that counter-slope faults can influence drainage evolution at intermediate catchment scales. The studied faults may consequently have influenced the evolution and pathways of these rivers before major incision at ~3 Ma by diverting drainage towards fault tips (Video 4; Supplementary Material 2), as proposed by (Vega-Ruiz et al., 2025b).

Upper-plate faulting may also have influenced regional base-level organisation. Faulting has been proposed to modulate the hydrological budget of the Tiliviche palaeolake, which acted as an intermediate base level for the Tana and Tiliviche rivers between ~10 and ~3 Ma, by promoting drainage reorganisation in the Central Depression (Kirk-Lawlor et al., 2013). FastScape cannot fully resolve internally drained systems controlled by local minima. Testing these configurations therefore requires landscape-evolution models explicitly designed to simulate closed basins (Garcia-Castellanos and Jiménez-Munt, 2015; Gailleton et al., 2024), although at greater computational cost.

5.4 Tectonic and Seismic Implications of the Studied Active Slow Faults

Post-riverbed warping indicates that most PTF and PSF displacement occurred since the Middle and Late Pleistocene, respectively. Despite uncertainties in establishing robust slip chronologies, the younger geomorphic markers record higher rates than the long-term averages. The younger displaced markers at the PSF relative to the PTF further support southward reactivation of counter-slope faults across the Central Depression (Binnie et al., 2020; Vega-Ruiz et al., 2025b) (Fig. 10a-b).

Regionally, the consistent east-vergent reverse motion and similar position (~45–50 km from the coast) of the studied and related faults suggest a >100-km-long, coherent fault system deforming the western Chilean forearc during the Quaternary (Fig. 10a). These structures are likely Mesozoic in origin and underwent repeated Cenozoic reactivation (Carrizo et al., 2008; García et al., 2013), similar to the Atacama Fault System (Allmendinger & González, 2010; González et al., 2003, 2006). Constraining their Late Cenozoic geometry and evolution at depth requires imaging beneath the thick Cenozoic cover, ideally through seismic imaging and geological balancing (Labbé et al., 2019; Martínez and Fuentes, 2022; Fuentes et al., 2023).

The spatial correspondence between faulting and seismicity provides further evidence for an active upper-plate fault system bounding a relatively deformed western forearc block. High-resolution upper-plate microseismicity reveals western and eastern clusters separated by a gap beneath the Central Depression (Fig. 10c) (Riedel-Hornig et al., 2025). The western cluster has a west-dipping limit that coincides with the counter-slope fault scarps, whereas the eastern cluster is associated with a major forearc structural and rheological boundary (Farías et al., 2005; Bloch et al., 2014; Salazar et al., 2017). The seismicity gap has been attributed to a relatively undeformed, low-permeability crustal block between major forearc fault systems (Cabrera et al., 2021; Contreras-Reyes et al., 2021; Hernández-Soto et al., 2026). The western fault systems may also provide pathways for slab-derived fluids into the forearc, potentially increasing pore pressure and influencing upper-plate seismicity (Peña et al., 2025).

Long- and short-term slip rates of the studied faults account for only 0.005–0.016% and 0.2–0.45% of the plate convergence rate, respectively, indicating that the Central Depression accommodates only a small fraction of the convergence within an otherwise highly active forearc (Landgraf et al., 2017). Nevertheless, the long-term rates are comparable to those of slow active

reverse faults (<1 mm/yr) in the Andes and worldwide, many of which are seismogenic (Melnick et al., 2006; Wen-Jun et al., 2013; Siame et al., 2015; Barnes and Ghisetti, 2016; Riesner et al., 2017; Shao et al., 2019; Costa et al., 2019; Liu et al., 2019; Rimando et al., 2021; Vega-Ruiz et al., 2025a) (Fig. 11b). Empirical fault-length–magnitude relationships (Thingbaijam & Mai, 2020; Thingbaijam et al., 2017), indicate that the studied faults could generate M6.1–6.8 earthquakes, highlighting an understudied seismic-hazard source whose structural role in forearc deformation remains poorly constrained (Fig. 10).

6 CONCLUSIONS

We provide geomorphic evidence that slow counter-slope faults systematically reorganise drainage networks across the hyperarid Atacama Desert during the Quaternary. The studied faults record averaged maximum long-term uplift rates of ~ 11 m/Myr, with substantially higher rates recorded by Middle to Late Pleistocene geomorphic markers, indicating recent enhanced activity that may reflect acceleration or episodic reactivation.

Numerical landscape-evolution models integrating east-vergent faulting reproduce the observed progression from blocking, beheading and ponding of interfluvial channels to the tectonic forcing of trunk channels. Drainage response is strongly scale dependent: small interfluvial channels are readily reorganized by faulting, whereas large trunk channels reaching the wetter Western Cordillera and Altiplano can maintain cross-fault connectivity under the observed long-term uplift rates. Together with field observations, these results show that the low-relief landscape and limited catchment areas across the forearc of set by the hyperarid conditions of the Atacama Desert allow even subtle tectonic uplift to outpace fluvial incision.

The consistent reverse kinematics, geometry and along-strike distribution of the studied and related faults, together with their recent deformation and spatial correspondence with upper-plate seismicity, indicate a >100-km-long system of east-vergent upper-plate faults deforming the western Andean forearc. Empirical fault-length–magnitude relationships yield characteristic magnitudes of M6.1–6.8 for the mapped fault lengths, providing an estimate of potential earthquake size but not recurrence or rupture likelihood. These results identify slow counter-slope faults as potentially seismogenic structures and highlight the need for further geophysical and structural investigations to constrain their subsurface geometry, connectivity and earthquake potential.

APPENDIX

The Supplementary Material 1 contains further descriptions of the claypan and active drainage network in the Pampa de Tiliviche site, as well as outputs from the numerical landscape evolution modelling. The Supplementary Material 2 contains the summary data frames for the estimation of the fault uplift and slip rates and videos of relevant outputs of the landscape numerical modelling. This material is referenced in the main manuscript.

