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1 **Kinematics of Active Deformation and Possible Segmentation of Seismic Slip along**
2 **the Foothills of the Western Kunlun (China).**

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16
17 **Key Points:**

- 18 • Slip rate on the blind footwall ramps beneath the Hotan anticline estimated to 1.2-2.8
19 mm/yr.
- 20 • Similar slip rate on the blind ramps all along the foothills, but possible structural and
21 kinematic segmentation of active deformation.
- 22 • Moderate seismicity along the foothills, but possibly large rare earthquakes rupturing the
23 wide frontal thrust sheet.
24

25 Abstract

26 The Tibetan Plateau stands as a prominent topographic feature at the Earth's surface,
27 characterized by intense seismic activity, in particular along its bounding mountain ranges. To
28 the northwest, the Western Kunlun Range has received increasing attention since the 2015
29 Pishan earthquake but its kinematics of deformation remain to be properly documented. Here, we
30 analyse the terrace record of active deformation along the Karakash River, where it crosses the
31 Hotan anticline. We date terraces using in-situ produced cosmogenic isotopes, and show that
32 terrace incision and uplift are spatially correlated with blind duplex ramps beneath the anticline.
33 From there, we quantify the slip rate of the overall duplex to be 1.2-2.8 mm/yr over the last ~250
34 kyr. Our data are not able to resolve the detailed kinematics on each blind ramp and we cannot
35 exclude that several of them are active at places along the anticline. By comparing to available
36 data, we propose that the system of blind structures all along the foothills of the Western Kunlun
37 has an overall slip rate of ~2 mm/yr. However, the way this slip rate is partitioned on the various
38 blind ramps is expected to vary along strike. The structural segmentation of the foothills may
39 therefore explain the moderate recorded seismicity in this region. Because this slip is transmitted
40 upward and forward onto the Mazar Tagh wide and geometrically simple frontal thrust sheet, we
41 question the possibility of large – but rare – earthquakes rupturing this structure.

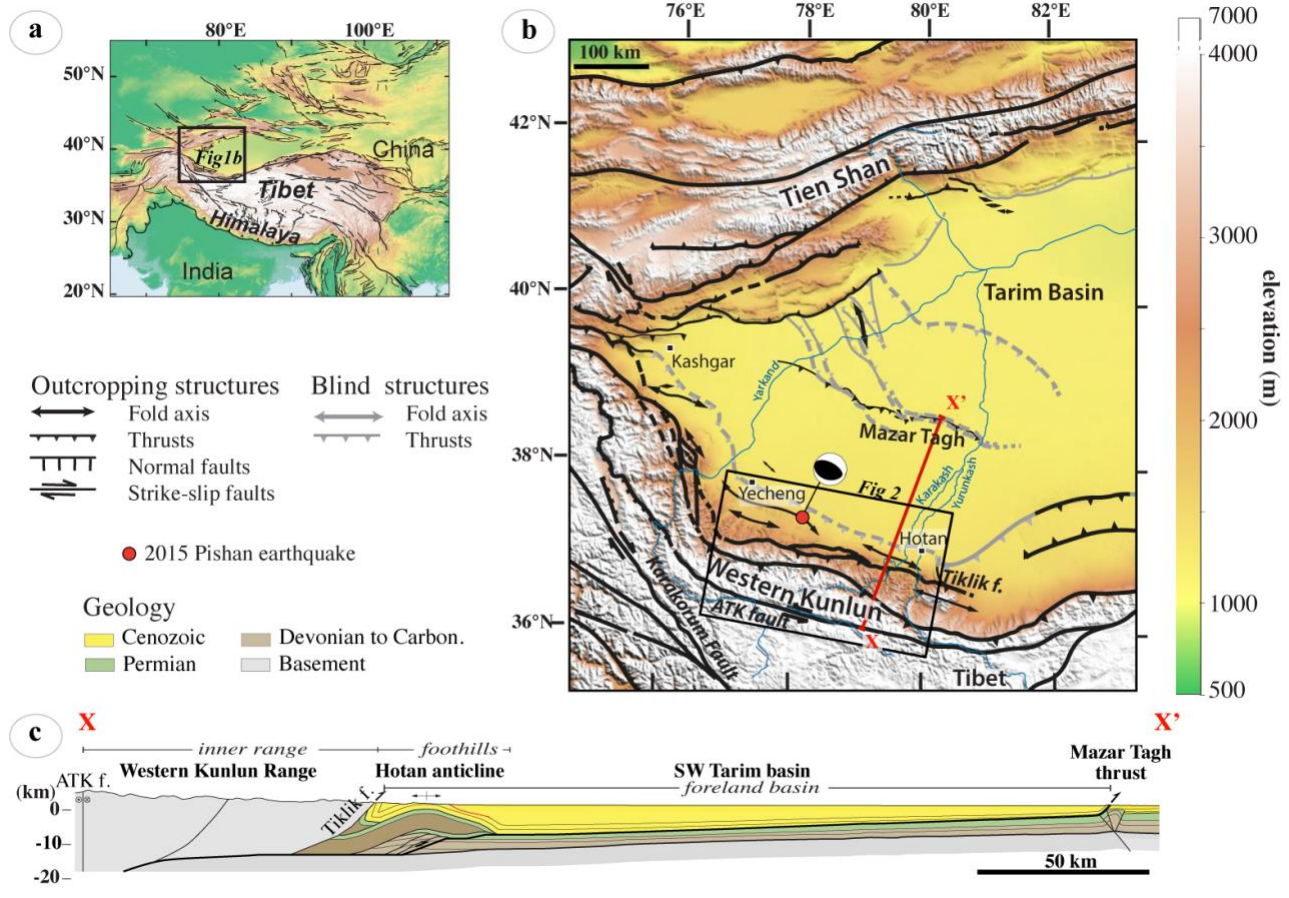
42 1 Introduction

43 The modern Tibetan Plateau results from the collision between India and Eurasia since
44 the early Cenozoic (e.g. [Allegre et al., 1984; Kapp and DeCelles, 2019]). Its building-up and
45 deformation has stood as a great field case to better understand the mechanics of the continental
46 lithosphere (e.g. [England and Houseman, 1989; P. Tapponnier et al., 2001]), as well as
47 earthquake mechanics as it features numerous major active faults and is a place of intense
48 seismic activity. As such, numerous studies have attempted to quantify the kinematics of
49 deformation of major continental faults, with particular attention on major strike-slip faults (e.g.
50 [Lacassin et al., 2004; Meriaux et al., 2004; Phillips and Searle, 2007; Replumaz et al., 2001;
51 Van Der Woerd et al., 2000; P Z Zhang et al., 2007], among many others), but also on the
52 mountain ranges forming the edges of the Plateau, such as the Himalayas to the south (e.g.
53 [Bollinger et al., 2006; Lavé and Avouac, 2000; Long and Robinson, 2021; Yin and Harrison,
54 2000], among many others), the Longmen Shan to the east (e.g. [Hubbard and Shaw, 2009;
55 Pitard et al., 2021; Royden and al, 1997]), and the Qilian Shan to the north-east (e.g. [P
56 Tapponnier et al., 1990; Yin et al., 2008; Zuza et al., 2016]). The 2001 Mw 7.8 Kokoxili (e.g.,
57 [Lasserre et al., 2005; Vallée et al., 2008]), the 2008 Mw 7.9 Wenchuan (e.g., [Liu-Zeng et al.,
58 2009; X Xu et al., 2009]), and the 2015 Mw 7.8 Gorkha (e.g., [Grandin et al., 2015; Hubbard et
59 al., 2016]) earthquakes are some recent examples of major earthquakes rupturing either one of
60 these major continental strike-slip faults or the thrusts that support the mountain ranges bounding
61 the Tibetan Plateau.

62 In contrast, the Western Kunlun Range, at the northwestern edge of the Plateau, has
63 remained much less studied and the first-order tectonic framework still remains to be properly
64 and consistently documented. The Western Kunlun forms a prominent topographic feature to the
65 southwest of the Tarim Basin in its foreland (Figure 1). Modern deformation rates are hardly
66 resolvable from geodesy [Guilbaud et al., 2017; H Wang et al., 2011] across this mountain
67 range. Nonetheless, crustal-scale mass-balance and structural cross-sections, indicate at least ~70
68 km of Cenozoic shortening that were probably mostly accommodated by the blind structures all

69 along the foothills [*Baby et al.*, 2022; *Laborde et al.*, 2019]. Even though available published
70 sections across these structures may not be all self-consistent and therefore directly comparable
71 (e.g. [*Baby et al.*, 2022; *Guilbaud et al.*, 2017; *X Jiang et al.*, 2013; *Laborde et al.*, 2019; *Li et*
72 *al.*, 2016; *Sun et al.*, 2019; *C-Y Wang et al.*, 2013]), lateral variations may exist with a
73 pronounced decrease in Cenozoic crustal shortening to the east of the mountain range, from ~64
74 km down to ~35 km eastward at the longitude of the Hotan anticline [*Baby et al.*, 2022] (Figure
75 1) or even less at the eastern termination of the range, at the transition towards the Altyn Tagh
76 strike-slip range [*Laborde et al.*, 2019]. In terms of recent active tectonics, only a couple
77 attempts at quantifying the kinematics of active deformation across the mountain front from
78 deformed young (Quaternary) geomorphic markers have been carried out since the 2015 Mw 6.4
79 Pishan earthquake [*Guilbaud et al.*, 2017; *J Xu et al.*, 2020]. They indicate a recent shortening
80 rate of ~0.5 to 3.9 mm/yr over the last ~300 kyr, but these data remain limited to the epicentral
81 area of the Pishan earthquake. Further east along the mountain front, significantly lower
82 shortening rates of <0.5 mm/yr were deduced across the Hotan anticline over the last 8-9 Myr
83 from structural cross-sections and growth strata analysis [*Baby et al.*, 2022]. Additional
84 constraints are therefore needed to further enlarge our view on the recent kinematics of
85 shortening all along the foothills of the Western Kunlun, and to better understand the seismic
86 potential of this region [*Guilbaud et al.*, 2017].

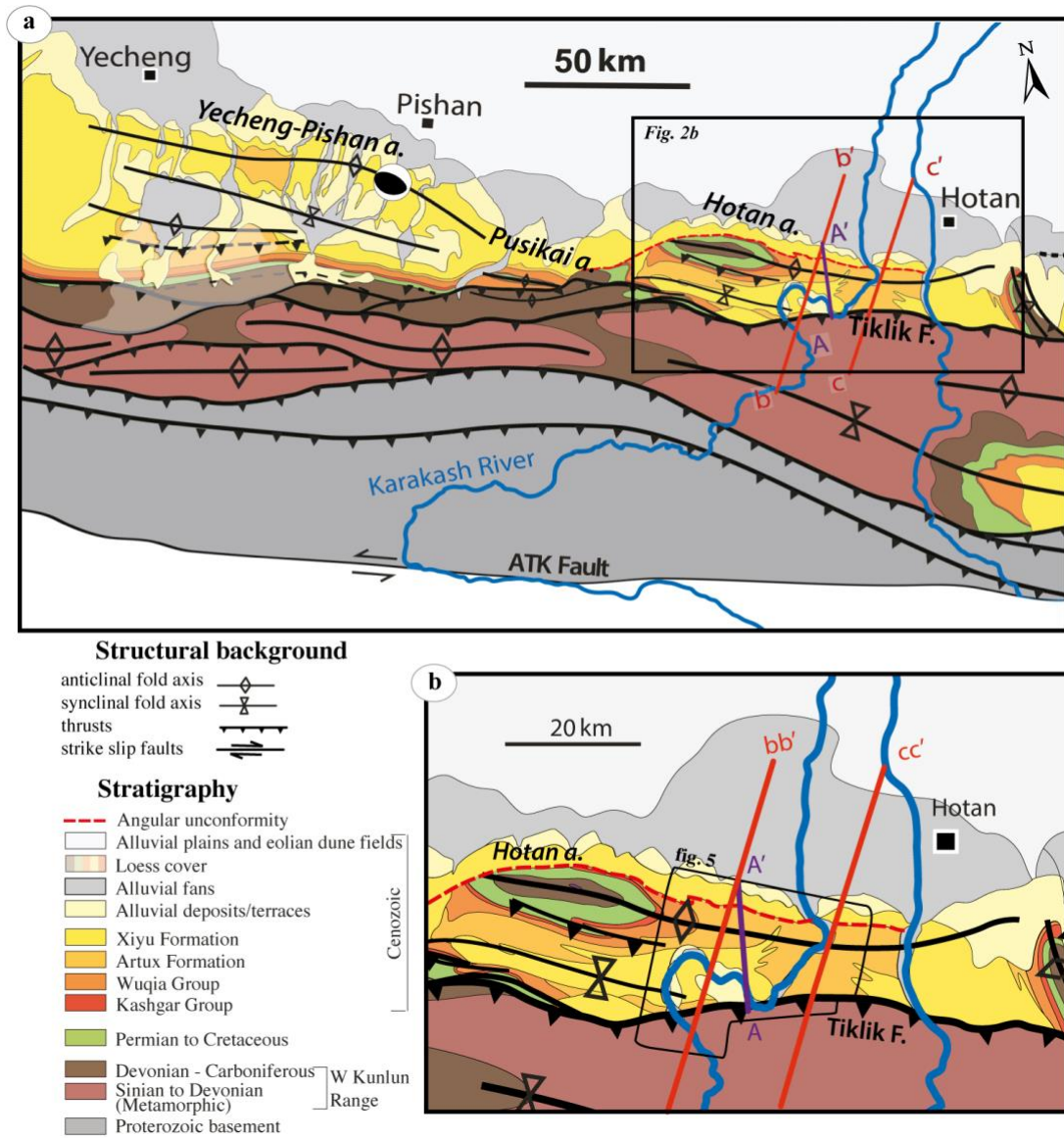
87 Additionally, structural investigations suggest that slip on the blind thrust ramps forming
88 the folds of the Western Kunlun foothills is transferred northward into the Tarim Basin,
89 indicating that the deformation front of the range is located at the Mazar Tagh ridge, i.e., ~150-
90 180 km north from the main topographic mountain front (e.g. [*Baby et al.*, 2022; *Chen et al.*,
91 2022; *Guilbaud et al.*, 2017; *Laborde et al.*, 2019; *Lu et al.*, 2016; *Wittlinger et al.*, 2004] (Figure
92 1c). This extremely large frontal thrust sheet has a remarkably simple geometry, showing little to
93 no internal deformation, above a ~4°S dipping decollement underlying the whole southwestern
94 Tarim Basin. This clearly contrasts with the apparent structural segmentation of the mountain
95 front, as deduced from the surface geology of the various folds forming the foothills (Figure 2).
96 Together with recorded seismicity, deformed recent geomorphic features clearly indicate that the
97 blind ramps forming the mountain front are active [*Ainscoe et al.*, 2017; *Guilbaud et al.*, 2017; *J*
98 *Xu et al.*, 2020; *Y Zhang et al.*, 2023], and by structural and kinematic deduction, that this is also
99 the case for the wide Mazar Tagh thrust sheet [*Guilbaud et al.*, 2017]. Given the dimensions and
100 the structural simplicity of this structure (Figure 1c), potential large continental earthquakes with
101 magnitudes $M > 8$ could be suspected. A better quantification of possible lateral variations in
102 shortening and slip rates on the various blind structures of the mountain front is expected to
103 indirectly help better assess how much slip and slip rate is transmitted forward onto the Mazar
104 Tagh thrust.



105

106 **Figure 1.** Location and structural context of the Western Kunlun Range. Topography is from the
 107 SRTM30 Digital Elevation Model. (a) General location in Asia, in the context of the Indo-
 108 Eurasian collision. Frame locates Figure 1b. (b) Topography and structural context of the
 109 Western Kunlun Range and southwestern Tarim Basin. The Western Kunlun Range overthrusts
 110 the southwestern rim of the Tarim Basin. Structural framework is taken from [Laborde et al.,
 111 2019]. Frame locates the geological map of Figure 2. ATK Fault: Altyn-Tagh Karakash strike-
 112 slip fault; Tiklik F.: Tiklik thrust fault. (c) Geological cross-section across the Western Kunlun
 113 Range, at the level of the Hotan anticline (after [Baby et al., 2022]). Section XX' is located on
 114 Figure 1b. The Western Kunlun Range extends from the Altyn-Tagh Karakash strike-slip fault to
 115 the south, to its deformation front at the Mazar Tagh thrust to the north. The Tiklik fault
 116 separates the inner range from the outer range and southwestern Tarim foreland basin [Laborde
 117 et al., 2019]. The Hotan anticline is one of the structures of the foothills, in the outer range.

118



119

120 **Figure 2.** Geological map of the study area. (a) Geological map of the Western Kunlun inner and
 121 outer ranges (location reported on map of Figure 1b). The inner range is constituted of
 122 Proterozoic to Paleozoic metamorphic basement, and is separated from the outer range by the
 123 Tiklik fault (F.: fault). The outer range and the structure of its foothills encompass Permian to
 124 Cenozoic series, folded in a series of anticlines and synclines along the mountain front, such as
 125 the Yecheng-Pishan, Pusikai or Hotan anticlines (a.: anticline). Structural cross-sections bb' and
 126 cc' (after [Baby et al., 2022]) (Figure 3) are located. ATK Fault: Altyn-Tagh Karakash fault. (b)
 127 Geological map of the Hotan anticline. The Hotan anticline shows a strong lateral variability as it
 128 exhumes and deforms variable stratigraphic levels, from the Cenozoic Xiyu formation to the east
 129 down to the Paleozoic to the west. The northern limb of the anticline is characterized by the
 130 presence of an erosional unconformity at the base of the Xiyu Formation. Section AA', on which
 131 our morphological observations are projected (Figure 6), is also reported. The polygonal frame
 132 locates the geomorphological map of Figure 5.

133

134 Here we propose to further explore these questions on the recent and potentially seismic
135 deformation of the Western Kunlun Range by quantifying the recent slip rate on the blind ramps
136 beneath the Hotan anticline. More precisely, we carry a morpho-tectonic analysis, in which we
137 map, date and analyze the deformation recorded by fluvial terraces along the Karakash River
138 where it flows out of the inner mountain range and crosses the anticline. By extension, this
139 analysis is also applied to the nearby Tiklik thrust where a scarp has been evidenced in the field.
140 From there, we quantify fault slip rates over the last ~250 kyr. These results are then compared
141 to longer-term estimates and variations on these estimates inferred from structural geology across
142 the Hotan anticline. They are also combined with existing data on recent fault slip rates from the
143 Pishan area to complement our view on the kinematics of active deformation all along the
144 Western Kunlun Range and southwestern Tarim Basin.

145 **2 Geological and kinematic context**

146 2.1 Geological setting

147 2.1.1 The Western Kunlun Range

148 The Western Kunlun Range is located along the southwestern edge of the Tarim Basin,
149 and extends over ~700 km from Kashgar to Hotan (Xinjiang, China) (Figure 1). It is divided in
150 two main segments of different orientations: 1) one trending approximately northwest-southeast
151 to the west, between the cities of Kashgar and Yecheng, in structural and stratigraphic continuity
152 with the northeastern Pamir Range, and 2) another segment trending approximately west-east to
153 the east, between the cities of Yecheng and Hotan. Hereafter, we only focus on this latter portion
154 of the Western Kunlun Range, where the deformation front has been described as extending far
155 north into the Tarim Basin (e.g. [*Chen et al.*, 2022; *Guilbaud et al.*, 2017; *Laborde et al.*, 2019;
156 *Lu et al.*, 2016; *Wittlinger et al.*, 2004]).

157 The Western Kunlun from Yecheng to Hotan presents very high elevations in the internal
158 part of the range, with peaks reaching altitudes over 5000-6000 m. Altitudes drop to ~1200 m
159 within the Tarim foreland basin. The Western Kunlun mountain range develops between the
160 Altyn Tagh-Karakash strike-slip fault to the south, and the Mazar Tagh emerging frontal thrust
161 ramp to the north (Figure 1c). North of the Karakash fault, the inner mountain range is composed
162 of imbricates of a Paleo-Proterozoic metamorphic basement, separated by thrust faults proposed
163 to root into a deep crustal decollement [*Laborde et al.*, 2019; *Matte et al.*, 1996; *Wittlinger et al.*,
164 2004]. Further north, the Tiklik fault separates the inner range from the outer range (Figure 1c).
165 In the outer range, the Paleozoic to Cenozoic sedimentary cover of the Tarim Basin is deformed
166 in a fold-and-thrust belt characterized by blind upper-crustal duplexes, as inferred from both
167 seismic profiles and field observations [*Baby et al.*, 2022; *Guilbaud et al.*, 2017; *X Jiang et al.*,
168 2013; *Laborde et al.*, 2019; *Li et al.*, 2016; *Lu et al.*, 2016; *C-Y Wang et al.*, 2013]. The thrusts of
169 these duplex stacks connect an intermediate decollement level located within Upper Cambrian
170 gypsiferous shales to a shallower decollement within the Paleogene evaporitic series at the base
171 of the Cenozoic sediments (e.g. [*Laborde et al.*, 2019]). This shallow decollement reaches the
172 surface ~180 km farther north into the Tarim Basin, at the latitude of the Mazar Tagh emerging
173 ramp, forming this way a uniquely wide frontal thrust sheet with little to no internal deformation
174 (e.g., [*Chen et al.*, 2022; *Guilbaud et al.*, 2017; *Laborde et al.*, 2019; *Lu et al.*, 2016; *Wittlinger*
175 *et al.*, 2004]) (Figure 1c).