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FIGURES AND TABLES

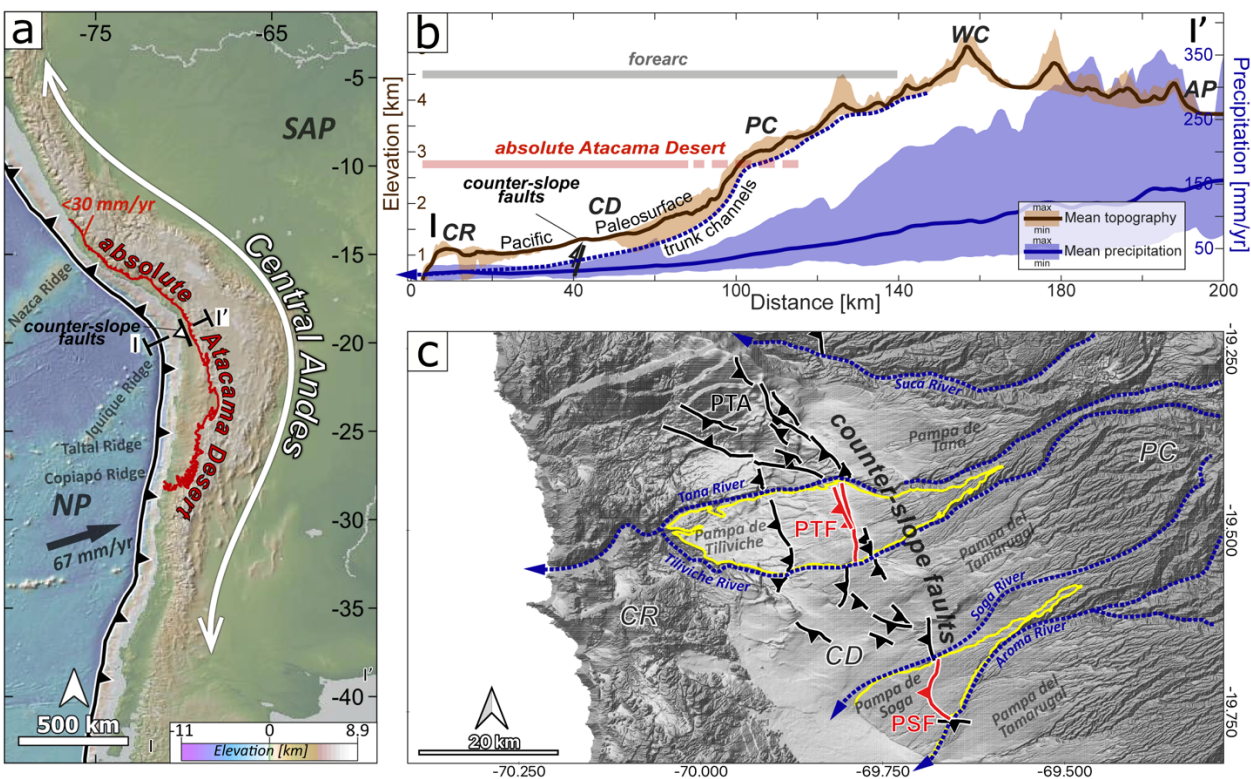


Figure 1: Regional setting. (a) Central Andes, Atacama Desert, and counter-slope faults of the forearc. Extension of the absolute Atacama Desert according to Latorre et al. (2002). The black arrow indicates the plate convergence vector (Norabuena et al., 1998). NP = Nazca Plate; SAP =

1026 South American Plate; **(b)** Geomorphologic and precipitation swath profiles of the Central Andean
1027 forearc across the study area. Topographic and climatic profile widths are 20 km and 50 km,
1028 respectively. Precipitation rates from Karger et al. (2017, 2018). AP = Altiplano; CD = Central
1029 Depression; CR = Coastal Range; PC = Precordillera; WC = Western Cordillera. Location of this
1030 profile is shown in (a). **(c)** Counter-slope faults along the Central Andean forearc. Studied
1031 interfluvies and faults are highlighted in yellow and red, respectively. Blue arrows are trunk
1032 channels. PTF = Pampa de Tiliviche fault; PSF = Pampa de Soga fault.

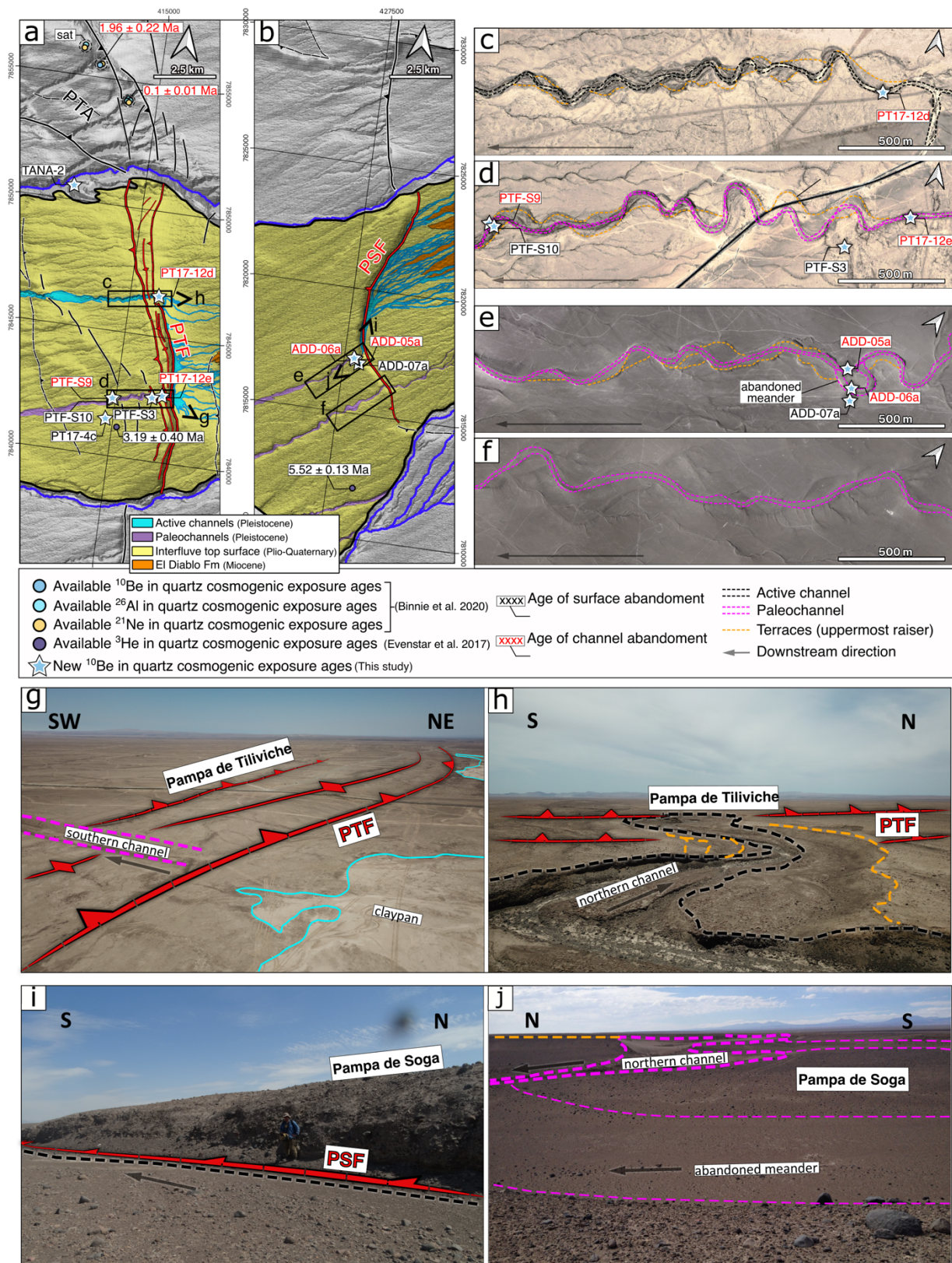


Figure 2: Close-up view of the faults scarps and dating sites (a-b) Pampa de Tiliviche and Quebrada Soga faults with available exposure age dating and new sampling dating sites. Detailed views of the studied channels incising the (c-d) Pampa de Tiliviche and (e-f) Quebrada Soga faults. Perspective views of the scarp and deformed channels along the (g-h) Pampa de Tiliviche and (i-j) Quebrada Soga faults. Location of river channels close-ups and field photos are shown in (a) and (b), respectively.

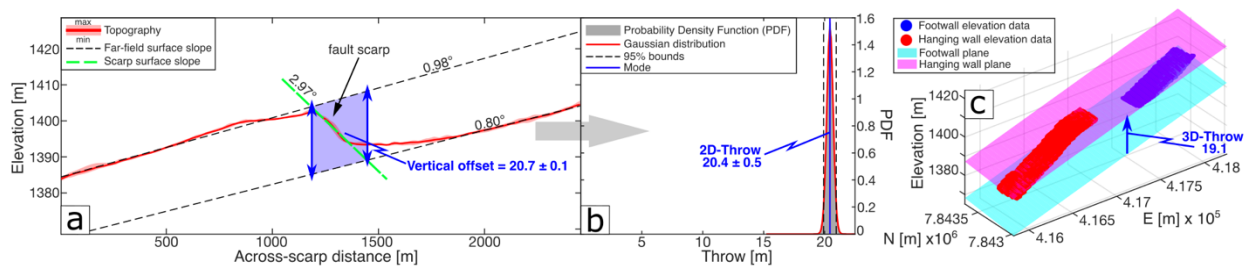


Figure 3: Vertical offset quantification. (a) Example of swath profile and quantification of vertical offset following Yang et al., (2015). The blue shaded represents the range between the minimum and maximum vertical offset estimates. **(b)** Monte Carlo simulation used to estimate the 2D-throw estimation based on the respective vertical offset quantification. **(c)** 3D-throw estimation based on the representative planes from planar geomorphic markers (i.e., interfluvial top surface) across the fault following. FW = Footwall; HW = Hanging wall. 2D- 3D- throw estimation approaches from Mackenzie & Elliott (2017).