176 The metamorphic Protero-Paleozoic basement of the inner range recorded the earlier
177 complex deformation history of the region during the Ordovician and Silurian closure of the
178 Proto-Tethys [Matte et al., 1996; Mattern and Schneider, 2000]. In contrast, the outer range and
179 Tarim Basin did not record this previous deformation episode, and is mostly constituted of a
180 Proterozoic igneous basement overlain by Neoproterozoic to Paleozoic marine calcareous and
181 terrigenous sediments (e.g., [Laborde et al., 2019]). This difference in stratigraphy between the
182 inner and outer ranges suggests that the localization of deformation along the present mountain
183 front is at least partly controlled by inherited Paleozoic structures [Baby et al., 2022; Laborde et
184 al., 2019].

185 Late Paleozoic to Cenozoic marine to continental sedimentary series characterize the
186 sedimentary cover of the outer range (Figure 2). Cenozoic thicknesses reach up to ~9 km nearby
187 Yecheng [Metivier and Gaudemer, 1997], as the sediments produced by the erosion of the
188 Western Kunlun Range and other surrounding mountains have been trapped inside the subsiding
189 endorheic Tarim Basin. The Cenozoic strata are subdivided in four main lithostratigraphic units
190 from early Paleogene up to present (e.g., [Bosboom et al., 2014; Laborde et al., 2019; Wei et al.,
191 2013]): the Kashgar and Wuqia Groups, and the Artux and Xiyu Formations. The Kashgar Group
192 is composed of marine limestones, siltstones, gypsum and marls dating from Paleocene to
193 Eocene. The Wuqia Group, consisting mostly of sandstones, siltstones and claystones, is marking
194 the transition from marine to continental deposits. Then, the Artux Formation, made up of
195 similar lithologies, is marked by the appearance of the first conglomerate layers. Finally, the
196 Xiyu Formation is mostly composed of conglomerates, interbedded with some sandy layers. This
197 upward coarsening of the Cenozoic series is interpreted to be related to the progressive thrusting
198 of the Western Kunlun mountain wedge over the flexural Tarim Basin [Laborde et al., 2019], as
199 suggested in other mountain ranges worldwide [Charreau et al., 2009; Dubille and Lavé, 2015;
200 Simoes and Avouac, 2006].

201 2.1.2 The Hotan anticline and Tiklik thrust

202 The Hotan anticline is one of the folds forming the mountain front of the Western Kunlun
203 Range. It is located to the south-southwest of the city of Hotan (Figure 2). This ~110 km long
204 and ~20 km wide anticline has an overall N110° direction to the west, turning N90° at its eastern
205 termination. To the east, the Hotan anticline is crossed by two large rivers, the Karakash and the
206 Yurunkash rivers (Figure 2).

207 The surface geology of the Hotan anticline reveals significant lateral variations. At its
208 eastern termination, only the upper Cenozoic series outcrops, and a lateral facies transition can
209 be observed from the Artux to the Xiyu formations (Figure 2b). Laterally, the whole Cenozoic to
210 Mesozoic series, down to Paleozoic limestones and metamorphic phyllites, is progressively
211 exhumed from east to west [Baby et al., 2022] (Figure 2b). Along the northern forelimb of the
212 Hotan anticline, the Xiyu Formation is deposited unconformably over the older formations with a
213 basal erosive contact [Baby et al., 2022; X Cheng et al., 2017; X Jiang et al., 2013] (Figure 2b).
214 South of the anticline, the Tiklik fault separates the inner and outer ranges of the Western
215 Kunlun (Figure 2b).

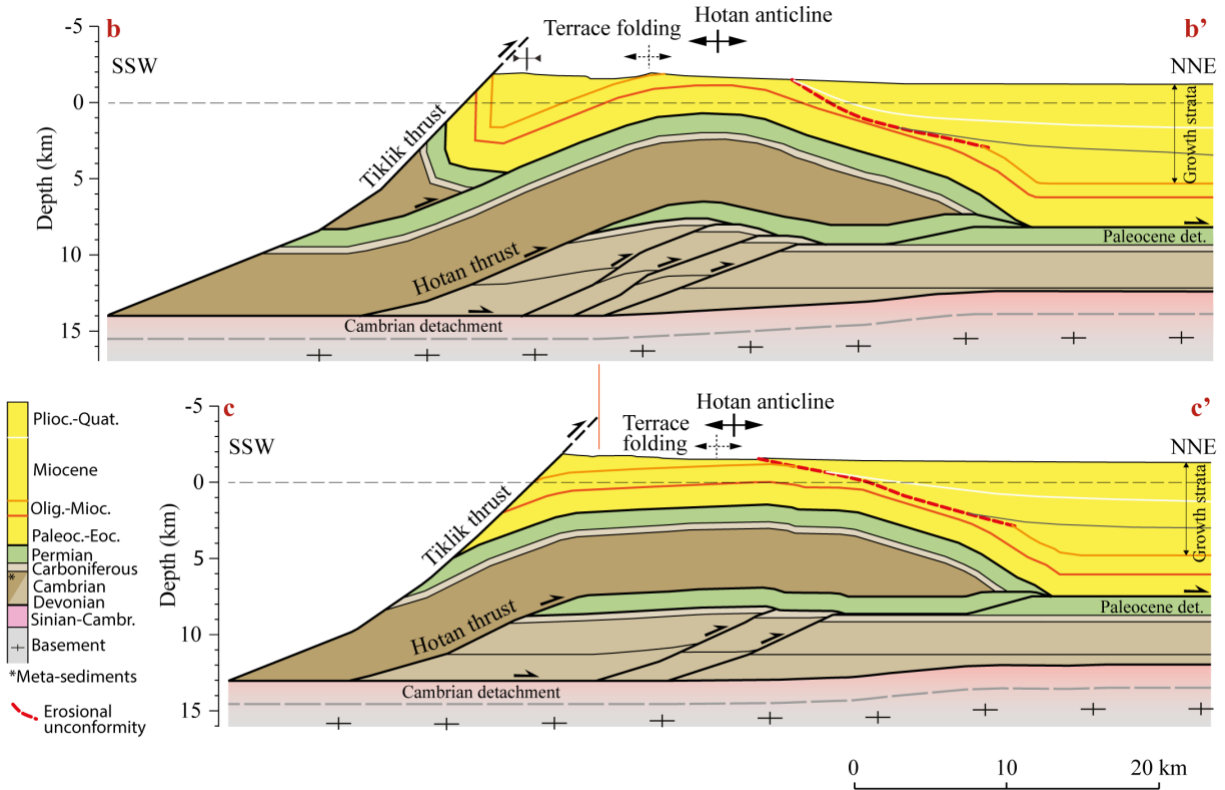
216 Several cross-sections based on the interpretation of seismic reflection profiles and of one
217 borehole have been proposed across the Hotan anticline [Baby et al., 2022; X Cheng et al., 2017;
218 X Jiang et al., 2013]. Of particular interest here, the recent work of [Baby et al., 2022] explores
219 the lateral variations in the sub-surface structure of the anticline, in line with the observed lateral

220 variations in surface geology. They interpret the anticline as resulting from a broad ramp fault-
221 bend anticline over the Hotan thrust, subsequently folded by a footwall duplex. All these
222 structures are blind and form the Hotan thrust system (Figure 3). The Hotan thrust involves
223 Paleozoic phyllites and develops at a probable stratigraphic contrast between the Paleozoic
224 metamorphic series of the Western Kunlun Range and the coeval sedimentary units of the Tarim
225 Basin [Baby et al., 2022; Laborde et al., 2019]. In contrast, the underlying duplex only implies
226 structures neo-formed within the Tarim sedimentary series. These thrusts connect the two
227 regional decollement layers: the Cambrian decollement and the one at the base of the Cenozoic
228 series. The geometry of layers imaged in seismic profiles implies that an intermediate
229 decollement level may be locally present at the base of the Permian [Baby et al., 2022] (Figure
230 3). In our subsequent analysis, we will rely on the sub-surface structure of the Hotan anticline as
231 described in the cross-sections bb' and cc' of [Baby et al., 2022] (Figures 2-3). Locally, at the
232 scale of these two geological sections, the underlying footwall duplex has orientations that differ
233 slightly from the N110 axis of the Hotan anticline (Figure 2). By considering the structural data
234 of [Baby et al., 2022], we find that the frontal duplex ramps locally have a N95°-N100°
235 direction.

236 Total shortening across the Hotan anticline amounts to 35-40 km, and is diversely
237 partitioned along-strike between the Hotan thrust and the footwall duplex [Baby et al., 2022].
238 The footwall duplex mostly developed to the west, uplifting and exhuming the deeper series
239 forming the anticline of the Hotan thrust ramp, and is absent at the eastern extremity of the fold
240 (Figures 2b-3). From the analysis of growth strata preserved in the forelimb of the anticline,
241 [Baby et al., 2022] suggested that deformation started by 16 Ma (or possibly earlier), in line with
242 previous work [X Cheng et al., 2017]. The transition from the basement Hotan thrust ramp to the
243 footwall duplex is recorded in growth layers, and appears diachronous along-strike, with an
244 eastward propagation of the duplex, together with the forward propagation of structures into the
245 Tarim Basin.

246 Thermochronological data indicate that fast exhumation of the basement in the hanging
247 wall of the Tiklik fault occurred sometime during the Late Oligocene and Early Miocene, and
248 has been more moderate since then [X Cheng et al., 2017]. These data record ~4-9 km of
249 exhumation, that may translate into >2-12 km of shortening, depending on how surface uplift and
250 paleo-elevations have evolved [X Cheng et al., 2017]. The recent activity of this thrust remains
251 undocumented.

252



253

254 **Figure 3.** Structural geometry of the Hotan anticline (location of sections on Figure 2). Sections
 255 bb' and cc' are taken from [Baby et al., 2022]. In this study, we only consider the sections that are
 256 closest to our site of investigation along the Karakash River and to the section AA' used for our
 257 morphological and morphotectonic interpretations (Figure 2b). The Hotan anticline results from
 258 a broad ramp anticline over the Hotan thrust, subsequently folded by the ramps of a footwall
 259 duplex. Axis of terrace folding (Figures 5-6) projected on structural sections bb' and cc'.

260

261 **2.2 Kinematics of crustal shortening across the Western Kunlun Range**

262 A total Cenozoic shortening of ~55-100 km across the Western Kunlun Range was
 263 estimated by crustal mass balance budgets [Baby et al., 2022; Laborde et al., 2019]. Considering
 264 a probable age of ~20-23 Ma for the initiation of the Cenozoic reactivation of the mountain
 265 range, as suggested from thermochronology (e.g., [Cao et al., 2015; X Cheng et al., 2017; Sobel
 266 and Dumitru, 1997; E Wang et al., 2003]) and the dated coarsening of the sediments within the
 267 foreland basin [Blayney et al., 2019; X-D Jiang and Li, 2014; Metivier and Gaudemer, 1997;
 268 Zheng et al., 2015], a long-term average shortening rate of ~2.4-5 mm/yr is obtained. This value
 269 is comparable, in its lower bound though, with the long-term (over several million years)
 270 shortening rates deduced from the structure and kinematics of some of the structures of the outer
 271 range, such as the Yecheng-Pishan [Guilbaud et al., 2017; Sun et al., 2019; J Xu et al., 2020] or
 272 the Hotan [Baby et al., 2022] anticlines, in line with the idea that most of the Cenozoic crustal
 273 shortening across the Western Kunlun Range has been absorbed in the foothills [Baby et al.,
 274 2022]. Based on the analysis of growth strata, [Baby et al., 2022] suggested that shortening rates
 275 across the Hotan anticline may have significantly decreased from ~5 mm/yr to <0.5 mm/yr by

276 ~8-9 Ma. Such deceleration has not yet been documented or corroborated elsewhere along the
277 mountain range.

278 Data on recent kinematics (i.e., over the last 100s kyrs), at the shorter time scale of
279 several seismic cycles, are too scarce along the mountain front to be compared to long-term
280 estimates. The 2015 Mw 6.4 Pishan earthquake ruptured the most frontal blind ramp of the
281 Yecheng-Pishan anticline [Li et al., 2016; Lu et al., 2016] and attracted some attention on this
282 particular structure where several flights of fluvial terraces are preserved and record recent
283 tectonic uplift [Ainscoe et al., 2017; Guilbaud et al., 2017; J Xu et al., 2020]. Due to the disputed
284 ages of some of these terraces, the analysis of these geomorphic markers of deformation suggests
285 fault slip rates ranging from 0.5 to 3.9 mm/yr over the last ~300-500 kyr [Guilbaud et al., 2017;
286 J Xu et al., 2020; Y Zhang et al., 2023]. As any slip on the blind ramps of the mountain front is to
287 be transferred forward onto the Paleocene decollement all the way northward to the Mazar Tagh
288 emerging ramp (Figure 1c), such slip rate values are also expected to be found there. A tentative
289 estimate of >0.9 mm/yr from an uplifted (but poorly dated) terrace has been proposed on the
290 Mazar Tagh [Pan et al., 2010].

291 Existing estimates of recent shortening and fault slip rates are therefore limited to draw a
292 clear and complete picture of the recent kinematics of shortening across the Western Kunlun
293 mountain front. More specifically, these data are absent in the particular case of the Hotan
294 anticline even though they would allow for exploring the idea of a recent deceleration of
295 shortening rates as proposed by [Baby et al., 2022], or for investigating possible lateral variations
296 along the mountain front, by comparison with the results obtained in Pishan [Guilbaud et al.,
297 2017; J Xu et al., 2020].

298 **3 Geomorphology of the Karakash River across the Hotan anticline and Tiklik thrust**

299 To investigate the recent kinematics of uplift and shortening of the Hotan anticline and
300 Tiklik fault, we conduct a geomorphological analysis of a site located along the Karakash River,
301 where it crosses these various structures. The Karakash River is one of the largest rivers flowing
302 out of the Western Kunlun Range, with a total drainage area of ~19000 km² extending well
303 beyond the Altyn-Tagh / Karakash fault at the rear of the mountain range (Figure 1b).

304 In our study area, the river crosses the Tiklik fault and flows out of the high-relief inner
305 range, meanders and follows its course behind and then across the Hotan anticline (Figure 2). It
306 then flows north into the Tarim Basin and forms a relatively large (>500 km²) active alluvial fan.
307 From the inner range to the alluvial fan, several fluvial terraces were incised and preserved
308 (Figure 4).

309 **3.1 Data and Methods**

310 **3.1.1 Field and satellite observations**

311 We map fluvial terraces within our study site using high-resolution (0.5 m) Pléiades
312 images. A high-resolution digital elevation model (DEM) is computed using the open-source
313 software suite MicMac developed by the Institut de physique du globe de Paris (IPGP, France)
314 and the Institut national de l'information géographique et forestière (IGN, France) [Rosu et al.,

315 2015; *Rupnik et al.*, 2017]. In some areas of the images where very smooth, bright and white
316 surfaces are located, the resultant DEM is not well-defined due to poor image correlation.

317 Therefore, these areas are excluded from our analysis. The DEM is subsequently down-
318 sampled to a 2m-resolution. To complement these data and cover a larger zone, a 30m-resolution
319 ALOS World 3D DEM together with the satellite images from the Google Earth database are
320 also used. Observations on satellite images and DEMs are completed by field observations
321 during four different field surveys in 2003, 2005, 2017 and 2019, with a cumulated on-site stay
322 of ~7-9 days due to limited field permits. A topographic profile across the Tiklik fault scarp is
323 measured in the field using a theodolite.

324 3.1.2 Mapping and correlation of fluvial terraces

325 Fluvial terraces are abandoned former river beds and as such, they appear in the field as
326 relatively planar surfaces, delimited by risers, above and along the present-day river bed (Figure
327 4). The relative altitudes of the various terraces reflect their relative ages, the higher terraces
328 being the older ones. Some of the mapped surfaces are complex as they may correspond to more
329 than one fluvial terrace. Indeed, lateral erosion by the Karakash River reveals that in some cases,
330 the loess or colluvial cover hides several generations of imbricated erosional fluvial strath
331 terraces (Figure 4a). This is the case in particular for our highest, and therefore oldest identified
332 surfaces. To avoid these issues, we will mostly rely on the lowest, and therefore youngest
333 surfaces. Because, terrace surfaces may be locally covered by colluvium from higher levels
334 (Figures 4 b-e), we will mostly consider in our quantitative analysis their altitudes nearest to their
335 riser, i.e. at distance from other higher terraces and closest to the present river.

336 The terrace record is discontinuous along the river, so that terrace patches are correlated
337 with each other using local across- and along-topographic profiles. Such correlation is
338 subsequently verified or completed with chronological constraints obtained from the dating of
339 some terrace patches (Section 4). Because of the presence of a fault scarp along the trace of the
340 Tiklik fault (Figure 4e), we consider the possibility that the terrace record may be disrupted and
341 therefore discontinuous across this fault. Therefore, correlation of terraces and their attributed
342 relative ages is first performed separately on either side of the fault, and then combined together
343 in our final map using the obtained absolute ages of terraces.

344 3.1.3 Computing terrace incision

345 Terrace elevations are extracted from available DEMs to compute terrace incision above
346 the present-day Karakash River. Incision is calculated along the long-distance profile of the river
347 course (Figure S1 in supplementary information). Because the river meanders and changes its
348 main flowing direction as it is deflected eastwards behind the Hotan anticline, the river course is
349 simplified by considering four segments of average similar flowing direction to compute terrace
350 incision (Figure 5, inset). When defining these main river segments, we verify that the average
351 long river profile is relatively simple and linear in each one of these segments (Figure S1). Some
352 terrace patches are preserved within major river meanders, i.e., at the intersection between two
353 main river segments. In this case, we compute the incision of these terraces along each one of the
354 two considered river segments separately, and deduce from there the range of their incision when
355 combining both results. In these cases, we find that incision is determined with a ~25-30 m

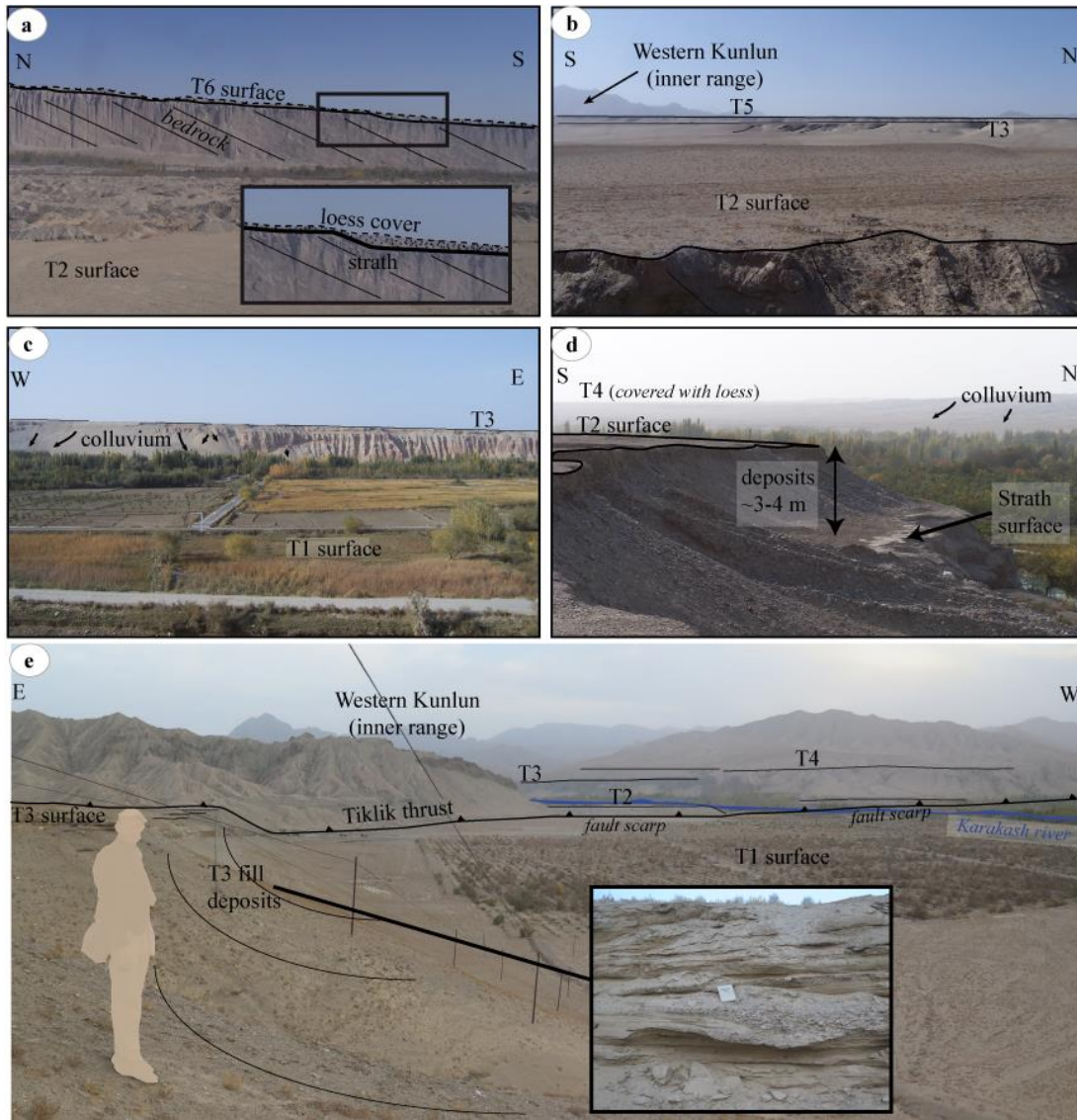
356 maximum uncertainty corresponding to the difference in river elevation on either side of these
357 major meanders. Once calculated, incision is projected perpendicular to the axis of tectonic
358 structures to finally derive incremental uplift.