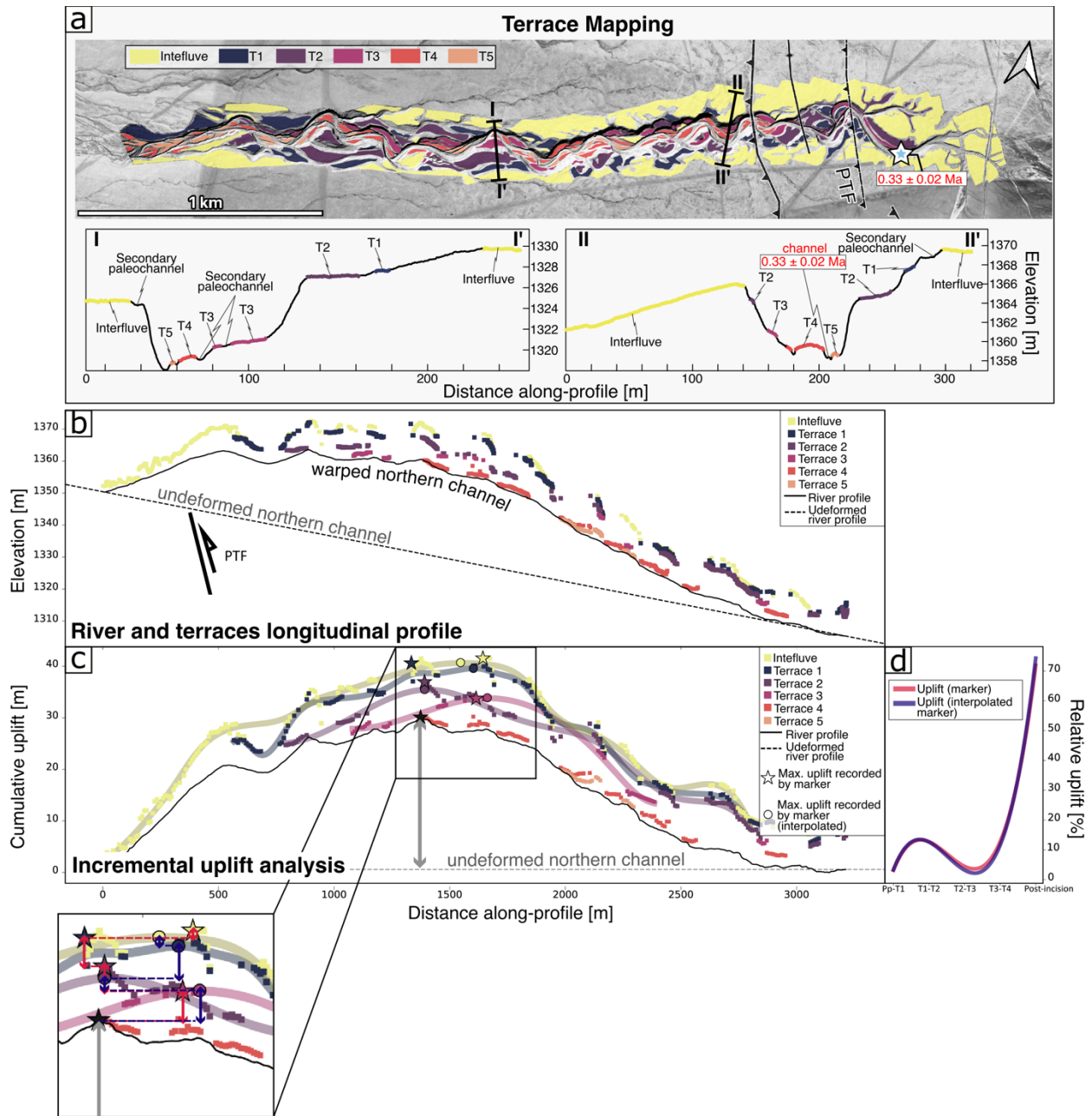
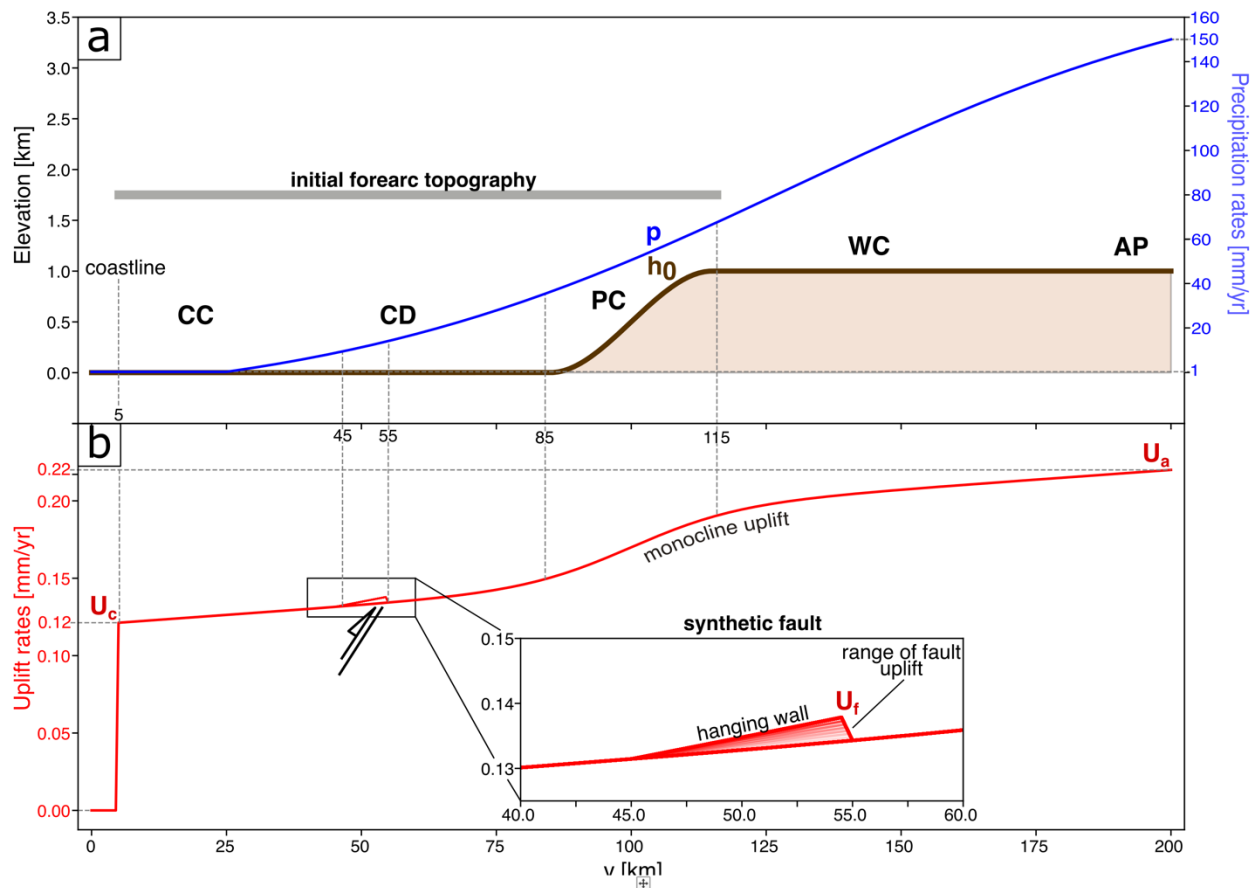


Figure 4: Geomorphic approach for incremental fault uplift and slip quantification. (a) Geomorphic mapping of terraces of the northern fluvial channel along the Pampa de Tiliviche fault (PTF). **(b)** Projection of elevation data of each mapped surfaces into the fluvial channel profile. **(c)** Analysis for each incremental uplift step based on mapped terraces, assuming a uniform undeformed river profile. Outset is a zoom in view of the maximum uplift recorded by each surface. **(d)** Contribution to total uplift of each incremental uplift step.