359 3.2 Geomorphological response to active tectonics along the Karakash River

360 The Tiklik thrust has a clear morphological expression where the Karakash River flows
361 out of the inner range, with a ~14 m high scarp disrupting fluvial terraces (Figure 4e), testifying
362 that the fault is active. Downstream from this scarp, the river first flows northward, is then
363 deflected southeastward along the backlimb of the Hotan anticline, and finally flows once again
364 northward to cross the anticline at its eastern termination (Figure 5). Major meanders are indeed
365 located over the backlimb of the Hotan anticline (Figures 5 and 6a). This change in the overall
366 river flow direction and its deflection towards the fold lateral termination is interpreted as related
367 to the uplift of the Hotan anticline, as similarly observed in other contexts (e.g., [Benedetti et al.,
368 2000; Burbank et al., 1996; Lavé and Avouac, 2000]), most probably in line with the eastward
369 propagation of folding over time [Baby et al., 2022].

370 Six levels of fluvial terraces are identified and mapped within our study area, labeled T1
371 to T6 from the lowermost and youngest to the topmost and oldest mapped surface (Figure 5).
372 The three uppermost terrace levels (T4 to T6) are scarcely preserved and form highly
373 discontinuous patches, but the remnants of the three other lower and younger levels (T1 to T3)
374 are abundant and relatively well-preserved along the river course. From field observations, we
375 cannot discard the possibility that other secondary generations of terrace levels are hidden below
376 the highest and oldest surfaces, as they are covered with loess and colluvium (Figures 4a and c).
377 Our field surveys also indicate that terraces are mostly strath terraces, with < 10 m of fluvial
378 deposits (cobbles, pebbles and/or sand) above the basal erosional strath surface (Figure 4d),
379 indicating ongoing active incision. Only locally, immediately downstream of the Tiklik fault
380 trace and where a syncline is described in the underlying geology behind the Hotan anticline, fill
381 terraces are suspected in the field. There, fluvial deposits are observed over > 10 m, with no clear
382 observation of the substratum beneath these sediments (Figures 4e and 5). To the north-east of
383 the investigated site, the Karakash River crosses the Hotan anticline and aggrades to form a large
384 alluvial fan (Figures 5 and 6a). These observations clearly indicate that the pattern of river
385 incision and aggradation follows the structural pattern of the Tiklik thrust and of the Hotan
386 anticline. Accordingly, active tectonics is found to exert a first-order control on the morphology
387 of our study site.

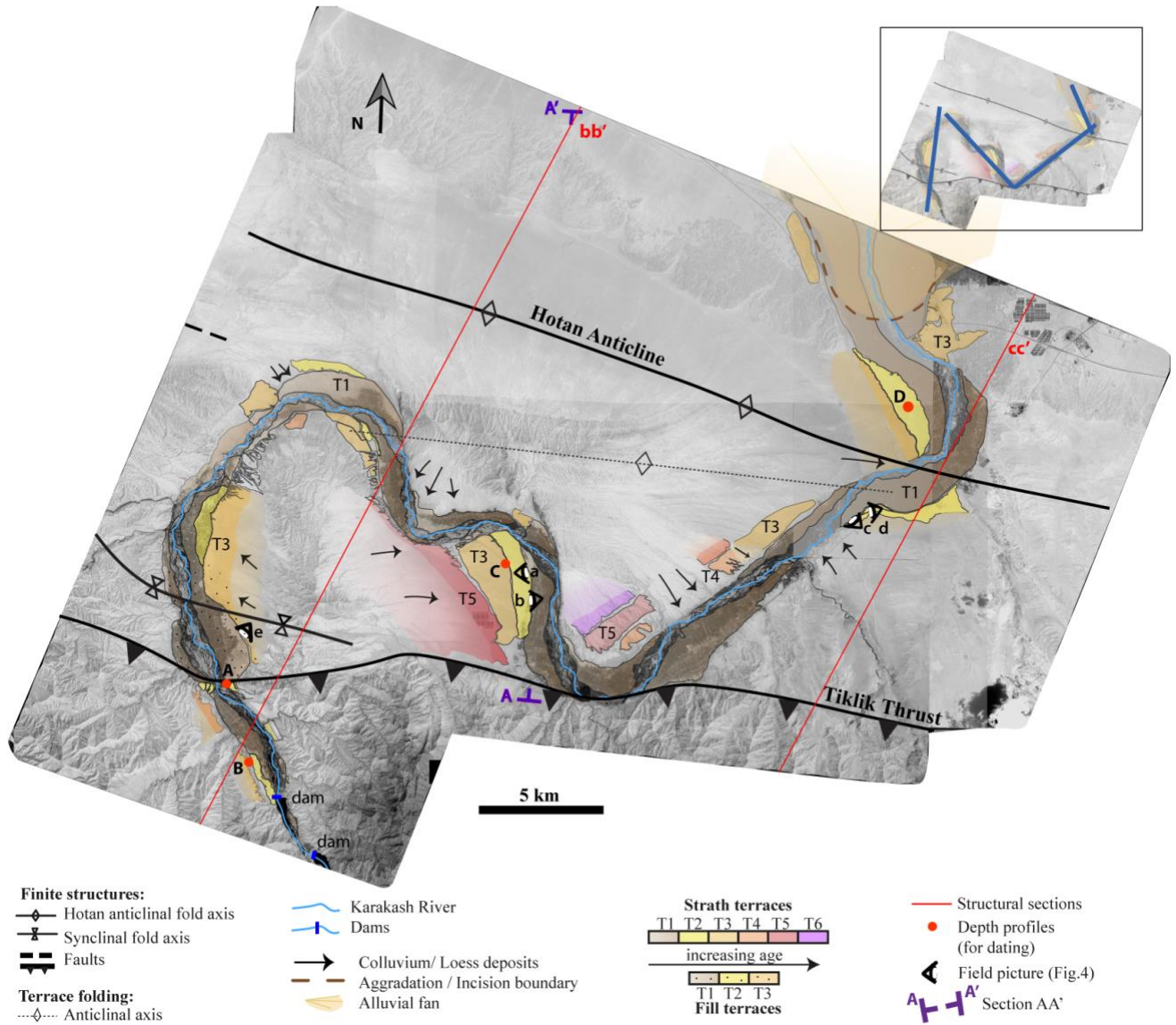
388 The incision pattern is relatively well-defined and continuous throughout the whole area
389 for terraces T1 to T3, and more scattered for the higher mapped levels. For a comparison with
390 sub-surface geology, terrace incision is projected perpendicular to structures (Figure 6b). As the
391 orientations of the Hotan anticline (as defined from surface geology, Figure 2) and footwall
392 duplex are slightly different, we tested both directions. When a projection perpendicular to the
393 Hotan anticline is considered (i.e., along a N24° axis), two zones of higher incision appear,
394 which are not directly comparable to underlying structures. Alternatively, we find a better
395 correlation between the pattern of terrace incision and the underlying geology when projecting
396 all data perpendicular to the direction of the footwall duplex ramps. More specifically, the best
397 correlation between all terrace data and the underlying geology (Figures 6b-c) is found when



398

399 **Figure 4.** Field pictures of the investigated fluvial terraces. These pictures are located on the
 400 morphological map of Figure 5 (numbering of pictures as in the map). (a) Planar surface of T6,
 401 on the left bank of the Karakash River. Even though continuous, this surface possibly hides, in
 402 the details, several generations of incisional terraces, as illustrated by the step in the basal strath
 403 and overlying fluvial deposits subsequently covered by loess (detail picture). (b) Planar surfaces
 404 of terraces T2, T3 and T5, on the right bank of the Karakash River. (c) Planar surfaces of terraces
 405 T1 and T3. T1 presently mostly hosts agricultural activities. Laterally, the terrace riser of T3 is
 406 hidden and covered by colluvial deposits (left side of the picture). (d) T2 terrace, illustrating the
 407 strath erosional basal surface, covered by ~3-4 m of fluvial deposits. (e) View over the Tiklik
 408 thrust fault and scarp, where the Karakash River flows out of the Western Kunlun inner range.
 409 The ~14 m high thrust scarp separates terrace T2 upstream from terrace T1 downstream.
 410 Downstream of the Tiklik thrust, terraces are identified as aggradational from their thick fill
 411 deposits, as illustrated here for T3. Inset: detail view on the fill deposits of T3, with field
 412 notebook for scale.

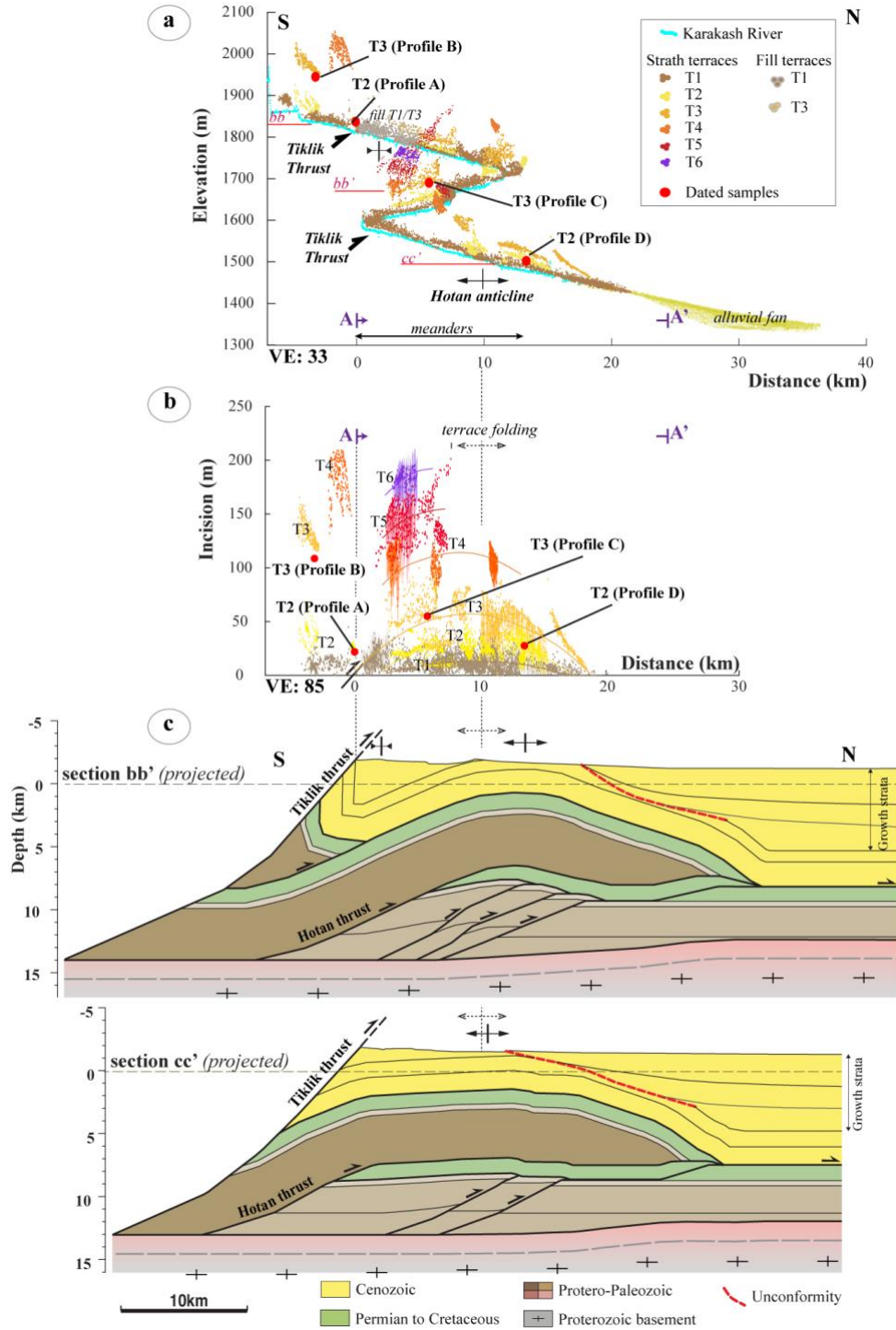
413



414

415 **Figure 5.** Map of the investigated site (location of map on Figure 2b). Various terrace levels are
 416 mapped along the Karakash River where it flows out of the Western Kunlun inner range,
 417 crossing the Tiklik thrust and the Hotan anticline. They are identified in the field as fill terraces
 418 immediately downstream of where the river crosses the Tiklik thrust (Field picture e, Figure 4e),
 419 and strath terraces elsewhere (Field picture d, Figure 4d). Four depth profiles within some of
 420 these terrace levels (sites A to D) have been sampled to date them using in-situ produced
 421 cosmogenic isotopes (Figure 7). Terrace incision is calculated all along the river course, by
 422 discretizing the river into four main flowing segments (upper right inset). A combined long-
 423 distance profile of the river and terraces can be found in Figure S1 (supplementary information).
 424 Incision is then projected on profile AA' (Figure 6), which is perpendicular to the direction of
 425 footwall duplex ramps and of terrace folding.

426



427

428 **Figure 6.** Projected terrace elevation and incision compared to the structural geometry of the
 429 Hotan anticline. All the data are projected parallel to profile AA' (Figures 2 and 5), and the origin
 430 of horizontal distances is set to where the Karakash River flows out of the Western Kunlun inner
 431 range and crosses the Tiklik thrust. VE: vertical exaggeration. (a) Projected river and terrace
 432 elevations. Main river meanders locate at km 0 and 10 along the section and are illustrated by the
 433 alternation of southward and northward changes in river flow, accompanied by a significant

434 (*Figure 6*) change in river elevation. Terrace levels are color-coded as in the map of *Figure 5*.
 435 Location of sampled depth profiles (A to D) are reported together with correlated terrace levels.
 436 Various structural features encountered along the Karakash River course are reported along the
 437 projected river profile, together with the intersections of the river course with structural sections
 438 bb' and cc'. (b) Projected terrace incision for documented strath terraces. Terraces are color-
 439 coded as in *Figure 6a*. Terrace incision is determined all along the longitudinal river profile (inset
 440 of *Figure 5*), before being projected. Transparent colored vertical lines indicate the possible
 441 range of incision values where the river meanders. Even though not continuous in particular for
 442 terrace levels higher than T3, the terrace record indicates folding, with a broad zone of higher
 443 incision by km 10. Terrace folding appears spatially correlated with thrusting over the frontal
 444 ramps of the footwall duplex (*Figure 6c*). (c) Structural sections bb' and cc' from [*Baby et al.*,
 445 2022] (*Figure 3*), projected onto section AA'.

446

447 projecting all data along section AA' oriented N05° (in purple on *Figure 5*), slightly
 448 oblique to the direction of the frontal duplex ramps. This minor obliquity may be related to data
 449 uncertainties, or rather relate to the combined eastward and forward propagation of duplexes
 450 over time, as this direction coincides with that obtained when kinematically relating the
 451 penultimate duplex ramp along section bb' to the most frontal one along section cc'. When
 452 considering all data projected along AA', incision is highest and delineates a broad ~10-15 km
 453 wide terrace anticline in the area of the major river meanders, along the backlimb of the main
 454 Hotan anticline (*Figure 6*). There is therefore a slight spatial discrepancy between the finite
 455 structure seen from surface geology and the morphological expression of active deformation. In
 456 fact, surface geology is mostly dominated by thrusting over the Hotan thrust along the Karakash
 457 River (*Figure 3*), whereas incision spatially coincides better with the location of the underlying
 458 duplex (*Figure 6*), indicating that these blind footwall ramps have been active over the time span
 459 represented by the terrace record.

460 Slip on the duplex ramps is to be transferred northward and upward to the frontal smaller
 461 blind ramp north of the anticline, and then into the Paleogene decollement. The frontal small
 462 blind ramp, connecting an intermediate decollement within the Permian and the base of the
 463 Cenozoic series, spatially coincides with the active alluvial fan of the Karakash River (*Figures 5*
 464 and *6a*) where growth layers have been described in seismic profiles [*Baby et al.*, 2022] (*Figures*
 465 *3* and *6c*). This further supports the recent activity of the blind structures in the footwall of the
 466 Hotan thrust.

467 **4 Ages of terraces along the Karakash River**

468 4.1 Data and Methods

469 4.1.1 Approach

470 Cosmogenic nuclides, such as ^{10}Be or ^{26}Al , are produced in situ in quartz minerals due to
 471 the exposure of near surface rocks to cosmic radiation, and undergo radioactive decay with time
 472 (e.g., [*Lal*, 1991]). Because of the attenuation of cosmic radiation across the rock mass, the
 473 production of nuclides decreases exponentially with depth. The nuclide concentration in surface
 474 rocks can be expressed as:

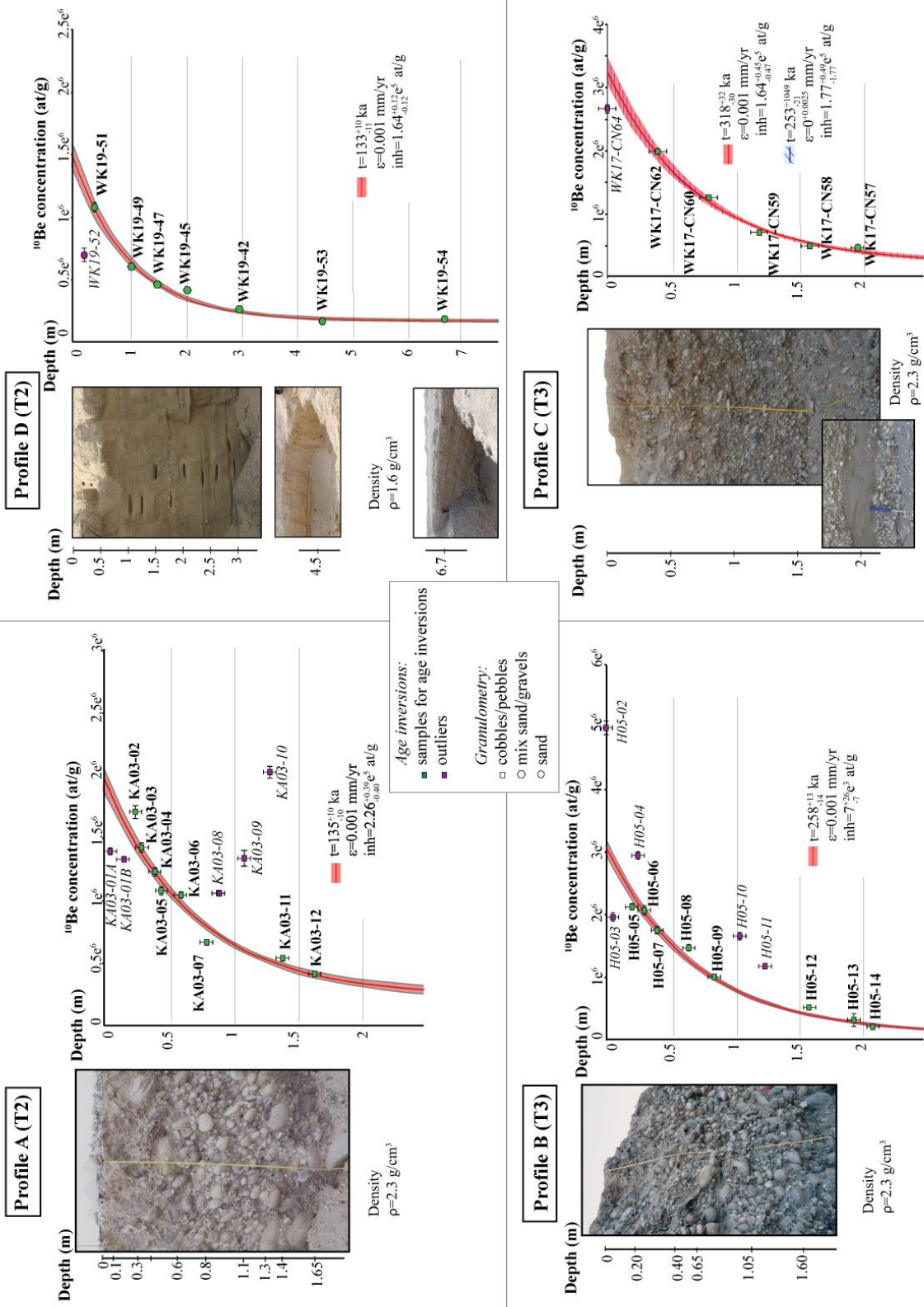
$$475 \quad C(z, t) = C(z, t_0) \cdot e^{-\lambda t} + \sum_i \frac{P_i}{\lambda + \mu_i \varepsilon} e^{-\mu_i z} (1 - e^{-(\lambda + \mu_i \varepsilon)t}) \quad (1)$$

476 where C is the concentration of atoms (atoms.g^{-1}), z is the depth (cm), t is the exposure time (yr),
 477 λ is the radioactive decay of the considered nuclide (yr^{-1}), P is the production rate at the surface
 478 ($\text{atoms.g}^{-1}.\text{yr}^{-1}$), μ is the attenuation coefficient (cm^{-1}), ε is the surface erosion rate (cm.yr^{-1}), and
 479 i represents the production pathways for neutrons and muons (e.g., [Braucher et al., 2003; Brown
 480 et al., 1992; Gosse and Phillips, 2001; Lal, 1991]). $C(z,t_0)$ corresponds to the initial
 481 concentration of atoms when exposure initiated, inherited from a prior exposure or transport
 482 history. Attenuation μ_i is determined as the ratio ρ/A_i , where ρ is the density of the material
 483 (g.cm^{-3}) and A_i the attenuation length (g.cm^{-2}) of the considered cosmic particle i (neutron or
 484 muon). Provided that parameters related to attenuation, production and radioactive decay are
 485 known, modeling the nuclide concentrations along a depth profile of an exposed surface is
 486 commonly used to derive the exposure age of the surface, as well as the average erosion rate and
 487 the initial inheritance (e.g., [Anderson et al., 1996; Guilbaud et al., 2017; Hidy et al., 2010;
 488 Repka et al., 1997; Yang et al., 2019]).

489 4.1.2 Collected samples and analyses

490 To date the abandonment of terraces using in situ produced ^{10}Be , we collected samples
 491 along depth profiles in four of the mapped fluvial terraces. During field campaigns in 2003 and
 492 2005, we sampled two terraces located in the hanging wall of the Tiklik thrust (Figure 5).
 493 Sampled terrace deposits are here formed of boulders and cobbles. At site A, a ~ 1.65 m deep
 494 profile was sampled into the man-made refreshed riser of the lower T2 terrace (Figure 7). At site
 495 B, a ~ 2.1 m deep profile was sampled along the steep riser wall of the highest terrace (terrace
 496 T3) (Figure 7). For this site, the top sample of the profile is constituted of amalgamated surface
 497 pebbles from the terrace surface (such surface sample could not be collected on the surface of T2
 498 at site A due to reworking). Further downstream, two terraces located over the limbs of the
 499 Hotan anticline have also been sampled in 2017 and 2019 for dating. At site C, a profile was
 500 realized into terrace T3, in the backlimb of the Hotan anticline. Amalgamated pebbles from 5
 501 cm-thick horizons and a sand sample were collected along a ~ 2 m-deep recently man-made
 502 excavation (Figure 7). The top sample is constituted of amalgamated surface pebbles. Finally, at
 503 site D, a fourth depth profile was collected along a recently man-made excavation into terrace
 504 T2, in the downstream portion of the river (Figure 5). This profile is ~ 3.3 m deep, and 5 cm-thick
 505 sand and amalgamated gravel horizons were collected from bottom to top (Figure 7). This profile
 506 is completed by two other deeper samples, accessible from nearby excavations (within ~ 100 m
 507 from the main profile) into the same terrace, at 4.5 and 6.7 m below the terrace surface (Figure
 508 7).

509 Chemical analyses of samples from profiles A to C were prepared at the Cosmogenic
 510 Isotope Laboratory of the University of Strasbourg (ITES, Strasbourg, France) and at the Centre
 511 Européen de Recherche et d'Enseignement des Géosciences de l'Environnement (CEREGE, Aix-
 512 en-Provence, France). Samples from profile D were prepared at the Purdue Rare Isotope
 513 Measurement Laboratory (PRIME Lab, Purdue University, USA). Most Accelerator Mass
 514 Spectrometer (AMS) measurements were performed at the French ASTER facility (CEREGE,
 515 Aix-en-Provence, France). Only two measurements were conducted at the Lawrence Livermore
 516 National Laboratory (LLNL, Livermore, USA). We obtained 40 measurements of ^{10}Be
 517 concentrations and 17 measurements of ^{26}Al concentrations. All analytical results are reported in
 518 Table 1. Further analytical details can be found in supplementary material (Text S1 and Table
 519 S1).



521 **Figure 7** ^{10}Be concentration with depth along the four depth profiles A to D. A field picture is
 522 provided for each depth profile, at scale with the graphs illustrating the measured concentrations
 523 with depth. Multiple pictures indicate that some of the samples were taken laterally from the
 524 main depth profile (Profiles D and C). The densities used for the age inversions have been
 525 determined from the average grain size (cobbles, gravels, sand) found all along the profiles. They
 526 are reported for each depth profile. The grain size of the samples is also reported by various
 527 symbols. Samples considered for age inversions are reported in green, and outliers in purple.
 528 Best-fit solutions of the age inversions when erosion is fixed (Model 1) are plotted (red line, with
 529 red confidence interval), and the associated parameters are indicated. In the case of profile C, the
 530 best-fit solution in the case that erosion is also inverted (Model 2, blue line and hatched interval)
 531 is also indicated. Analytical details for all samples in Table 1 and model parameters for age
 532 inversion in Table 2.

533

534 4.1.3 Age determination

535 The exposure time (taken for the time of abandonment of the terraces) is determined from
 536 the modeling of the ^{10}Be concentrations along depth profiles A to D, following equation (1) and
 537 using the parameters detailed below. Even though the ^{26}Al concentration has been measured for
 538 17 of our 40 samples, we do not model it. The available ^{26}Al concentrations are rather used to
 539 question the reliability of ^{10}Be measurements whenever the $^{26}\text{Al}/^{10}\text{Be}$ ratio significantly departs
 540 from the theoretical production ratio of 6.61 [Braucher et al., 2011].

541 The radioactive decay λ of ^{10}Be is set to $5.13 \cdot 10^{-7} \text{ yr}^{-1}$ (^{10}Be half-time of $1.35 \cdot 10^6$ yrs)
 542 [Chmeleff et al., 2010; Korschinek et al., 2010]. The production rate P varies as a function of
 543 time and space, depending on the solar activity and on the Earth's magnetic field. Here, we
 544 however use the time constant scaling scheme “St” [Lal, 1991; Stone, 2000] found to be
 545 satisfactory [Balco and Rovey, 2008; Borchers et al., 2016], with a production rate of 4.01 ± 0.33
 546 $\text{atoms} \cdot \text{g}^{-1} \cdot \text{yr}^{-1}$ at sea level and at high latitudes to calculate the expected local ^{10}Be production
 547 rate at our study site. To derive the slow and fast muonic production rates, we followed
 548 [Braucher et al., 2011] and scaled the rates with atmospheric pressure ignoring the latitudinal
 549 effect. Here, a production rate ranging between 11.71 and $14.94 \text{ atoms} \cdot \text{g}^{-1} \cdot \text{yr}^{-1}$ is determined,
 550 assuming that the present river base level represents the altitude of terraces when folding
 551 initiated. We use attenuation lengths A of $160 \text{ g} \cdot \text{cm}^{-2}$, $1500 \text{ g} \cdot \text{cm}^{-2}$, and $4320 \text{ g} \cdot \text{cm}^{-2}$ for neutron,
 552 and slow and fast muons, respectively [Balco, 2017; Braucher et al., 2011; Marrero et al., 2016].
 553 The density ρ of the sampled terrace profile is set to $2.3 \text{ g} \cdot \text{cm}^{-3}$ when these terraces are
 554 constituted of boulders and pebbles (Sites A, B and C), and to $1.6 \text{ g} \cdot \text{cm}^{-3}$ when they are
 555 constituted of finer material such as sand or gravels (Site D). We first set the erosion rate ε to 10^{-6}
 556 $\text{m} \cdot \text{yr}^{-1}$ (Model 1), following the value obtained from terraces known to be in erosional steady-
 557 state in Pishan (see supplementary material in [Guilbaud et al., 2017]). Erosion is also tested as a
 558 free parameter (Model 2). Further details on how these various parameters were determined are
 559 provided in supplementary material (Text S2).

560 Equation (1) is first inverted for time (exposure age t) and inheritance ($C(z, t_0)$) by
 561 minimizing the misfit between a theoretical curve and the observed data, following a χ^2 criterion:

$$562 \chi^2 = \sum_{i=1}^n \left(\frac{c_i - C(z_i, t)}{\sigma_{c_i}} \right)^2 \quad (2)$$

Sample	Rock type	Thickness (cm)	Depth (cm)	10Be (1e5 at/g qtz)	26Al (1e5 at/g qtz)	26Al/10Be
Profile A (36.88743°N, 79.41021°E, 1837 m) - Terrace T2						
KA03-1A	cobble	5	5	13.966 ± 0.395		
KA03-1B	cobble	5	15	13.343 ± 0.342		
KA03-2	cobble	5	25	17.102 ± 0.589	94.256 ± 3.094	5.51
KA03-3	cobble	5	30	14.274 ± 0.475	78.749 ± 2.605	5.52
KA03-4	cobble	5	40	12.331 ± 0.428		
KA03-5	cobble	5	45	10.822 ± 0.413	70.137 ± 2.547	6.48
KA03-6	cobble	5	60	10.482 ± 0.338		
KA03-7	cobble	5	80	6.69 ± 0.198	39.459 ± 1.467	5.90
KA03-8	cobble	5	90	10.611 ± 0.336		
KA03-9	cobble	5	110	13.437 ± 0.71	44.86 ± 2.193	3.34
KA03-10	cobble	5	130	20.295 ± 0.641		
KA03-11	cobble	5	140	5.447 ± 0.211	36.035 ± 1.51	6.62
KA03-12	cobble	5	165	4.155 ± 0.135		
Profile D (36.99381°N, 79.72292°E, 1500 m) - Terrace T2						
WK19-52	sand	5	20	7.294 ± 0.59		
WK19-51	sand	6	39	11.301 ± 0.417		
WK19-49	gravels	5	105	6.292 ± 0.233		
WK19-47	sand/gravels	5	153	4.826 ± 0.208		
WK19-45	sand /gravels	5	207	4.332 ± 0.23		
WK19-42	sand /gravels	4	300	2.713 ± 0.126		
WK19-53	sand	5	450	1.722 ± 0.064		
WK19-54	sand	5	670	1.897 ± 0.074		
Profile B (36.85861°N, 79.42104°E, 1944 m) - Terrace T3						
H5-2	amalgamated qtz vein pieces	2	0	49.892 ± 1.14		
H5-3	cobble	6	5	19.665 ± 0.707		
H5-4	cobble	6	20	21.284 ± 0.544		
H5-5	cobble	6	25	29.43 ± 0.831	143.33 ± 4.273	4.87
H5-6	cobble	6	30	20.725 ± 0.749	157.137 ± 5.305	7.58
H5-7	cobble	6	40	17.517 ± 0.628	109.146 ± 4.003	6.23
H5-8	cobble	9	65	14.717 ± 0.513		
H5-9	cobble	6	85	9.972 ± 0.442	50.795 ± 2.234	5.09
H5-10	cobble	6	105	16.575 ± 0.635		
H5-11	cobble	6	125	11.713 ± 0.38	59.222 ± 2.416	5.06
H5-12	cobble	7	160	5.161 ± 0.18		
H5-13	cobble	6	195	3.098 ± 0.123	17.307 ± 0.744	5.59
H5-14	cobble	10	210	2.123 ± 0.074		
Profile C (36.93355°N, 79.53800°E, 1691 m) - Terrace T3						
WK17CN-64	amalgamated qtz vein pieces	1	0	26.61 ± 0.679	172.539 ± 4.223	6.48
WK17CN-62	amalgamated pebbles(n≥50)	7	40	19.809 ± 0.52	121.646 ± 3.292	6.14
WK17CN-60	amalgamated pebbles(n≥50)	7	80	12.502 ± 0.35	74.505 ± 1.997	5.96
WK17CN-59	amalgamated pebbles(n≥50)	7	120	6.931 ± 0.222	46.64 ± 1.696	6.73
WK17CN-58	amalgamated pebbles(n≥50)	7	160	4.791 ± 0.143		
WK17CN-57	sand	5	198	4.54 ± 0.159	18.396 ± 1.524	4.05

564 **Table 1** Sampling and isotopic data for all collected samples for cosmogenic analysis. Depth
 565 profiles labeled A to D and interpreted terrace levels are indicated and located with geographic
 566 coordinates and altitudes at the surface. All the samples taken along these profiles are reported
 567 with their depth along the profile, their granulometry or sample characteristics (qtz: quartz; n:
 568 number of amalgamated samples). Sampled thickness is integrated vertically along the profile.
 569 Complete analytical details are provided in supplementary material (Table S1).

570

571 where C_i are the concentrations measured at depth z_i with an uncertainty σ_{C_i} , and $C(z_i, t)$ is the
 572 predicted concentration at the same depths using equation (1). Acceptable values for exposure
 573 time and inheritance are defined according to various possible confidence intervals: (1) as
 574 obeying the condition defined by: $\chi^2 < \chi^2_{min} + 1$ (e.g., [Bevington and Robinson, 2003;
 575 Braucher et al., 2009; Charreau et al., 2017; Saint-Carlier et al., 2016; Yang et al., 2019]; and
 576 (2) within a 1σ - or (3) 2σ -confidence interval.

577 For comparison, these inversions for exposure age and for inheritance in the case of a
 578 fixed erosion rate of 10^{-6} m.yr⁻¹ (Model 1) are completed by another set of inversions where
 579 erosion rates are set free and inverted (Model 2). Ages from Model 1 will be preferred, as
 580 erosion is in this case well-defined from a nearby site by steady-state terraces, and Model 2 will
 581 be mostly informative. All results are presented in Table 2. The variation of the χ^2 criterion as a
 582 function of the various tested parameters, as well as the possible trade-offs between inverted
 583 parameters are illustrated in the case of both Models 1 and 2 in supplementary material (Text S2,
 584 Figures S2-S9).

585 4.2 Ages of fluvial terraces

586 4.2.1 Age inversions

587 All samples are included in the modeling, except for (1) surface amalgams or samples
 588 from stratigraphic levels observed to be perturbed along the depth profiles, such as most often
 589 close to the terrace surface (cryoturbation, fissures, etc); and (2) samples whose ¹⁰Be content
 590 appears too high relative to the overall visually expected exponential decay with depth. Even
 591 though disregarded in the inversion of the depth profiles, these samples defined as outliers from
 592 stratigraphic, sampling or analytical arguments, are reported with our results for comparison
 593 (Figure 7).

594 ²⁶Al/¹⁰Be ratios are not used *a priori* to define outliers in the modeling of ¹⁰Be
 595 concentrations along depth profiles, but are rather taken *a posteriori* as informative of possible
 596 complexities. Indeed, these ratios are known for only one third of the samples, and any departure
 597 from the expected production ratio could be related to a complexity impacting either the ¹⁰Be or
 598 the ²⁶Al concentration. Overall, most ²⁶Al/¹⁰Be ratios are relatively close to the expected
 599 production ratio (Table 1), signifying simple exposure histories. Only 2 samples show low ratios
 600 (<3.5). One is a clear outlier along profile A (sample KA03-9) because of a ¹⁰Be concentration
 601 too high in view of the overall trend at depth (Figure 7); the other is a sand lens along profile C
 602 (sample WK17-CN57), when all other samples of this profile are amalgamated pebbles with
 603 consistent ²⁶Al/¹⁰Be ratios. Because the ¹⁰Be concentration of the sand lens is visually within the
 604 expected exponential decrease with depth, we have no valuable reason to exclude it from our

605 modeling. All other samples considered as outliers from field or analytical reasons have
 606 consistent $^{26}\text{Al}/^{10}\text{Be}$ ratios.

607

Sampled profiles (Interpreted terrace level)	Parameters	Explored Range	Model 1 - fixed erosion rate							
			Results Model 1							
			best fit	interval 1 σ		interval 2 σ		interval $\chi^2_{\min}+1$		
	min	max	min	max	min	max				
A (T2)	Age (kyr)	80 - 600	135	129	141	125	145	131	139	
	Erosion (mm/yr)	fixed to 0.001	0.001	-	-	-	-	-	-	
	Inheritance (at/g)	10 000 - 300 000	226 000	202 000	250 000	186 000	265 000	211 000	241 000	
B (T3)	Age (kyr)	100 - 600	258	250	266	244	271	253	264	
	Erosion (mm/yr)	fixed to 0.001	0.001	-	-	-	-	-	-	
	Inheritance (at/g)	0 - 300 000	7 000	0	23 000	0	33 000	0	17 000	
C (T3)	Age (kyr)	100 - 600	318	299	337	288	350	306	330	
	Erosion (mm/yr)	fixed to 0.001	0.001	-	-	-	-	-	-	
	Inheritance (at/g)	0 - 300 000	164 000	136 000	192 000	117 000	209 000	146 000	182 000	
C (T3)	Age (kyr)	100 - 600	without common outliers and deep sand sample				344	374		
	Erosion (mm/yr)	fixed to 0.001	0.001	336	383	322	399	-	-	
	Inheritance (at/g)	0 - 300 000	66 000	26 000	105 000	0	130 000	40 000	92 000	
D (T2)	Age (kyr)	80 - 600	133	127	139	122	143	129	137	
	Erosion (mm/yr)	fixed to 0.001	0.001	-	-	-	-	-	-	
	Inheritance (at/g)	10 000 - 300 000	164 000	157 000	171 000	152 000	176 000	159 000	168 000	
Sampled profiles (Interpreted terrace level)	Parameters	Explored Range	Model 2 - free erosion rate							
			Results Model 2							
			best fit	interval 1 σ		interval 2 σ		interval $\chi^2_{\min}+1$		
	min	max	min	max	min	max				
A (T2)	Age (kyr)	80 - 3000	123	116	1 353	113	3000	120	325	
	Erosion (mm/yr)	0 - 0.01	0	0	0.0055	0	0.006	0	0.005	
	Inheritance (at/g)	0 - 300 000	228 000	64 000	260 000	0	274 000	183 000	245 000	
B (T3)	Age (kyr)	100 - 600	258	209	292	205	297	231	264	
	Erosion (mm/yr)	0 - 0.01	0.001	0	0.0015	0	0.0015	0.0005	0.001	
	Inheritance (at/g)	0 - 300 000	7 000	0	54 000	0	64 000	0	30 000	
C (T3)	Age (kyr)	100 - 1500	253	238	1100	232	1302	246	486	
	Erosion (mm/yr)	0 - 0.01	0	0	0.0025	0	0.0025	0	0.002	
	Inheritance (at/g)	0 - 300 000	177 000	10 000	210 000	0	226 000	128 000	194 000	
C (T3)	Age (kyr)	100 - 1500	without common outliers and deep sand sample				269	639		
	Erosion (mm/yr)	0 - 0.01	0.001	260	664	253	1060	0	0.002	
	Inheritance (at/g)	0 - 300 000	85 000	0	132 000	0	154 000	0	110 000	
D (T2)	Age (kyr)	80 - 2000	124	117	621	114	1 661	121	289	
	Erosion (mm/yr)	0 - 0.01	0	0	0.008	0	0.0085	0	0.007	
	Inheritance (at/g)	0 - 300 000	166 000	115 000	175 000	81 000	179 000	144 000	170 000	

608

609 **Table 2** Ages obtained from the inversion of depth profiles A to D (Figures 5 and 7). Interpreted
 610 terrace levels are reported for all depth profiles. In Model 1, we invert for age and inheritance,
 611 and consider a fixed erosion rate of 0.001 mm/yr. In the case of Model 2, the erosion rate is also
 612 inverted. The best-fitting solutions are indicated with the minimum and maximum limits of
 613 various confidence intervals (1 σ , 2 σ and $\chi^2_{\min}+1$). Two model runs are reported for the inversion
 614 of profile C. The first one considers all samples, except outliers disregarded following commonly
 615 defined criteria for all profiles (here: surface amalgamated samples). The second one also
 616 excludes the deep sampled sand layer. See text and supplementary material for further details.

617

618 Inversion results are reported in Table 2 and represented in Figure 7 for depth profiles A
 619 to D. We here favor age results derived from models with erosion rates fixed to a well-
 620 established regional value (Model 1), and conservatively consider 2 σ confidence intervals. An
 621 age of 135 ± 10 ka is found for the abandonment of terrace T2 from profile A, and $258 +13/-14$
 622 ka for that of T3 from profile B, upstream and south of the Tiklik thrust. Downstream,

623 abandonment of terrace T2 is dated to $133 \pm 10/-11$ ka from profile D, and that of terrace T3 to
 624 $318 \pm 32/-30$ ka from profile C. Inheritance varies from 0-33 000 atoms.g⁻¹ (Profile B, T3) to 186
 625 000-265 000 atoms.g⁻¹ (Profile A, T2) and remains overall limited (up to ~10% of the
 626 concentration at the surface). The variability in inheritance does not seem to relate simply to
 627 terrace age, distance from the range front, or granulometry (sand, pebbles or cobbles) as
 628 sometimes observed (e.g., [Belmont et al., 2007; Carretier et al., 2009; Lupker et al., 2017]). In
 629 models where erosion is inverted (Model 2), similar best-fit ages are obtained, but the range of
 630 acceptable ages is most often asymmetrically widened toward older values, in line with lower
 631 inheritance and/or higher erosion rate (Figures S6-S9 in supplementary material).