Note	Input parameter	Value	Unit
x, y	Model dimension	200×10^3	m
$\Delta x, \Delta y$	Cell size	500	m
t	Running time	10	Myr
Δt	Timestep	1	Myr
h_0	Initial topography		m
p	Precipitation ^a		mm/yr
U_c	Coastal uplift ^b	0.12	mm/yr
U_a	Andean uplift ^c	0.22	mm/yr
U_f	Fault uplift	0 – 0.012	mm/yr
ΔU_f	Fault uplift step	0.001	mm/yr
K_d	Hillslope diffusion coefficient ^c	5×10^{-2}	m^2/yr
K_f	Fluvial erodibility	5×10^{-7}	m/yr
G	Deposition coefficient	0.001	

Note	Output scenario	Fault uplift rate
M_{ref}	Reference scenario	0 mm/yr
M_{1-12}	Faulting scenarios 1 – 12	0.001 – 0.012 mm/yr

Note	Output parameter	Unit
N_{ch}	Number of hanging wall incising channels	
A_p	Ponded drainage area	m^2
D	Channels depth	m
W	Channels width	m
T_p	Channel deflexion index	
A_c	Critical catchment area	km^2

1055

1056 **Figure 5: (a) Landscape evolution model (LEM) set up.** Initial topography, input precipitation

1057 rates, and **(b) uplift rates** used in the numerical LEM. These parameters are extended across the x-

1058 distance axis. Left table shows a summary of all input parameters. The table on the left summarizes

1059 the model input parameters, whereas the tables on the right summarize the model runs and output

metrics. AP = Altiplano; CD = Central Depression; CR = Coastal Range; PC = Precordillera; WC = Western Cordillera. a = Karger et al. (2017, 2018); b = Melnick (2016); c = Coudurier-Curveur et al. (2015).

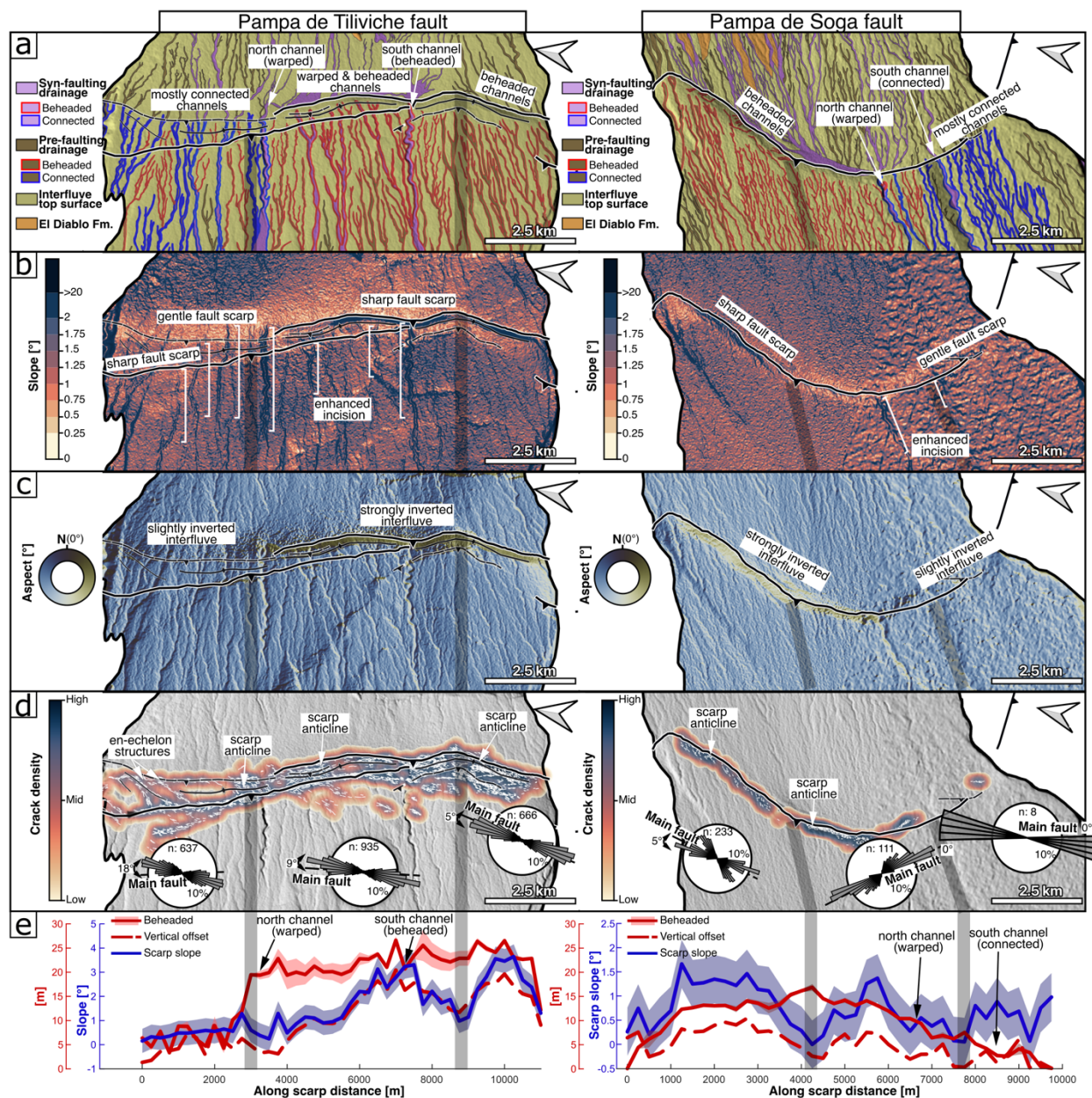


Figure 6: Geomorphic mapping of the studied interfluvies. (a) Classification of drainage networks, **(b)** slope maps highlighting faults scarps sharpness and differential incision degree of channels, **(c)** aspect map highlighting disrupted portions of the pediplains and channels, and **(d)**

crack density and orientation along the fault scarps and. **(e)** Height, vertical offset and scarp face slope and aspect of fault scarp along the fault traces. Segments are delimited by black dashed lines.