632 4.2.2 Interpreted ages

633 The results from the age inversions performed on the various profiles on either side of the
 634 Tiklik fault are consistent with the correlation established from their morphological analysis
 635 (Section 3). The inversion of depth profiles A and D from T2 terrace remnants converges to
 636 remarkably consistent age intervals of 135 ± 10 ka and $133 \pm 10/-11$ ka, respectively (Model 1,
 637 Table 2, Figure 7). These ages are slightly older than those retrieved for terraces labelled T3
 638 ~130 km further west in Pishan [Guilbaud et al., 2017], but remain compatible with the known
 639 regional climatic evolution, in particular with wetter regional periods as documented from the
 640 Kesang's Cave in the northern Tian Shan [H Cheng et al., 2012]. Even though climatic forcing
 641 has been most probably similar at this regional scale, the upstream drainage basins of the
 642 Karakash and Pishan rivers are dissimilar, being significantly more extensive and glaciated in its
 643 upstream portions in the case of the Karakash River. We propose that these differences explain
 644 the slightly different results in the timing of river aggradation and incision, and from there in the
 645 timing of terrace abandonment between these two sites, following the findings of [Malatesta et
 646 al., 2017] on the variable responses of rivers to a same regional climatic forcing during glacial-
 647 interglacial periods. A common age interval of 125-143 ka will be hereafter considered for all
 648 terraces T2 along the Karakash River (Table 3).

649 The situation is not as simple in the case of the inversion of depth profiles B and C into
 650 terrace remnants of T3, with ages of $258 \pm 13/-14$ and $318 \pm 32/-30$ ka, respectively (Model 1,
 651 Table 2, Figure 7). The age of $258 \pm 13/-14$ ka obtained from profile B fits into the same
 652 reasoning as that previously exposed for profiles A and D into terraces T2: this age is slightly
 653 older than that derived for terraces labelled T4 in Pishan, but is in line with the known regional
 654 climatic evolution. However, such is not the case for profile C, with a slightly older and
 655 inconsistent age (given uncertainties) when compared to that from profile B, questioning the
 656 possibility that the sampled terrace remnant might be part of an older T4 level. This is even more
 657 pronounced when the deepest sand sample WK17-CN57 is not considered in the inversions,
 658 because of its different granulometry or its anomalous ²⁶Al/¹⁰Be ratio (Table 2). However, when
 659 erosion is inverted (Model 2), the best-fit solution favors a non-realistic null erosion and ages
 660 compatible with the idea that profile C was dug into terrace level T3 (Table 2). Results obtained
 661 for profile C rely on the inversion of a relatively smaller set of samples (5 samples instead of 7-8
 662 samples for the other profiles; Figure 7), so that results may be highly dependent on the
 663 consideration/exclusion of any single one of them. This is in particular critical at shallow depths
 664 (<1m), where the exponential decay of ¹⁰Be concentrations is here only defined by 2 unique
 665 samples constituted of pebbles and cobbles, and therefore with a possible variability in their
 666 concentrations, as observed for shallow samples along Profiles A and B of similar granulometry.
 667 As a result, the solution to profile C drawn by these fewer samples is clearly not unique, as both

668 solutions issued from Models 1 and 2 provide the same fit to the data, despite different resultant
 669 ages, as long as the other parameters are adjusted (Figure 7). Given this, we have considered the
 670 possibility that profile C was in fact sampled into an older T4 terrace level. However, we were
 671 not able in this case to find a satisfactory solution for terrace correlation, combining relative ages
 672 from mapping and absolute ages from dating. Given all these considerations, we favor the idea
 673 that profile C is part of a T3 terrace. Because of limited sampling or because of a locally lower
 674 erosion rate, its age could not be well-defined by following the same approach as that of the
 675 other profiles. From this, and providing the age solution for profile B (Model 1) as well as one of
 676 the solutions for profile C (Model 2), we will hereafter attribute and consider the common age
 677 interval of 244-271 ka for terraces T3 along the Karakash River (Table 3).

678

Terrace level	Interpreted ages (ka)	Base level changes (m)	Sinuosity	Maximum excess incision (sinuosity)	Uplift (m) - average (<i>extended</i>)	Fault slip (m) - average (<i>extended</i>)
<i>Present river</i>	-	-	2.14	-	-	-
T1	30-60	12-30 (9-36)	2.14	5.45	16-63 (16-77)	34-192 (33-248)
T2	125-143	50-71 (37-86)	2.08	15.7	79-120 (64-120)	169-367 (131-389)
T3	244-271	97-136 (73-163)	2.03	28.24	146-224 (116-242)	311-689 (239-784)

679

680 **Table 3** Characteristics attributed, measured or calculated for the various terrace levels.
 681 Interpreted ages cover the time span common to the 2σ age intervals retrieved from the inversion
 682 of depth profiles for each terrace level T2 and T3, and is attributed to T1 by comparison to the
 683 known regional climatic evolution. Base level changes are calculated by considering these age
 684 intervals together with a sedimentation rate of 0.4-0.6 mm/yr. Sinuosity of present-day and
 685 paleo-river valleys is determined from Figures 5 and 10. Maximum excess incision related to
 686 sinuosity variations is measured upstream of our study area in the case of T2 and T3, but behind
 687 the Hotan anticline in the case of T1 (Figure 10). Uplift is retrieved where it is highest above the
 688 blind footwall ramps. It is then translated into fault slip using equation (4). For uplift and slip,
 689 most probable values are indicated (average), together with values if larger uncertainties are
 690 considered (*extended*), corresponding to continuous and dashed boxes in Figures 12 and 13,
 691 respectively.

692

693 Given the results and interpretations proposed for terraces T2 and T3, and following the
 694 idea that terrace abandonment relates to the regional climatic evolution as proposed in Pishan
 695 [Guilbaud et al., 2017], we propose that terrace T1 has a probable age of 30-60 ka (Table 3). The
 696 proposed time span is here purposely conservatively large to account for the fact that we have no
 697 absolute tight age constraints on T1. The reasoning is more difficult to apply with confidence to
 698 terrace levels older than T3: in our subsequent analysis, we will therefore mostly rely on terrace
 699 levels T1 to T3, which are the best and most continuously preserved along the Karakash River
 700 (Figures 5 and 6), as well as those with direct or indirect chronological constraints (Tables 2 and
 701 3).

702 5 Kinematics of deformation

703 5.1 General approach

704 Because the overall geomorphology of the study site is primarily controlled by active
705 tectonics, we expect terrace incision to picture approximately tectonic uplift cumulated since
706 terrace abandonment.

707 Where terraces are offset by a fault, the vertical fault throw can be easily determined by
708 comparing the position (altitude or incision) of the disrupted markers on either side of it. Over a
709 much larger spatial scale or when the position of the terraces (or equivalent former riverbeds) is
710 not directly observable in the fault footwall, a complementary approach is followed. Following
711 [Lavé and Avouac, 2000], the pattern of incremental uplift U since terrace abandonment can be
712 quantified from terrace incision I from:

$$713 \quad U = I + D - \Delta I \quad (3)$$

714 where D is the river base level change (positive in the case of sedimentation), and ΔI is the
715 excess incision related to changes in river sinuosity or gradient (positive in the case of an excess
716 of incision) (Figure 8). Uplift is here given relative to the fault footwall (structural uplift).
717 Variations in lithospheric subsidence across the investigated fault systems are neglected given
718 the length scale of these structures relative to that of the whole flexural basin (Figure 1c) and the
719 flexural rigidity of the Tarim Block [Lyon-Caen and Molnar, 1984].

720 Once structural uplift is determined locally across a fault scarp or over a much larger
721 spatial scale, the associated slip on the fault can be assessed using a kinematic model of faulting
722 and folding. In the case of fault-bend folding with bedding-plane slip, as derived here from
723 existing structural constraints on the Hotan anticline [Baby et al., 2022; X Cheng et al., 2017; X
724 Jiang et al., 2013] (Figure 3), it follows that:

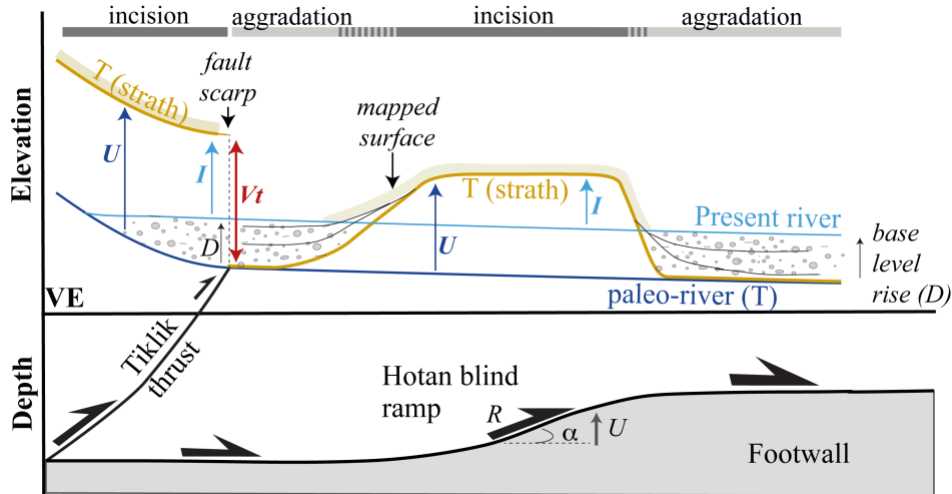
$$725 \quad U = R \cdot \sin \alpha \quad (4)$$

726 where R is the incremental slip on the fault plane since terrace abandonment, and α the
727 underlying local fault dip angle reported to the surface [Bernard et al., 2007; Lavé and Avouac,
728 2000] (Figure 8). This approach is applicable to any terrace record, even though spatially
729 discontinuous, as long as the geometry of the underlying active fault is known. Finally, the fault
730 slip rate is derived by comparing incremental slip and terrace ages.

731 5.2 Kinematics of deformation across the Hotan anticline

732 5.2.1 Quantifying incremental uplift from terrace incision

733 To quantify the recent kinematics of the faults underlying the Hotan anticline, we first
734 derive incremental structural uplift from equation (3) (Figure 8) by correcting terrace incision for
735 changes in river base level, gradient and sinuosity.



736

737 **Figure 8** General principle of our morphotectonic approach (after [Lavé and Avouac, 2000]).
 738 The case of a terrace T record is taken as an example. Once the paleo-riverbed corresponding to
 739 T (dark blue line) is abandoned, it becomes a geomorphic marker (orange line) that is
 740 progressively deformed because of uplift U above the underlying faults (equation 4). This
 741 geomorphic marker is incised (I) by the active riverbed (light blue line), or is buried by
 742 aggradation as the river base level rises over time (D), depending on the local balance between
 743 uplift and sedimentation. As such the present-day river stands as a reference line. The deformed
 744 former riverbed (orange line) is an erosive surface where uplift and incision prevail (strath), but
 745 is a depositional surface where aggradation occurs. Because aggradation may be time-
 746 transgressive [Weldon, 1986], the aggradational mapped surface in geometrical continuity with
 747 the strath terrace T may be younger. Uplift can be determined by comparing the terrace record T
 748 to the geometry of the corresponding former riverbed, and is given here in a structural reference
 749 frame relative to the fixed fault footwall. Uplift can be approximated from terrace incision I , as
 750 long as it is corrected for base level rise D (equation 3). Across the Tiklik thrust, the vertical
 751 throw V_t since the abandonment of terrace T is here the sum of the incision I of the terrace in the
 752 thrust hanging wall and of the river base level rise D in the thrust footwall. Here the
 753 representation of river incision is simplified and does not integrate non-tectonic incision related
 754 to river sinuosity or gradient changes over time. Note that elevation (and the corresponding
 755 terrace record) is vertically exaggerated (VE) compared to the geometry of structures at depth.

756

757 5.2.1.1 River base level

758 Within our study site, the base level of the Karakash River is the endorheic Tarim
 759 foreland basin, immediately north of the Hotan anticline. We do not know of any data on the
 760 sedimentation rate in the basin at the time scale of our terrace record, encompassing probable
 761 climatically-induced variations during glacial and interglacial periods. To overcome this
 762 difficulty, we calculate a long-term sedimentation rate of $\sim 0.5 \text{ mm.yr}^{-1}$ over the last 5 Myr from
 763 the $\sim 2.5 \text{ km}$ thick Plio-Quaternary sediments immediately downstream of the river outlet within
 764 the Tarim foreland (Figures 3 and 6). This value should however be considered as a long-term
 765 minimum as we do not have the data needed to properly decompact the considered sediment
 766 thickness.

767 Estimating the uncertainties on this value is not straightforward, and we arbitrarily
768 consider a wide possible range of $\sim 0.4\text{-}0.6\text{ mm.yr}^{-1}$ for the sedimentation rate at the base level of
769 the Karakash River, with the idea of encompassing the various unknowns on this estimate.
770 Possible time variations in the sedimentation rate at the several 100s kyr time scale of the terrace
771 record should be smoothed out and the large uncertainty (i.e., $\pm 20\%$) stands in line with possible
772 average porosity loss during sediment compaction in the Tarim (see supplementary text S2 in
773 [Laborde et al., 2019]). The considered range of sedimentation rates is overall consistent with
774 the rates proposed in the Pishan area $\sim 150\text{ km}$ further west [Guilbaud et al., 2017], and with
775 previous published local values in the region using decompacted well data [Metivier and
776 Gaudemer, 1997].

777 Over a time interval of $\sim 250\text{ kyr}$, as considered here, the correction for sedimentation at
778 the base level may rise up to $\sim 125\text{ m}$, a substantial value when compared to the maximum $\sim 75\text{ m}$
779 of incision of terrace T3 downstream of the Tiklik thrust (Figure 6). The base level correction to
780 be considered for each terrace is calculated and reported in Table 3.

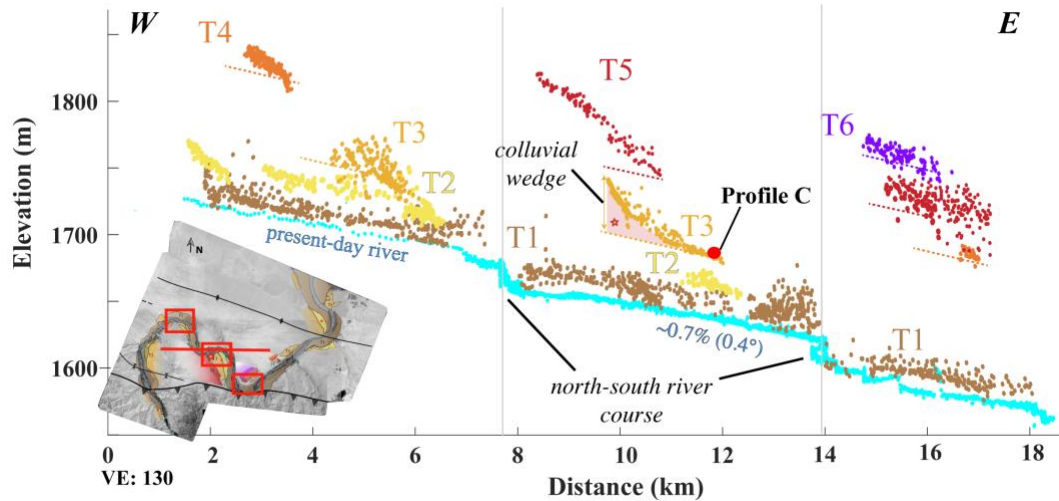
781 5.2.1.2 River gradient

782 The present-day gradient of the Karakash River is of 0.7% . A change in river gradient
783 over time is expressed by: (1) an excess of incision upstream and sedimentation downstream in
784 the case of a decreasing gradient; or (2) an excess of incision downstream and sedimentation
785 upstream for an increasing gradient [Lavé and Avouac, 2000]. At the scale of the Hotan anticline,
786 we do not have any evidence for such broad variations, independent of tectonics, in the incision
787 pattern (Figure 6).

788 To further tentatively track variations in river gradient, we analyse terrace profiles along
789 portions of the river course that are approximately parallel to the axis of the duplex ramps (i.e.,
790 perpendicular to section AA') with the idea that they are expected to have endured
791 approximately constant uplift. Residual slopes in such profiles (relative to present river gradient
792 and/or in between various terrace profiles), if existent, could be partly related to changes in river
793 gradient over time [Poisson and Avouac, 2004; Simoes et al., 2014].

794 We here consider the $\sim 15\text{-}20\text{ km}$ long portion of the Karakash River flowing
795 southeastward along the backlimb of the Hotan anticline (Figure 9). There, three smaller sections
796 of the river flow approximately eastward, separated by others flowing approximately southward.
797 The three eastward sections are located at different distances across strike (i.e., along section
798 AA'), so that structural uplift may be approximated as constant within each one of them. Within
799 each one of these sections, the profiles of terraces T1 to T2 are observed to be parallel to the
800 present-day river when projected along an east-west axis perpendicular to AA' (red line in the
801 inset of Figure 9), with no particular residual slope that could be related to a change in river
802 gradient. This is also mostly the case for T3, except in the central river segment where a residual
803 slope (relative to the present-day river) is observed. We verified that this slope is in fact an
804 artefact as the highest T3 points sample here the top of a colluvial wedge along a T5-T3 riser
805 (Figure 9). In the case of higher and older terraces, a residual slope is observed, which could be
806 partly due to surface processes (in particular in the central segment) or be real. Given that our
807 analysis will rely only on terraces T1 to T3, we hereafter neglect possible changes in river
808 gradient since T3 abandonment.

809



810

811 **Figure 9** River and terraces along the backlimb of the Hotan anticline. The red frames in the
 812 lower left inset locate three segments where the river locally flows in a direction close to that of
 813 terrace folding (Figure 5). Within each of these three segments, terrace incision is therefore
 814 expected to have operated under a similar uplift rate. These segments are separated by portions
 815 of the river flowing in a north-south direction. River and terrace data are projected parallel to the
 816 axis of terrace folding (red line in inset map). Dashed colored lines report the local river gradient
 817 at the altitude of the terraces. A slight apparent excess gradient is observed for T3 in the case of
 818 the central river segment, but this is an artefact as these T3 points sample a colluvial wedge
 819 along a T5-T3 riser near Profile C (colored area in graph and small red star on the inset map).
 820 Overall, terraces T1 to T3 appear relatively parallel to the present-day river within each one of
 821 the east-west flowing segments, indicating that there has been no major change in river gradient
 822 since T3 abandonment.

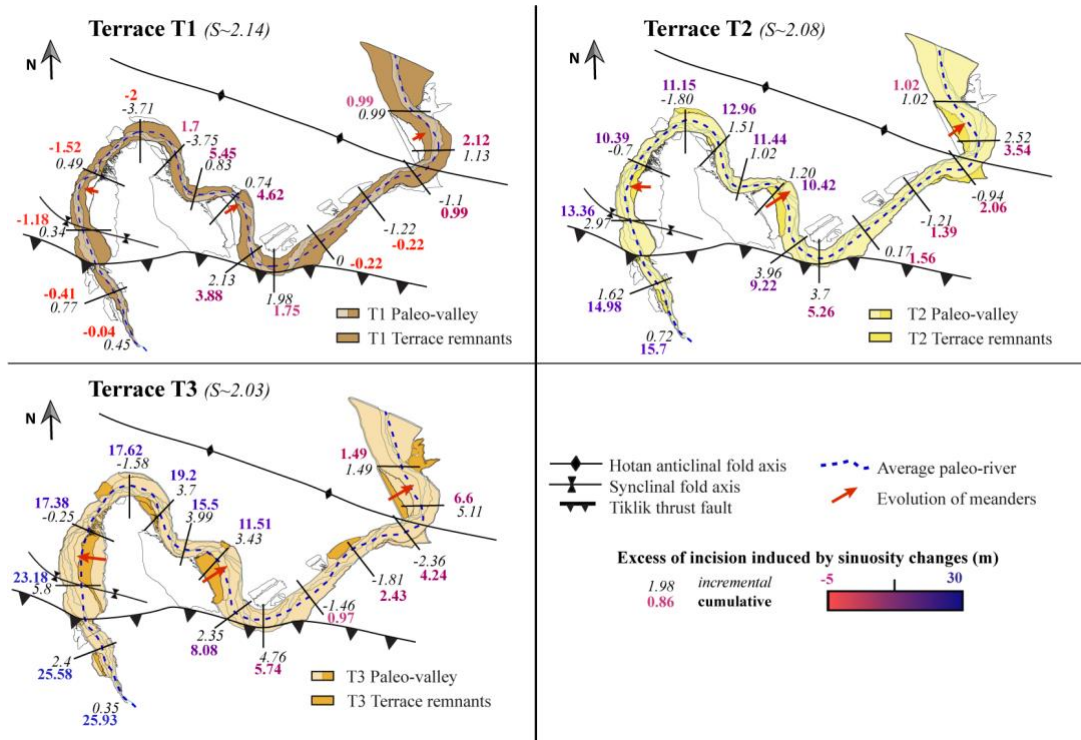
823

824 5.2.1.3 River sinuosity

825 The geomorphological map of Figure 5 is suggestive of changes in valley sinuosity over
 826 time, in particular where meanders appear to have been amplified since the abandonment of T5-
 827 T3, such as nearby our depth-profiles C and D collected for the dating of T3 and T2,
 828 respectively. To put tighter constraints on these changes, we reconstruct river paleo-valleys and
 829 define past river valley sinuosities by using our geomorphological map (Figure 5) and by
 830 adapting the approach of [Lavé and Avouac, 2000] to our field conditions (Figure 10). For these
 831 reconstructions, we follow some simple rules: (1) for each time step, equivalent to the time just
 832 prior to terrace abandonment, we consider the associated terrace record, (2) the extent of the
 833 corresponding paleo-valley is confined within parts of the landscape where elevations are equal
 834 to, or lower than, that of the preserved remnants of the considered terrace level, and (3) we
 835 consider an average paleo-channel flowing within the middle of the reconstructed paleo-valley
 836 and quantify its sinuosity. We call here for a particular attention on the fact that we are only able
 837 to reconstruct paleo-valleys and not the various paleo-channels that once flew within these
 838 valleys. This nuance between river channels and river valley is illustrated by the present-day
 839 Karakash River: the present river flows through various channels, with a seasonal and yearly
 840 variability, that will tend to occupy alternatively over time the whole present-day active valley.