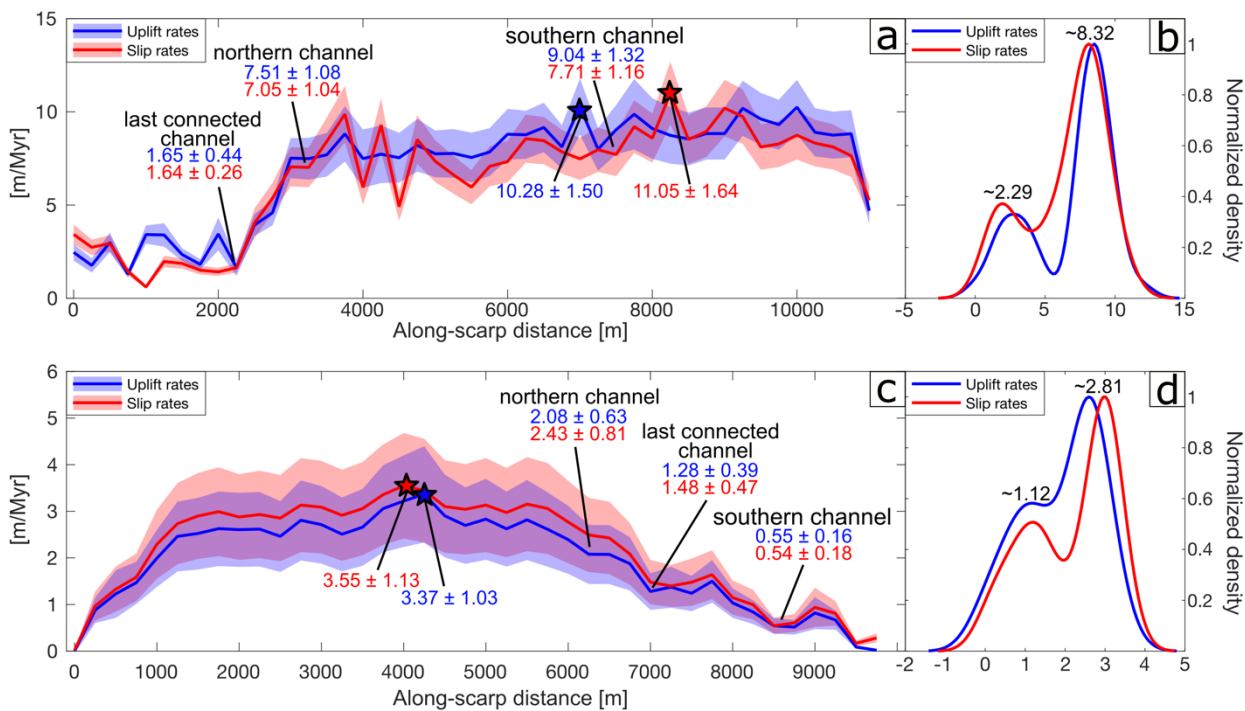


Figure 7: Along-scarp distribution and PDF of long-term slip and uplift rates. (a-b) Pampa de Tiliviche site. (c-d) the Pampa de Soga site.

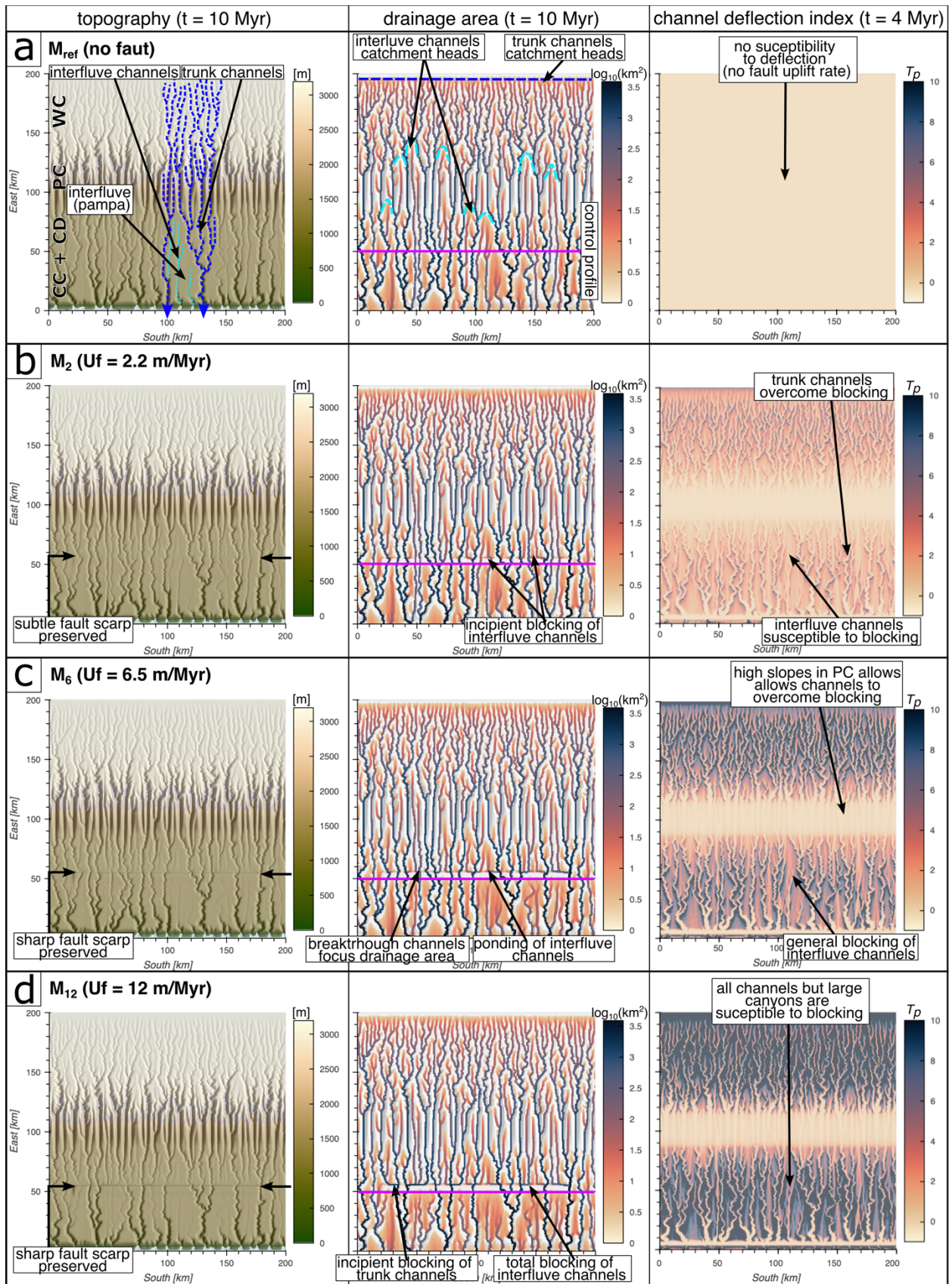
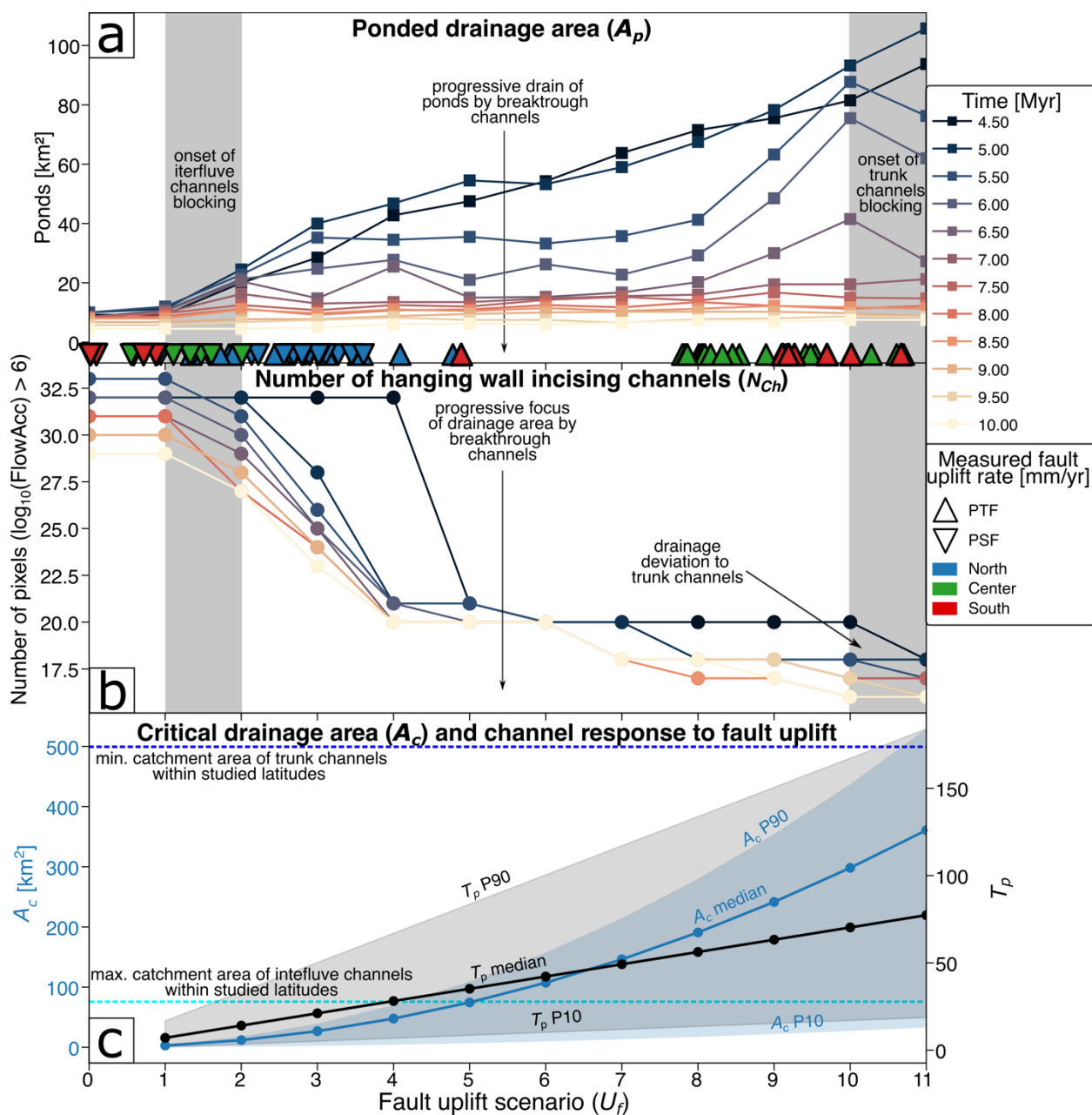