841 In other words, the whole river valley is not fully occupied at any time; extrapolating this
 842 observation to the past river record, this signifies that reconstructed past river courses may not
 843 flow through all their corresponding terrace remnants at once. The quantified sinuosities rather
 844 represent the sinuosities of the river valleys rather than that of the actual river channels.
 845 Therefore, for a proper comparison between present and past rivers, we define an average
 846 modern river course flowing through the center of the present river valley.

847



848

849 **Figure 10** Reconstructed paleo-valleys, evolution of river sinuosity and associated excess
 850 incision. Panels represent the paleo-valley (colored areas) reconstructed from remnants of
 851 terraces T1 to T3 (deep colors). An average river course (dashed blue line) is determined as if
 852 flowing in the center of each paleo-valley. The reconstructed river course is segmented (dark
 853 cross-lines), and variations in river sinuosity and associated excess incision are calculated within
 854 each segment. Increments of excess incision from one segment to the other are reported (italic
 855 dark values, in meters), and cumulative ones (bold colored values, in meters) are determined by
 856 integrating all incremental values, upstream from base level.

857

858 Figure 10 shows the reconstructed valleys and average river courses for terraces T1 to
 859 T3. The paleo-valleys of T4, T5 and T6 could not be reconstructed since terrace remnants are too
 860 scarce or most often covered with loess or colluvium (Figures 4-5). The average river valley
 861 sinuosity S is derived from:

862

$$S = \frac{L}{D} \quad (5)$$

863 where L is the length of the average reconstructed channel and D is the downvalley path. The
 864 downvalley path is defined as the shortest direct distance between the considered downstream
 865 and upstream points. Here, from the river outlet into the Tarim Basin to the river outlet out of the
 866 inner range, we have a direct downvalley path of ~ 32 km. Calculated sinuosities are reported in
 867 Figure 10 and in Table 3. From there, we observe an overall increase in river valley sinuosity
 868 over time, from a sinuosity of ~ 2.03 in the case of T3 river to the present-day value of ~ 2.14 .

869 The contribution of this increase in sinuosity to incision along the river course can be
 870 quantified from:

$$871 \quad \Delta I_s = G \cdot (L_{x,t0} - L_{x,t}) \quad (6)$$

872 where ΔI_s is the excess incision related to changes in sinuosity, G the river gradient (in %), $L_{(x,t0)}$
 873 and $L_{(x,t)}$ the river longitudinal distances at point x for present-day ($t0$) and paleo-rivers at time t ,
 874 respectively. We remind here that considered river courses are taken from the average course
 875 within the center of the reconstructed river valleys. Longitudinal distances are determined
 876 upstream from base level, and point x is taken along the present-day river valley profile for
 877 reference. More precisely, we discretize the river course in various segments in which we
 878 compare average present-day and paleo-rivers to compute increments of ΔI_s from T1 to T3
 879 within these segments (Figure 10). These increments of ΔI_s are then summed and integrated all
 880 along the river course to quantify the cumulated value of ΔI_s , upstream from base level.

881 The results are reported in Figure 10, and in Table 3. In the upstream portions of our
 882 study site, variations in sinuosity have resulted in ~ 25 - 30 m of excess incision at most over the
 883 last ~ 250 kyr. In fact, the incision of the fill terraces found in this area, just downstream of the
 884 Tiklik thrust, correlates well with this excess incision (Figures 11a-b) and may simply result
 885 from river sinuosity variations over time.

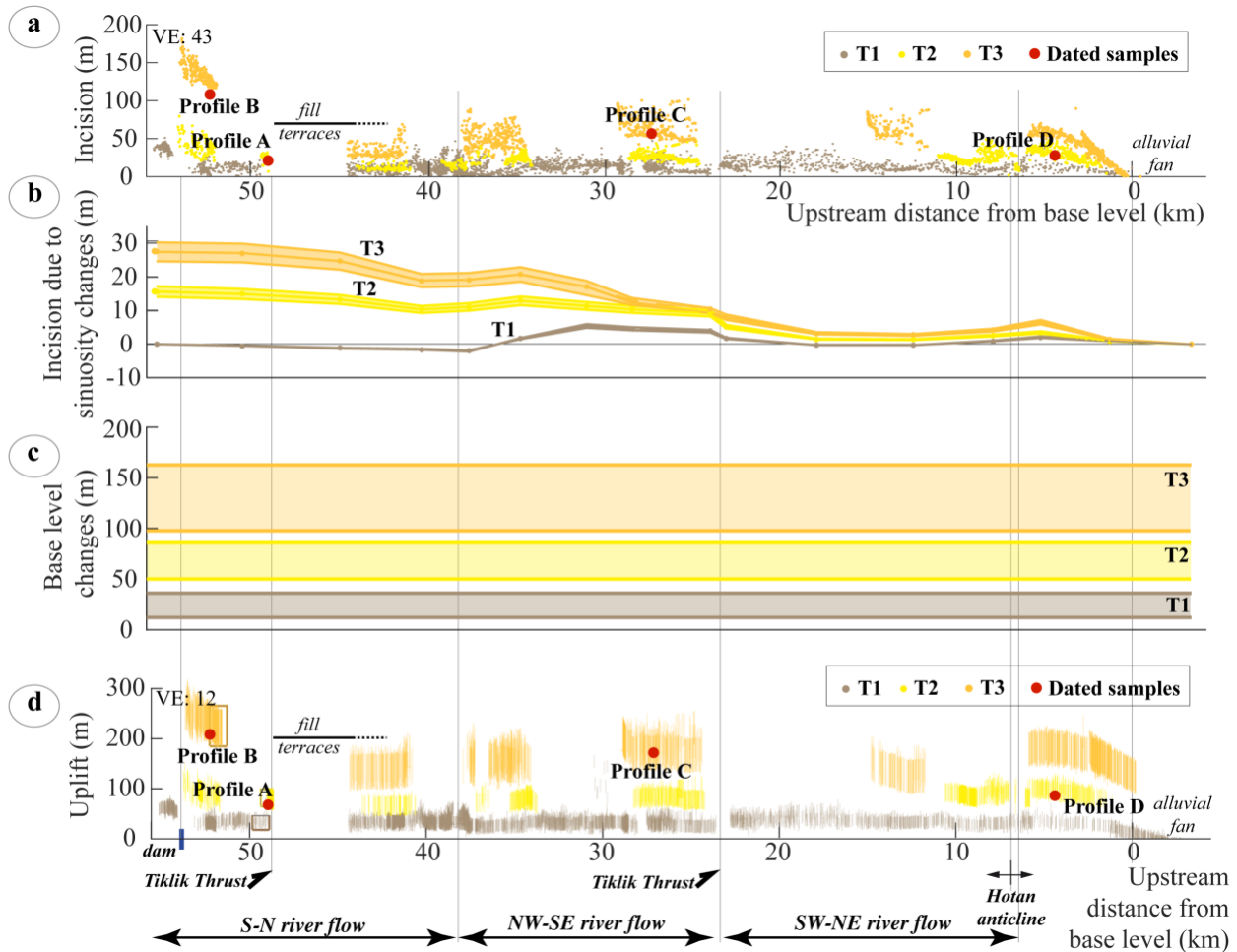
886 5.2.1.4 Incremental uplift

887 Incremental uplift since terrace abandonment is determined using equation (3) by
 888 following the long-distance profile of the Karakash River upstream from base level (Figure 11).
 889 All incision values are considered. Cumulative excess incision related to changes in sinuosity is
 890 attributed an arbitrary $\sim 10\%$ uncertainty, in the high range of the relative variation in sinuosity
 891 since T3 and by comparison to previous work [*Lavé and Avouac, 2000*] (Table 3). Corrections
 892 for base level integrate both the possible range of sedimentation rates, as well as the age intervals
 893 attributed to each terrace level (Table 3). Minimum uplift is determined for each related incision
 894 value by considering the minimum correction for base level (calculated from the minimum
 895 sedimentation rate of 0.4 mm/yr and the lower bound of attributed age) and the maximum excess
 896 incision related to sinuosity changes - and conversely, for maximum uplift. Overall, the range of
 897 values for calculated uplifts is most probably too large as uncertainties are most certainly
 898 conservatively overestimated.

899 Our computations reveal that base level changes contribute the most significantly to
 900 uplift, when compared to incision. As an example, incision related to the uplift of terrace T3 is of
 901 ~ 60 m in the zone of highest uplift behind the Hotan anticline (nearby depth-profile C, 28 km
 902 upstream from base level, Figure 11), when correcting the measured ~ 75 m incision (Figure 11a)
 903 for the ~ 15 m of incision related to variations in river sinuosity since T3 abandonment (Figure

904 11b). Comparatively, sedimentation at the base level amounts to ~125 m (Table 3, Figure 11c),
 905 that is ~2/3 of the total ~185 m uplift (Figure 11d).

906



907

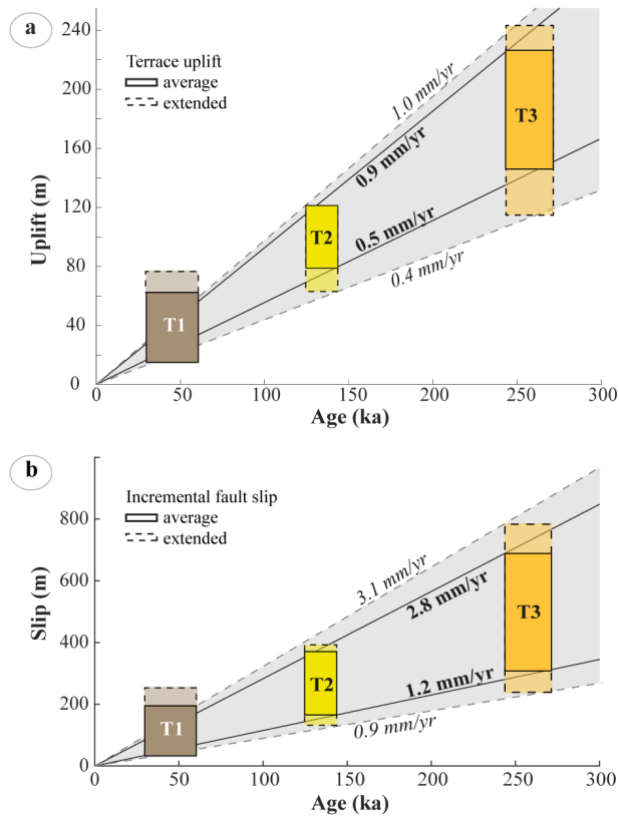
908 **Figure 11** From terrace incision to structural uplift. Only the strath terraces of levels T1 to T3
 909 are considered, along the long-distance river profile upstream from base level. Location of
 910 sampled depth profiles (Profiles A to D) are reported, together with that of main structural
 911 features encountered along the Karakash River course, in addition to the main directions of river
 912 flow, for reference. (a) Terrace incision as in Figure 6b, but reported here along the river course.
 913 (b) Cumulative excess of incision related to changes in river sinuosity over time (Figure 10). (c)
 914 Changes in base level, as deduced from the sedimentation rate and the interpreted ages of terrace
 915 abandonment (Table 3). (d) Structural uplift, calculated following equation (3) by correcting
 916 terrace incision (Figure 11a) from sedimentation at the base level (Figure 11c), and from incision
 917 related to sinuosity changes (Figure 11b). Vertical lines represent the range of uplift values found
 918 when considering minimum and maximum corrections for non-tectonic incision. Boxes in the
 919 hanging wall of the Tiklik thrust (to the left) indicate the values used to quantify the maximum
 920 vertical offset across this fault (Figure 14c).

921

946 been active over the time span of the terrace record and contributed to broad, even though
 947 possibly limited, uplift in this area. Overall, terrace anticlinal folding spatially coincides well to
 948 the first order with the ramps of the footwall duplex, confirming the idea that these structures
 949 have been active over the time span covered by the uplifted terraces.

950 **5.2.2 Fault slip rate on the Hotan footwall duplex**

951 We extract the range of maximum uplift for terrace levels T1 to T3 above the footwall
 952 duplex ramps (boxes in Figure 12), and compare it to terrace ages (Figure 13a). On average, the
 953 maximum uplift rate above the footwall ramps has been of 0.5-0.9 mm/yr over the last ~250 kyr.
 954 This range of uplift rate values could possibly be extended to ~0.4-1.0 mm/yr when accounting
 955 for the largest uncertainties on uplift estimates (dashed boxes on Figures 12-13).



956 **Figure 13** Uplift (a) and incremental fault slip (b) compared to terrace age. The range of
 957 considered terrace uplift is taken from the uplift determined above the ramps of the footwall
 958 duplex (boxes on Figure 12). The considered range of values can be extended when considering
 959 the largest uncertainties in uplift estimates (dashed boxes, here and Figure 12). Incremental fault
 960 slip is determined from terrace uplift and the dip angle of the underlying ramps following
 961 equation (4). We recall that only terraces T2 and T3 were dated, the age of T1 being estimated
 962 from regional considerations.
 963

964
 965 Terrace uplift is used to derive the incremental fault slip from equation (4) (Figure 13b).
 966 Because the pattern of terrace uplift is not sufficiently defined in detail, in particular in the
 967 backlimb of the anticline, we only consider uplift where it is maximum (boxes in Figure 12),

968 away from possible growth kink bands in the terrace record. The dip angles of the underlying
969 blind duplex ramps are all measured to $\sim 23\text{-}24^\circ\text{S}$ on average (within the range $18\text{-}29^\circ\text{S}$) from
970 available structural cross-sections projected along the direction of section AA' (Figure 12). The
971 dip angles of the footwall ramps slightly increase from the most frontal to the most internal ones.
972 However, all dip angles remain within a comparable value, so that we simplify the system and
973 assimilate the duplex to one single ramp with an identical average dip angle of $18\text{-}29^\circ\text{S}$. Using
974 equation (4), incremental fault slip is deduced from terrace uplift for each level T1-T3 (Figure
975 13b). Maximum fault slip values are determined from maximum uplift and minimum fault dip
976 angle values, and conversely for minimum fault slip, so that the range of fault slip represented by
977 the boxes on Figure 13b is probably too wide but conservative. These various results are reported
978 on Table 3. When incremental slip is compared to terrace ages, an average fault slip rate of 1.2-
979 2.8 mm/yr is found over the last ~ 250 kyr, a rate possibly from 0.9 to 3.1 mm/yr when
980 accounting for the larger uncertainties on incision and uplift values (dashed boxes on Figure 13).

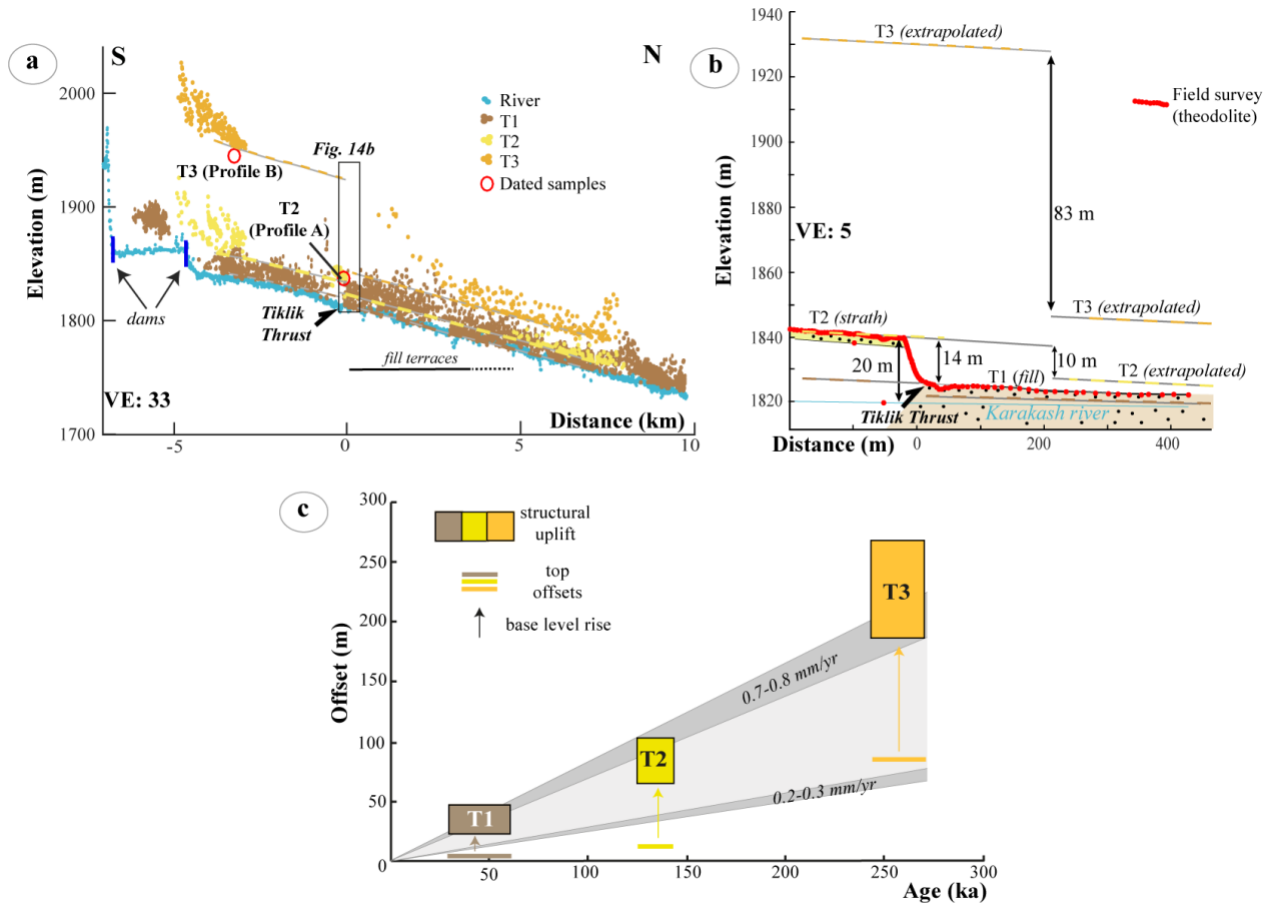
981 The broad pattern of terrace uplift does not allow to derive the detailed kinematics of the
982 blind duplex ramps (Figure 12), i.e. whether all ramps are active or not, and how slip is
983 partitioned between them. The fault slip rate of $1.2\text{-}2.8$ mm/yr determined here is that on the
984 overall duplex system, independently of the number and location of actual active ramps. This slip
985 rate is that transmitted from the deeper Cambrian decollement at the back to the footwall duplex,
986 and from there to the shallow Paleogene decollement northward and upward into the Tarim
987 Basin.

988 5.3 Slip rate on the Tiklik thrust

989 The Tiklik fault is active, as indicated by the ~ 14 m high scarp where it crosses the
990 Karakash River (Figure 4e). This scarp separates terrace levels T2 and T1 on the hanging-wall
991 and footwall of the fault, respectively. In addition, it should be reminded that terraces
992 immediately downstream of the Tiklik fault scarp were identified as fill terraces (Figure 4e), in
993 contrast with the strath terraces observed in the hanging wall of the thrust fault or further
994 downstream toward the Hotan anticline. These geomorphic features are not directly comparable
995 as they indicate lateral changes in the river behaviour, from incision to aggradation.

996 First, we consider the northward flowing Karakash River segment to the west of our
997 investigated area where the fault scarp is observed (Figure 14a). We laterally extrapolate the
998 geometry of the top of the various terrace levels T1 to T3 on either side of the Tiklik thrust
999 (Figure 14a) to quantify their apparent vertical offset, and combine these data with the fault scarp
1000 measured in the field (Figure 14b). From there, we derive a possible fault offset of ~ 4 m across
1001 T1, even though not clearly measurable. We get values of ~ 10 m and ~ 83 m for the offsets of T2
1002 and T3, respectively (Figure 14b). When combined to terrace ages, we find an average vertical
1003 displacement rate of $\sim 0.2\text{-}0.3$ mm/yr, with a faster rate of ~ 0.6 mm/yr between T2 and T3 when
1004 compared to the rate of <0.1 mm/yr since T2 (Figure 14c). However, we cannot exclude the fact
1005 that the top of the fill terraces in lateral geometric continuity with the identified and dated strath
1006 terrace levels are not precisely of the same age, as fill terraces have been documented elsewhere
1007 to be time-transgressive [Weldon, 1986] (Figure 8). The above values may therefore represent
1008 apparent offsets and should be considered as minimizing the actual vertical throw across the
1009 Tiklik thrust.

1010



1011

1012 **Figure 14** Analysis of terrace offsets across the Tiklik thrust. (a) River and terrace profiles,
 1013 along the northward flowing segment of the Karakash River to the west of our study site (Figure
 1014 5). Elevation data are projected parallel to section AA', which is also approximately orthogonal
 1015 to the Tiklik thrust. The dashed colored lines, following the color code of terrace data,
 1016 extrapolate the geometry of the terrace surfaces towards the Tiklik thrust. Box locates Figure
 1017 14b. (b) Zoom into the area around the Tiklik fault scarp. A topographic theodolite profile (red
 1018 points) documents the fault scarp, as well as the thickness of the T2 deposits or the elevation of
 1019 the river in the fault hanging wall. These data are complemented by topographic data from
 1020 available DEMs, in this area or extrapolated to it (dashed colored lines, from Figure 14a).
 1021 Terraces are strath terraces on the hanging wall of the Tiklik fault, with <5m thick deposits, and
 1022 are observed to be fill terraces in the fault footwall. (c) Vertical offsets across the Tiklik thrust,
 1023 compared to terrace ages. Color-code represents terrace levels T1 to T3. Horizontal thick lines
 1024 correspond to the offset of terrace surfaces extrapolated to the fault scarp (from Figures 14a-b).
 1025 These values should be taken as minimum estimates of the actual fault offsets. Boxes represent
 1026 the structural uplift of terraces, relative to the Tarim fixed footwall, as quantified in Figure 11d.
 1027 These values should be taken as maximum estimates of the actual fault offsets. Altogether the
 1028 uplift rate of the Tiklik thrust over its footwall is estimated to 0.2-0.8 mm/yr.

1029

1030 Indeed, the former riverbed, in continuity with a dated and investigated strath terrace
 1031 level, is most probably buried within the fill terraces or below the present-day river in the

1032 footwall of the Tiklik thrust. This burial amounts at most to the rise in river base level since
1033 terrace abandonment, in the case that there is no structural uplift of the Tiklik footwall behind the
1034 uplifting Hotan anticline during the considered time span. This situation is illustrated in Figure 8.
1035 In this case, the vertical throw of the paleo-river remnant across the Tiklik thrust can be
1036 calculated by adding the amount of base level rise in the footwall to that of terrace incision in the
1037 hanging wall, provided that this incision is corrected for non-tectonic parameters such as changes
1038 in river sinuosity and gradient. This calculation is equivalent to that of the structural uplift of
1039 terrace remnants in the hanging wall of the Tiklik thrust relative to the Tarim footwall, as
1040 illustrated in Figure 11. From this, we get maximum vertical throws of 20.7-44.8 m, 64.7-99.9 m
1041 and 186.5-265.0 m across the Tiklik thrust for terrace levels T1, T2 and T3, respectively (boxes
1042 in Figure 11d). When combined with terrace ages, we get an average maximum uplift rate of 0.7-
1043 0.8 mm/yr over the Tiklik thrust, with possible but limited evidence for variations of this rate
1044 over time (Figure 14c).

1045 We cannot exclude a slight structural uplift of the immediate Tiklik footwall with respect
1046 to the Tarim footwall, behind the active ramps of the footwall duplex, or related to the Hotan
1047 thrust if still active. As such, we remind that the latter uplift values should be taken as
1048 maximizing the actual vertical throw of terraces across the Tiklik thrust and the derived uplift
1049 rate. The difference between this uplift rate and the formerly derived rate of 0.2-0.3 mm/yr from
1050 the offset of terrace surfaces resides in the ~0.5 mm/yr rise in base level at the fault footwall
1051 (Figure 14c).

1052 To summarize, our data indicate an uplift rate in the range of 0.2-0.8 mm/yr over the
1053 Tiklik thrust over the last ~250 kyr (Figure 14c). The balanced cross-sections of [Baby *et al.*,
1054 2022] indicate that the thrust dips 45-50°S. From equation (4), we therefore get that the Tiklik
1055 thrust has slipped at a rate of 0.3-1.1 mm/yr on average over this time period. In the details, with
1056 the available data, we cannot exclude that fault slip has not been steady over time, with higher
1057 rates from T3 to T2, i.e. from ~250 ka to ~130 ka -, and limited fault displacement since ~50-100
1058 ka (Figure 14c).

1059 **6 Discussion**

1060 **6.1 Kinematics of thrusting of the Tiklik Thrust**

1061 **6.1.1 Uncertainties on our results**

1062 The Tiklik fault is segmented on various branches further west, at places even hidden
1063 below loess deposits (Figure 2a), and terminates rapidly further east towards the Altyn-Tagh
1064 strike-slip fault system (Figure 1). As such, field conditions are limited to provide direct fault
1065 slip rate estimates at sites other than the one investigated here, where the fault is well marked in
1066 the landscape by a scarp. One of the difficulties encountered here, though, lies in the fact that
1067 geomorphic markers on either side of the scarp are not directly comparable, with strath and fill
1068 terraces on the hanging-wall and footwall, respectively (Figure 14b). We overcome this difficulty
1069 by considering two extreme cases when analysing the terrace record, conservatively enclosing
1070 the fault slip rate within a minimum and a maximum value (Figure 14c).

1071 6.1.2 Comparison to previous estimates

1072 Along the Yurunkash River, ~50 km east of the fault scarp investigated here, apatite
1073 fission-track (AFT) ages from the broad basement anticline overthrusting the Tiklik fault record
1074 ~4-9 km of exhumation since the Late Oligocene, with proposed higher earlier rates up to the
1075 Early Miocene and more moderate exhumation since then [X Cheng *et al.*, 2017]. In more
1076 details, sample KLKS-7 of [X Cheng *et al.*, 2017], taken within the immediate hanging wall of
1077 the Tiklik thrust, has an AFT central age of ~22.5 Ma. This age is not well resolved as the
1078 sample did not pass the χ^2 test, possibly indicating that it was within the AFT partial annealing
1079 zone when exhumation initiated ~22.5 Myr ago. It is presently located ~300 m above the Tarim
1080 Basin base level, a value neglected hereafter with respect to estimates on exhumation and burial
1081 since 22.5 Ma. Considering a ~20° C/km geotherm, this sample suggests locally ~3-6 km of
1082 exhumation of the hanging wall of the Tiklik thrust above the base level for erosion since 22.5
1083 Ma. On the footwall side of the thrust, the corresponding ~22.5 Ma old sedimentary horizon is
1084 presently buried at an estimated depth of ~6 km below base level along the mountain front
1085 (Figure 3). Altogether, this suggests a total vertical throw of ~9-12 km across the Tiklik thrust
1086 since ~22.5 Ma, a value in line with the possible minimum structural uplift of basement rocks on
1087 either side of the Tiklik thrust when considering that basement is at the surface in the hanging
1088 wall (Figure 2) and at a minimum depth of ~10 km in the footwall (Figure 3). This suggests an
1089 average long-term structural uplift rate of ~0.4-0.6 mm/yr over that time span. When considering
1090 the ~45-50°S dip angle of the fault down to the depth of ~10 km retrieved from balanced cross-
1091 sections (Figure 3), this translates into a long-term fault slip rate of ~0.5-0.9 mm/yr. These
1092 results are comparable to the rates derived here over a much shorter time span of ~250 kyr
1093 (Figure 14c). It suggests that the Tiklik Thrust has been active most probably continuously since
1094 the Miocene.

1095 6.1.3 Contribution of the Tiklik thrust to crustal shortening

1096 We documented here for the first time the recent slip rate on the Tiklik thrust, consistent
1097 with the long-term ~0.5-0.9 mm/yr rate deduced from thermochronological data and structural
1098 reasoning. Even though this fault is documented to dip at ~45-50°S down to a depth of ~10 km
1099 (Figure 3), it is expected to flatten at depth below the inner range (Figure 1c). Given this
1100 geometry, the ~0.5-0.9 mm/yr slip rate is used as a proxy for the shortening rate accommodated
1101 by the thrust. Over the last ~20 Myr, the Tiklik thrust has therefore absorbed a possible
1102 cumulated crustal shortening of ~10-18 km. This value is to be added to the shortening taken by
1103 the blind structures of the Western Kunlun foothills, which may reach values of 64 km as along
1104 section aa' across the Hotan anticline and Tiklik thrust [Baby *et al.*, 2022]. When compared to
1105 the total Cenozoic shortening of ~55-100 km estimated across the Western Kunlun Range by
1106 crustal mass balance budgets [Baby *et al.*, 2022; Laborde *et al.*, 2019], this confirms the idea that
1107 most of the crustal shortening has been absorbed in the foothills [Baby *et al.*, 2022], with some
1108 contribution of the Tiklik thrust at the front of the inner range.

1109 6.2 Kinematics of shortening and active deformation across the Hotan anticline

1110 6.2.1 Uncertainties on our results

1111 Using the terrace uplift record (Figure 12), we obtained a 1.2-2.8 mm/yr fault slip rate on
1112 overall Hotan duplex system (Figure 13b). Most uncertainties on the data and hypotheses used to

1113 derive this rate have been already discussed, step by step, while progressively reaching this
1114 result.

1115 We were not able to derive the detailed kinematics of the blind duplex ramps, and the
1116 broad uplift pattern may be indicative that several of the duplex ramps are (or have been
1117 recently) active, in particular the most frontal ones, as proposed in Pishan further west [*Y Zhang*
1118 *et al.*, 2023]. Because the dip angle values of the various ramps derived from structural sections
1119 are comparable, and because the overall slip rate is to be partitioned onto the various active
1120 ramps, we simplified the duplex system into an analog single structure slipping at the total rate
1121 transmitted from the deeper Cambrian decollement into the shallower Paleocene one. We
1122 recognize that the most internal ramps are slightly steeper than the most frontal ones and could
1123 contribute more to the recorded uplift, from a simple geometrical argument (Figure 8). However,
1124 based on mechanical arguments, we may also consider that the overall slip partitioning is most
1125 favorable onto the more shallowly dipping frontal ramps. We believe, however, that these details
1126 only slightly impact the overall slip rate value retrieved here, and that they are already
1127 encompassed within the large conservative uncertainties on our estimates, in particular when
1128 considering a broad range of dip angles for the considered analog single ramp.

1129 We also simplified our analysis by considering that tectonic deformation is cylindrical,
1130 even though there are evidence for lateral structural variations along the Hotan anticline (Figure
1131 2) [*Baby et al.*, 2022]. However, [*Baby et al.*, 2022] found that these variations were in fact
1132 mostly related to the eastward progressive initiation of the blind duplex in the footwall of the
1133 Hotan thrust, and that the total shortening, the age of deformation inception and from there the
1134 long-term shortening rate across the whole Hotan anticline did not vary significantly along strike.
1135 At the scale of the investigated area, which extends over ~25-30 km along strike (Figure 5),
1136 major structural changes mostly rely on the presence of a syncline at the back of the duplex only
1137 along the western section bb', and in the eastward decreasing cumulated deformation of the
1138 duplex (Figure 3). Despite these variations and the extent of the area, we find that the pattern of
1139 terrace incision and uplift projects well onto a single broad antiform that correlates with the
1140 underlying duplex (Figures 6b-c and 12). This signifies that the lateral structural changes of the
1141 blind duplex and the potential lateral variations in its overall slip rate, if existent, are not
1142 sufficient to be resolved by the ~250 kyr long terrace record.

1143 Finally, it should be reminded that terrace uplift is here mostly dominated by the
1144 significant amount of base level changes (Figure 11), determined from sedimentation rates over
1145 the last ~5 Myr, considered laterally constant since we have no indication for lateral or temporal
1146 changes in these rates in available cross-sections (Figure 3). Any variation in the pattern of
1147 terrace uplift related to the previously discussed unknowns are therefore expected to be
1148 smoothed by this substantial correction for base level, and should therefore be considered as
1149 encompassed within the conservatively derived uncertainties.

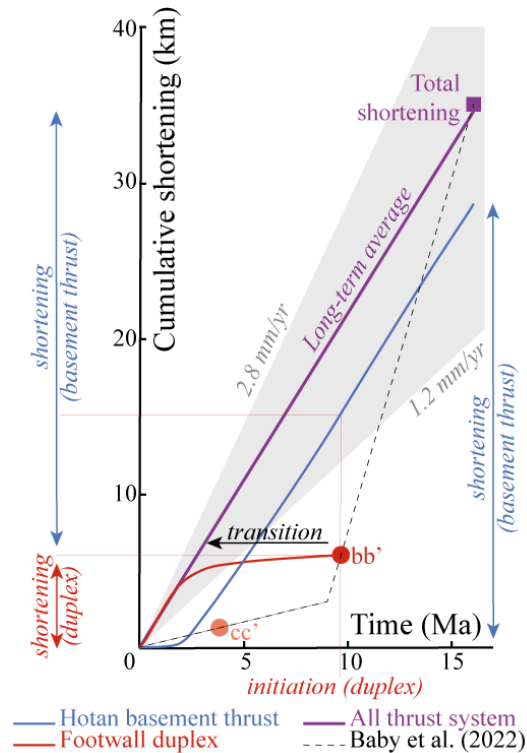
1150 6.2.2 Reconciling the kinematics of the Hotan anticline over various time scales:
1151 evidence for the recent activity of the Hotan thrust to the east?

1152 A 1.2-2.8 mm/yr slip rate is derived for the Hotan blind duplex, over a time span of ~250
1153 kyr, with a range of rates that could be conservatively extended to values from 0.9 to 3.1 mm/yr
1154 (Figure 13b). Given the large uncertainties on this estimate (Figure 13b), these values are only an
1155 average and we cannot rule out any temporal variability within the ~250 kyr time span of the
1156 terrace record. These values are in any case significantly higher than the long-term <0.5 mm/yr

1157 shortening rate proposed over the last ~8-9 Myr on the same duplex system from structural
1158 cross-sections [*Baby et al.*, 2022]. We have re-considered in detail the data used and the methods
1159 followed to reach each one of these results, independently, and did not find any particular reason
1160 to favor one more than the other as both analyses were found robust.

1161 One possible explanation lies in the fact that these results encompass a different time
1162 duration. If the fault system alternates periods of tectonic quiescence and activity, rates may be
1163 highly variable over a shorter time scale, even though they are expected to average over time to
1164 the geological long-term value. In other terms, if applicable here, this would suggest that the
1165 Hotan duplex system has been particularly active over the last ~250 kyr, with rates 3 to 6 times
1166 faster than the geological longer-term (Myr time scale) rate of <0.5 mm/yr – also suggesting
1167 ~100 kyr long periods of relative inactivity prior to ~250 ka. Episodic fault activity and related
1168 variable slip rates have been documented in various contexts since the first observations by
1169 Wallace [*Wallace*, 1987], and were attributed to variations in fault stresses or properties related
1170 to external climatic forcing (e.g. [*Calais et al.*, 2010; *Chéry and Vernant*, 2006; *Hetzl and*
1171 *Hampel*, 2005; *Luttrell and Sandwell*, 2010]), to variations in the strength of the fault zone or of
1172 the lithosphere, and/or to fault interactions (e.g. [*Chéry et al.*, 2001; *Chéry and Vernant*, 2006;
1173 *Dolan et al.*, 2007; *Dolan and Meade*, 2017; *Gunderson et al.*, 2018; *Peltzer et al.*, 2001]).
1174 However, such episodic fault activity has been documented in most cases over shorter millennial
1175 time scales (i.e. from geodetic time scales to few 1-10s kyrs). If real, the variations in rates
1176 documented in Hotan would therefore take place over a much longer time span than most often
1177 documented on continental active faults, and may call for other causal intrinsic processes,
1178 possibly related to the initiation and accretion of each duplex slice (e.g. [*Hoth et al.*, 2007;
1179 *Menant et al.*, 2020]), or to variable strain partitioning on connected thrusts [*Gunderson et al.*,
1180 2018].

1181 Interestingly, even though the 1.2-2.8 mm/yr rate derived over the last ~250 kyr is
1182 significantly higher than the one proposed over the last ~8-9 Myr by [*Baby et al.*, 2022], it
1183 coincidentally compares well with the longer-term geological average of ~2.2-2.5 mm/yr when
1184 considering the 35-40 km total shortening across the Hotan anticline and the 16 Ma time of
1185 deformation initiation [*Baby et al.*, 2022] (Figure 15). We recall here that [*Baby et al.*, 2022]
1186 hypothesized that thrusting over the Hotan thrust fully ended with the initiation of footwall
1187 duplexing, using this change in tectonic style as a marker of ongoing deformation. However, this
1188 change in tectonic style may not have been sharp in time but progressive, raising the possibility
1189 that basement thrusting has remained active together with footwall duplexing for a certain time,
1190 during a transitional period, as long as the basement thrust has not been significantly deformed
1191 by the underlying duplex. This situation would imply that not all the shortening attributed to the
1192 Hotan thrust from cross-sections had been consumed prior to footwall duplexing, but that some
1193 remained to be accommodated afterwards. Indeed, if this had been the case, ~10-15 km of
1194 shortening on the Hotan thrust needs to have been absorbed after the footwall duplex initiated,
1195 during a ~5-8 Myr long transitional period, to keep the long-term average shortening rate
1196 constant at ~2 mm/yr (Figure 15).



1197

1198 **Figure 15** Long-term kinematics of shortening across the Hotan anticline, re-evaluated from our
 1199 results on fault slip rates over the last ~250 kyr. The case of structural cross-section bb' [*Baby et*
 1200 *al.*, 2022] is illustrated here. The data on total shortening and time of overall deformation
 1201 initiation (purple square) across the anticline is reported, together with the shortening
 1202 accommodated by the footwall duplex and its timing of initiation (red circle). In case the
 1203 initiation of thrusting on the Hotan thrust abruptly ends when duplexing initiates, such data on
 1204 section bb' and other sections studied by [*Baby et al.*, 2022] (such as cc' reported here in light
 1205 red) indicate fast initial shortening rates, decelerating to slower rates by 8-9 Myr (dashed lines).
 1206 However, in case deformation is progressively transferred from the Hotan thrust (blue line) to the
 1207 footwall duplex (red line), over time during a transitional period (arrow), it might be possible
 1208 that the overall shortening rates remain constant over time, at the long-term geological rate
 1209 (purple line). The slip rate determined over the last ~250 kyr from terrace uplift is illustrated by
 1210 the grey area.

1211

1212 Figure 15 illustrates our latter reasoning for section bb', to the west of the area
 1213 investigated here. When also applied to section cc' further east, it raises the question as to
 1214 whether the Hotan thrust remains presently active along this section, given the fact that the
 1215 duplex initiated only ~4 Myr ago here [*Baby et al.*, 2022]. We have no evidence for significant
 1216 lateral variations in terrace folding and uplift along the Karakash River, suggesting that the
 1217 recent kinematics of the footwall duplex is laterally comparable over structural sections bb' and
 1218 cc', given the resolution of our data. If still active, the Hotan thrust is therefore expected to only
 1219 accommodate a limited slip rate along section cc', not resolvable from our terrace analysis.
 1220 Further east (section dd' of [*Baby et al.*, 2022], east of Hotan City and of the Yurunkash River),
 1221 the footwall duplex is absent and we might speculate that the Hotan thrust is here active at a rate

1222 of, at most, the value estimated from the folded terraces along the Karakash River, or maybe less,
1223 as the overall Hotan anticline vanishes eastward (Figure 2).

1224 To summarize, even though we cannot rule out episodic fault slip on the structures
1225 forming the Hotan anticline at the 1s-100s kyr time scale, we favor the idea that our rate
1226 estimates and those by [Baby et al., 2022] can be reconciled by considering a progressive
1227 transfer of slip and deformation from the Hotan thrust to the various ramps forming the footwall
1228 duplex. A similar situation, with a progressive forward transfer of deformation between
1229 structures, has been also proposed for the Yecheng-Pishan anticline, based on the similar
1230 observation that the recent slip rate derived from terraces is higher than the rate derived from
1231 only the most recent and frontal blind duplex ramp, even though the recent and long-term
1232 average rates are consistent [Guilbaud et al., 2017] – an idea further corroborated in Pishan by
1233 the broad terrace pattern [Y Zhang et al., 2023]. In the westernmost portions of the Hotan
1234 anticline, we therefore expect that only the duplex is presently active, as cumulative deformation
1235 of the basement thrust mechanically impedes slip on it. However, at the eastern termination of
1236 the anticline, this thrust is only slightly deformed - or even the unique structure beneath the
1237 Hotan anticline – and is therefore expected to be active. The transition between these two styles
1238 of active structures beneath the Hotan anticline is expected to be somewhere nearby section cc'
1239 of [Baby et al., 2022], or slightly further east (Figure 2).