Figure 8: Geomorphic insights from numerical LEM. (a) Reference model of the forearc at the latitudes of the study area at $t = 10$ Myr. (b, c, d) Landscape evolution models of the forearc with a counter-slope fault imposing a long-term homogeneous uplift rate of ~ 2.2 m/Myr, ~ 6.5 m/Myr, and ~ 9.8 m/Myr, respectively. The pink line at East = 50 km is the control profile where parameters shown in Fig. 10 are measured.



1079 **Figure 9: Output parameters from LEM. (a)** Ponded drainage area along footwall of fault. **(b)**
1080 Number of fluvial channels incising the hanging wall of the fault, including trunk channels. **(c)**
1081 Median deflection index (T_p) and critical drainage area (A_c) measured in models at a given fault
1082 uplift rate. Catchment areas of interfluvial and trunk channels agree with modelling observations at
1083 which they get blocked.

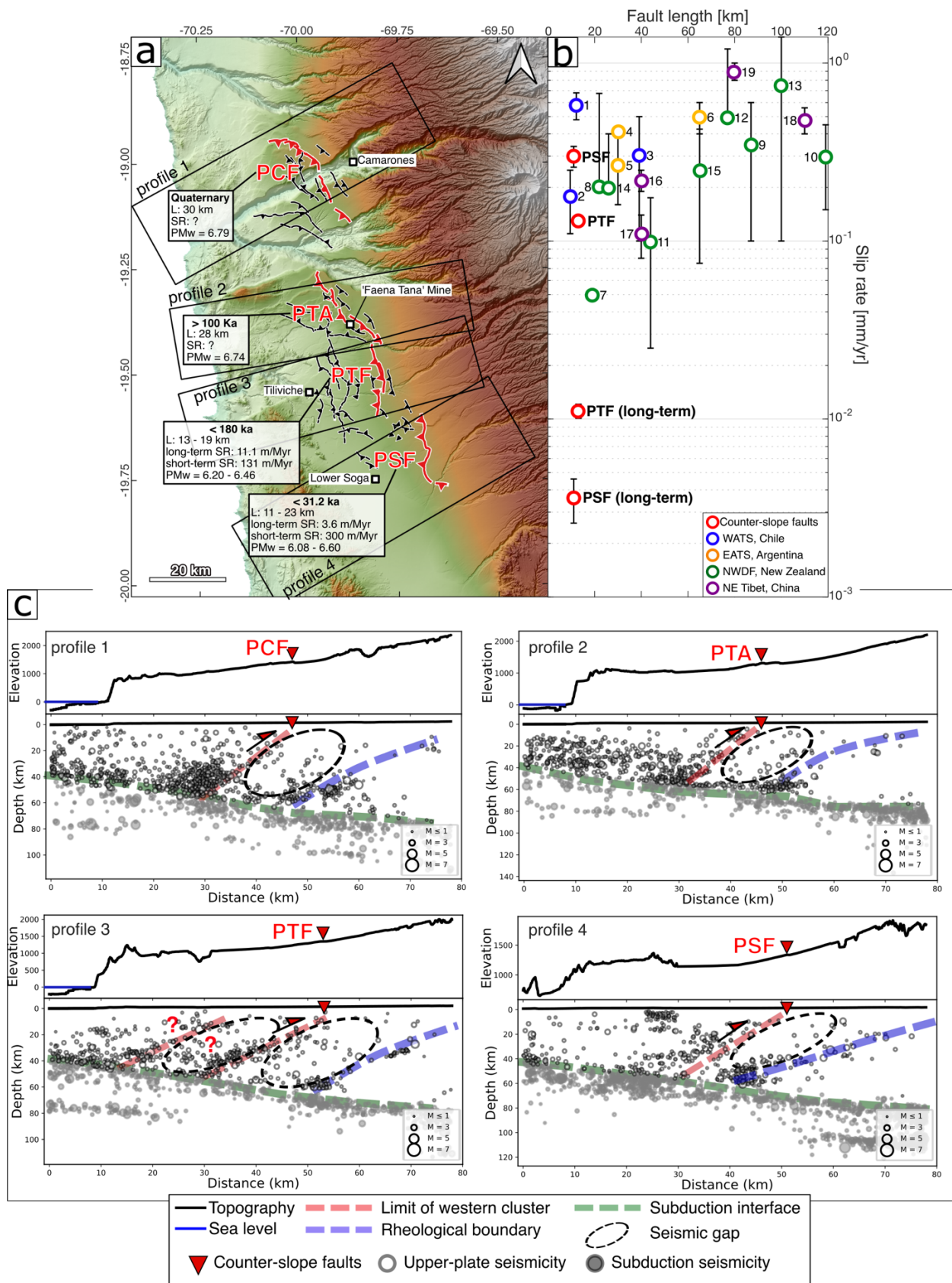


Figure 10: Seismic parameters of the forearc counter-slope faults. (a) Counter-slope faults and summary of their Quaternary activity and seismic potential. Fault trace from (Vega-Ruiz et al., 2025b). PCF = Pampa de Camarones fault; PTA = Pampa de Tana Anticline; PTF = Pampa de Tiliviche fault; PSF = Pampa de Soga fault **(b)** Slip rates versus fault length for slow slipping active reverse faults (< 1 mm/yr) coloured by tectonic setting. 1 – Mesamavida (Vega-Ruiz et al., 2025a); 2 – Huepil (Melnick et al., 2006); 3 – San Ramon (Riesner et al., 2017); 4 – La Rinconada (Rimando et al., 2021); 5 – Las Penas (Costa et al., 2019); 6 – Eastern Sierra de Palo (Siame et al., 2015); 7-15 – Elizabeth, Razorback, Khurangi, Farewell, Fouldwind 1-4 & Kongahu (Barnes and Ghisetti, 2016); 16-17 – Jiayuguan & Jintanan (Wen-Jun et al., 2013); 18 – Zongwulong Shan (Shao et al., 2019); 19 – Yumen–Beidahe (Liu et al., 2019). EATS = East Andean Thrust System; NWDF = North Westland Deformation Front; WATS = West Andean Thrust System **(c)** Profiles of microseismicity showing the occurrence of a west-dipping ramp limit matching location of the counter-slope fault scarps. Location of profiles is shown in (a). Microseismicity catalogue from Riedel-Hornig et al. (2025).

Table 1: Fault parameters used for the calculation of 2D-, 3D-throw, and dip-slip.

<i>Fault parameters</i>	<i>Pampa de Tiliviche</i>	<i>Quebrada de Soga</i>
<i>Azimuth / Dip of fault plane (°)</i>	$352 \pm 14 / 72 \pm 10$	$358 \pm 24 / 62 \pm 10$
<i>Slope / Aspect fault scarp (°)</i>	$1.51 \pm 0.5 / 85 \pm 10$	$0.7 \pm 0.5 / 101 \pm 10$
<i>Dip tolerance of surfaces (3D-throw) (°)</i>	5	
<i>Rake (°)</i>	Reverse 90 ± 20	Right-lateral 145 ± 35
<i>Sl/Sd</i>	0.22 ± 0.18	reverse 1.87 ± 2.66

1100 **Table 2:** Summary of ^{10}Be cosmogenic nuclide exposure ages. Shaded samples are considered as
1101 most representative of the true ages for the respective sampled feature.

<i>Site</i>	<i>Feature</i>	<i>Sample</i>	<i>Age (Myr)</i>	<i>External uncertainty (1σ)</i>
<i>Pampa de Tiliviche</i>	<i>Pediplain</i>	PT17-4	2.33	0.32
		PTFS-3a	2.61	0.36
		PTFS-3b	1.91	0.22
		PTFS-3c	1.58	0.16
		TANA- 2a	1.99	0.23
		TANA- 2b	2.61	0.36
	<i>Northern channel (riverbed)</i>	PT17- 12a	0.328	0.024
		PT17- 12d	1.13	0.11
		PT17-12f	1.35	0.14
	<i>Southern channel (T3)</i>	PTFS- 10a	0.262	0.019
		PTFS- 10b	0.535	0.042

<i>Quebrada de Soga</i>		PTFS-10c	0.410	0.031
	<i>Southern channel (riverbed)</i>	PT17-2d	0.735	0.067
		PT17-2e	0.788	0.067
		PTFS-9	0.182	0.013
	<i>Pediplain</i>	ADD-07a	4.130	0.926
		ADD-07b	3.850	0.789
		ADD-07c	4.998	1.519
		ADD-07d	4.130	0.932
		ADD-07e	2.547	0.349
	<i>Northern channel (meander/T2)</i>	ADD-06a	0.0451	0.0033
		ADD-06b	0.0329	0.0024

	ADD-06c	0.154	0.011
<i>Northern channel</i> <i>(riverbed)</i>	ADD-05a	0.0254	0.0019
	ADD-05c	0.0300	0.0022
	ADD-05d	0.0366	0.0026
	ADD-05e	0.0329	0.0023