1240 The deceleration of shortening rates across the Hotan anticline ~8-9 Myr ago proposed by
1241 [Baby et al., 2022] could therefore be an artefact, resulting from the progressive transfer of
1242 deformation from the Hotan thrust to the ramps of the footwall duplex, in time and in space, in a
1243 complex system where all structures are blind (Figure 15). The regional kinematic reorganization
1244 deduced by [Baby et al., 2022] from this deceleration is therefore also to be questioned, even
1245 though other detailed investigations, on other structures or locally in Hotan from better-resolved
1246 seismic data on growth strata or from higher and older fluvial terraces, would be needed to fully
1247 resolve this issue. In any case, the comparison between rates of deformation at various time
1248 scales, from long-term geological estimates to more recent Quaternary data, proves to be
1249 insightful by overcoming the limits of each approach to better reveal the presently active
1250 structures, their lateral extent and their slip rates.

1251 6.3 Active deformation and seismic segmentation along the mountain front of the 1252 Western Kunlun range

1253 Geomorphology indicates that the Tiklik thrust and the blind ramps of the duplex below
1254 the Hotan anticline are active, and therefore possibly seismic. This is also the case for the other
1255 blind ramps, connecting the Cambrian to the Paleogene decollements all along the Western
1256 Kunlun foothills, such as below the Yecheng-Pishan anticline [Guilbaud et al., 2017; J Xu et al.,
1257 2020]. A slip rate of 0.5 to 3.9 mm/yr has been proposed over the last ~300-500 kyr on the
1258 frontal blind ramp below the Pishan anticline, with most probable values of ~2-2.5 mm/yr when
1259 comparing all published estimates [Guilbaud et al., 2017; J Xu et al., 2020]. These rates are
1260 consistent with our findings in Hotan (Figures 13b), indicating that there may not be significant
1261 lateral variations in the recent slip rate on the blind ramps all along the Western Kunlun foothills,
1262 at least from Pishan to Hotan, and within the resolution of available geomorphic data.

1263 Even though slip rates have consistent values and are accommodated by frontal ramps
1264 that have comparable structural positions all along the mountain front, there may be a strong
1265 structural and kinematic lateral segmentation in the way recent deformation is absorbed. Indeed,

1266 geomorphology indicates that the most frontal blind ramp below the Pishan anticline, which
1267 ruptured during the 2015 Pishan earthquake, is most probably the only one to be presently active
1268 [Guilbaud et al., 2017; J Xu et al., 2020; Y Zhang et al., 2023]. In Hotan, we cannot discard the
1269 possibility that several of the blind footwall ramps are active together (Figure 12b) and that there
1270 may be some lateral variations in the actual active ramps below the anticline, in a context where
1271 these ramps have been propagating forward into the Tarim foreland basin and eastward towards
1272 the lateral termination of the anticline [Baby et al., 2022]. In fact, the exact lateral extent of each
1273 blind ramp, either in Hotan or all along the mountain foothills, is unknown. In addition, as
1274 discussed in section 6.2.2, the blind Hotan thrust could remain active at places, in particular
1275 along a <40 km long segment east of the section cc' across the anticline (Figure 2). Altogether,
1276 this indicates that active deformation is taken by one or several different active blind structures
1277 all along the mountain front, even though all these structures have a similar structural position by
1278 connecting the Cambrian and the Paleogene decollements, and even though they accommodate
1279 together a comparable slip rate at the time scale of several seismic cycles.

1280 This situation results in the fact that the kinematics of active deformation is probably
1281 more segmented along the mountain foothills than suspected from surface geology (Figure 2), as
1282 slip is distributed on various of the blind structures that all have a limited lateral extension.
1283 Because seismic ruptures may be subdued by structural complexities, as observed in other
1284 contexts (e.g., [Cubas et al., 2022; Hubbard et al., 2016; King and Nabelek, 1985; Wesnousky,
1285 2006]), such segmentation in the way active deformation is absorbed all along the Western
1286 Kunlun foothills may limit the magnitude of the earthquakes that release the cumulated crustal
1287 stresses, at least partly. As a matter of fact, the recorded historical seismicity is rather moderate,
1288 with ~10-15 thrust earthquakes with magnitudes Mw5-6 since the 1950's (US Geological
1289 Survey, 2023, Earthquake Catalog), the 2015 Mw 6.4 Pishan earthquake being the largest recent
1290 event recorded in the area (Figure 1).

1291 [Guilbaud et al., 2017] determined that an earthquake like Pishan is expected to rupture
1292 the same patch every ~200 years, given the average ~40 cm coseismic slip [He et al., 2016; Wen
1293 et al., 2016] and the ~2 mm/yr long-term slip rate on the frontal blind ramps. A more integrated
1294 view on the seismicity over the whole length of the Western Kunlun foothills may be provided
1295 by comparing the potency rate (integral of the fault slip rate over the fault area) over this region
1296 to the coseismic potency of the Pishan earthquake (coseismic slip over the ruptured area) (e.g.,
1297 [Simoes et al., 2007]). We get a potency rate of $\sim 9.10^6$ m³/yr over the whole ~ 300 km long
1298 foothills from Hotan to Yecheng, by considering ~15 km wide blind ramps slipping over the
1299 long-term at a rate of ~2 mm/yr. Given the ~40 x 20 km² patch that ruptured during the Pishan
1300 earthquake with an average coseismic slip of ~40 cm [He et al., 2016; Wen et al., 2016], this
1301 earthquake is found to have released a potency of $\sim 3.8.10^8$ m³. Given these results, we find that
1302 Pishan-like earthquakes are expected to occur every ~35 years all along the foothills – or longer
1303 in case slip is not released only during earthquakes. This recurrence time is slightly higher than
1304 that suggested from observations since the 1950s from global catalogs (US Geological Survey,
1305 2023, Earthquake Catalog), even though the ~60-70 years time window of observation is
1306 probably too short to conclude on a possible deficit of Mw6 (or higher) earthquakes.

1307 While the Tiklik thrust reaches the surface, the blind structures below the various
1308 anticlines forming the foothills connect upward onto the Paleogene decollement. As such, slip is
1309 transmitted northward and reaches the surface at the Mazar Tagh deformation front (Figure 1c).
1310 This situation kinematically implies that the Mazar Tagh thrust sheet is also active [Guilbaud et

1311 *al.*, 2017], with a slip rate of similarly ~2-2.5 mm/yr. In contrast with the structural and
1312 kinematic segmentation of the foothills, the geometry of the active Mazar Tagh thrust sheet
1313 appears remarkably simple, with a ~150-180 km wide and ~350 km long continuously shallowly
1314 dipping basal decollement (Figures 1b-c), only slightly deformed locally at places [*Chen et al.*,
1315 2022]. In other contexts, wide and smooth fault geometries have been proposed to favor the
1316 rupture of mega-earthquakes (e.g., [*Bletery et al.*, 2016; *Cubas et al.*, 2022]). If interseismically
1317 locked, the Mazar Tagh thrust sheet could therefore possibly slip during rare but major
1318 earthquakes, as hinted by its dimensions and its structural simplicity, with recurrence times of at
1319 least ~1.6-2 kyr [*Guilbaud et al.*, 2017]. Such situation would suggest the possibility of a
1320 bimodal seismic behavior, as proposed for instance in the Himalayas [*Bilham*, 2019; *Dal Zilio et al.*,
1321 2019], with on one hand relatively frequent moderate earthquakes that rupture the segmented
1322 deeper blind structures of the foothills and that transmit crustal stresses upward, and on the other
1323 hand rare major earthquakes rupturing the wide frontal Mazar Tagh thrust sheet. Further
1324 elucidating this question will require in the future to document the interseismic mechanical
1325 behavior of the Mazar Tagh thrust and explore the geological record of potential past major
1326 earthquakes in the region.

1327 **7 Conclusions**

1328 We analysed the incision of terraces along the Karakash River, where this river crosses
1329 the Hotan anticline along the eastern foothills of the Western Kunlun mountain range (Figure 5).
1330 From samples collected in the field within depth-profiles for in-situ produced ^{10}Be cosmogenic
1331 isotope dating, we determined the ages of two main recent terrace levels (Figure 7). From there,
1332 we quantified structural uplift (Figure 11) and the slip on the underlying blind structures since
1333 these terraces were abandoned. These various results indicate a slip rate of 1.2-2.8 mm/yr on the
1334 blind ramps of a duplex beneath the Hotan anticline over the last ~250 kyr (Figures 12-13). In
1335 addition, the Tiklik thrust, which separates the inner range from the foothills, forms a clear
1336 morphologic scarp that disrupts the terrace record. Our analysis and age results suggest that this
1337 thrust slipped at a rate of 0.3-1.1 mm/yr over the same time period (Figure 14).

1338 The slip rate determined on the blind ramps beneath the Hotan anticline over the last
1339 ~250 kyr is 3-6 times faster than the average shortening and slip rates derived across the same
1340 structures over the last ~8-9 Myr from structural sections [*Baby et al.*, 2022]. We propose that
1341 this discrepancy reflects the progressive transfer of deformation from the Hotan basement thrust
1342 to the footwall duplex, over a probable time span of ~5 Myr after the duplex initiates (Figure 15).
1343 This may suggest that the Hotan thrust remains active east of our study area, where the blind
1344 duplex is absent or incipient.

1345 When compared to existing data further west in Pishan [*Guilbaud et al.*, 2017; *J Xu et al.*,
1346 2020], our findings reveal that the blind structures all along the mountain foothills are active and
1347 slip at an overall rate of ~2 mm/yr. However, in the details, this kinematics is expected to endure
1348 a certain geometric segmentation, because of the possible partitioning of slip among several of
1349 the most frontal ramps, and because of the possibly limited lateral geometric extent of each
1350 structure. The ~2 mm/yr slip rate is then transferred upward and northward onto the shallowly
1351 dipping Paleogene decollement, all the way to the Mazar Tagh deformation front. This ~150-180
1352 km wide structure has a simple structural geometry that contrasts with the kinematic
1353 segmentation suggested for the blind ramps forming the mountain front (Figure 1). We suggest
1354 that this segmentation may explain partly the moderate seismicity recorded all along the

1355 mountain front. From there, we also question the possibility of a bimodal seismicity in the case
1356 of the Western Kunlun, as proposed in other contexts (e.g. [Bilham, 2019; Cubas et al., 2022;
1357 Dal Zilio et al., 2019]), with rare large ($M > 8$) earthquakes rupturing the Mazar Tagh thrust
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1380 **Open Research**

1381 Pléiades satellite imagery (<https://earth.esa.int/eogateway/missions/pleiades>, last access: January
1382 2018) was obtained through the ISIS program of the CNES under an academic license and is not
1383 available for open distribution. On request, the DEMs calculated from this imagery can be
1384 provided to any academic researcher, but only after approval from the CNES (contact: isis-
1385 pleiades@cnes.fr, with copy to simoes@ipgp.fr and referring to this paper). Numerical
1386 computations for the DEMs were performed using the free and open-source MicMac software
1387 suite [Rosu et al., 2015; Rupnik et al., 2017] freely available at <https://micmac.ensg.eu/index.php>
1388 (last access: April 2018). We completed our topographic data with the ALOS World 3D – 30m
1389 digital elevation model provided by the Japanese Aerospace Exploration Agency freely
1390 accessible at <https://www.eorc.jaxa.jp/ALOS/en/aw3d30/index.htm> (last access: April 2018). We
1391 also refer in this work to the US Geological Survey Earthquake Catalog, freely accessible at
1392 <https://earthquake.usgs.gov/earthquakes/search/> (last accessed: January 2024). The data collected
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Supporting Information for

Kinematics of Active Deformation and Possible Segmentation of Seismic Slip along the Foothills of the Western Kunlun (China).

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Additional Supporting Information (Files uploaded separately)

Caption for Table S1

Introduction

This file contains additional information on the geomorphology of the Karakash River (Figure S1), analytical procedures followed to analyze collected samples (Text S1 and Table 2), accompanied by a table with complete analytical data on each sample (Table S1, uploaded separately). Further details on the age inversion procedures and models are provided (Text S2), and possible trade-offs between inverted parameters are illustrated by additional figures (Figures S2 to S9).

Figure S1. Long-distance profile of the Karakash river

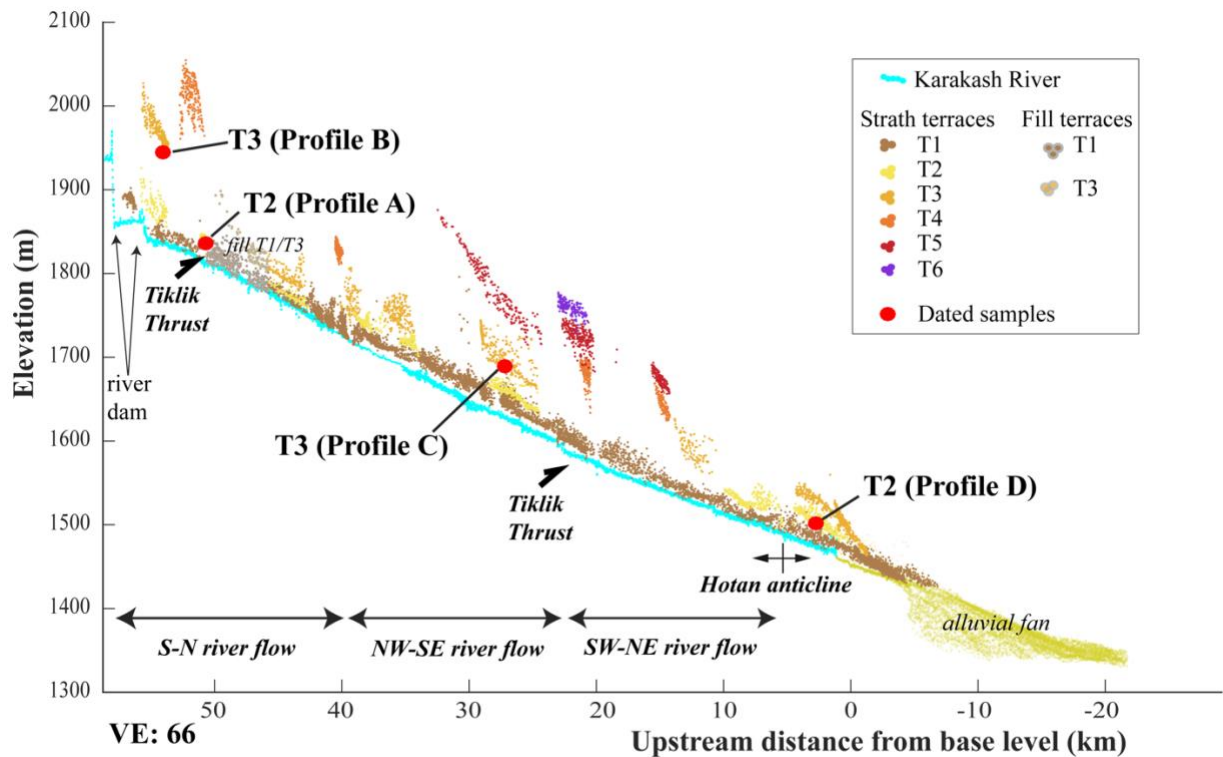


Figure S1. Long-distance profile of the Karakash River and associated terraces. Distances are indicated upstream from base level. Terrace levels are color-coded as in the map of Figure 5 and profiles of Figure 6. Location of sampled depth profiles (A to D) are reported together with correlated terrace levels. Various structural features encountered along the Karakash River course are reported for reference.

Text S1. Analytical procedures

Samples have been prepared and analyzed in the course of several years (2006 to 2020, after field campaigns in 2003, 2005, 2017 and 2019), in different laboratories (see Table S1 for analytical details), including *Institut Terre et Environnement de Strasbourg* (ITES, Strasbourg, France), *Centre Européen de Recherche et d'Enseignement des Géosciences de l'Environnement* (CEREGE, Aix-en-Provence, France) and *Purdue Rare Isotope Measurement Laboratory* (PRIME Lab, Purdue University, USA). These laboratories have different quartz isolation and purification protocols that we briefly describe below.

To isolate the quartz-rich fraction contained in the samples, all samples were crushed, sieved, and leached in concentrated hydrochloric acid. Quartz is then isolated and purified by several low concentrated fluorine and nitric acid leaches (at ITES and PRIME Lab) or hydrochloric and hexafluoric acid leaches followed by sequential fluorine acid leaches (CEREGE). This is to dissolve aluminum-rich feldspar minerals, etch the exterior portion of the quartz crystals in order to remove atmospheric ^{10}Be adsorbed on the mineral surfaces, and reduce the total amount of aluminum Al [Brown *et al.*, 1991; Gosse and Phillips, 2001; Kohl and Nishizumi, 1992].

About 0.2-0.5 mg of ^9Be carrier was added to the purified quartz that was then dissolved in concentrated fluorine acid. Preparing the samples for ^{26}Al measurements implied that after dissolution, total Al was measured in the sample solutions. ^{27}Al carrier is then added if the total mass of Al is less than 2 mg. Nuclides were then chemically separated using ion exchange columns. After precipitation, $\text{Be}(\text{OH})_2$ and $\text{Al}_2(\text{OH})_3$ were heated at 750°C to obtain BeO and Al_2O_3 .

The $^{10}\text{Be}/^9\text{Be}$ and $^{26}\text{Al}/^{27}\text{Al}$ ratios were then measured at ASTER (CEREGE), except for 2 samples (KA03-1A and KA03-1B) that were measured at the *Center for Accelerator Mass Spectrometry* at *Lawrence Livermore National Laboratory* (CAMS – LLNL, Livermore, USA) (see Table S1). Finally, the ratios were converted to ^{10}Be or ^{26}Al concentrations using the measured total Be or Al concentrations prior to chemical separation (Tables 1 and S1). As we do not have $^{26}\text{Al}/^{27}\text{Al}$ ratios for all samples, only those available are presented in Tables 1 and S1. They provide additional information on the quality of the analytical processing and can help point out possible exposure complexities (see main text).

The obtained AMS ratios were overall at least one order of magnitude larger than those of the analytical blanks (Table S1). It is therefore unlikely that large errors were introduced due to the processing of the various samples in different laboratories.

Table S1 (uploaded separately). Complete sampling and analytical data for all the samples collected in depth profiles A to D. These data are provided in an accompanying Excel file, where the first sheet (*Table Notice*) provides additional information on some of the columns of the second sheet (*Data*) where all data are reported. See Figure 5 for location of samples, and Figure 7 for a representation of concentrations of in-situ produced ^{10}Be along all depth-profiles. A simplified version of this table is provided in the main text (Table 1).

Text S2. Profile inversions and age interpretations.

In this study, we used the time constant scaling scheme “St” [Lal, 1991; Stone, 2000] with a production rate of 4.01 ± 0.33 atoms/g/yr at sea level and at high latitudes to constrain the local ^{10}Be production rate at our study site. We use here the CRONUS-Earth online calculator (v3, last consulted on December 2022; http://hess.ess.washington.edu/math/v3/v3_age_in.html) (e.g. [Balco et al., 2008; Borchers et al., 2016; Marrero et al., 2016]), and find values ranging from 11.87 (profile D) to 16.17 atoms/g/yr (profile B). Assuming the present river base level to represent the topographic level at the beginning of folding, taking into account the time-changing elevation of the sampled sites, those rates drop to values ranging from 11.71 to 14.94 atoms/g/yr, respectively. To model the data, we used attenuation lengths of ~ 160 , 1500 and 4320 g/cm^2 for neutrons, slow muons and fast muons, respectively [Balco, 2017; Braucher et al., 2011; Marrero et al., 2016].

Before inverting for the ages of each depth profile, we determined the density to be used for modeling (e.g., [Ackerer et al., 2016; Hancock et al., 1999; Perrineau et al., 2011]). Although the four terraces we sampled belong to the same river system and are distributed over a relatively small area, they have somewhat different granulometries, ranging from cobbles to sand (see Figure 7). Densities were measured in the lab or in the field for each profile, as well as estimated by in-depth cross-section photograph analyses and/or by inverting our sample concentrations as a function of depth (Table S2). Based on these estimates, we decided to use a constant density of 2.3 for the profiles mostly comprised of cobbles and pebbles (profiles A, B and C) and a density of 1.6 for the remaining profile (D) mostly comprised of sand and smaller gravels.

Sampled profiles (terrace level)	Measured density (laboratory/field)	Estimated density (profile analysis from photographs)	Estimated density (inversion of concentration profiles)	Density used in age inversions
A (T2)	-	2.37-2.25-2.21	2.5	2.3
B (T3)	-	2.62-2.59-2.58	2.0	2.3
C (T3)	2.34	2.37-2.25-2.21	2- 2.2	2.3
D (T2)	1.6	-	1.3-1.5	1.6

Table S2. Densities measured and/or estimated for each depth profile, and values considered in our age inversions.

We then modelled our profiles using two different approaches:

- A first model (Model 1), with a fixed erosion value set at 0.001 mm/yr (or 1 mm/kyr) corresponding to a local long-term estimate (sample at saturation for more than 1 Myr) on a nearby drainage basin (see [Guilbaud et al., 2017]) and a constant

density. The two remaining parameters (age and inheritance) are then inverted in a specific range of values (see Table 2). The best solution is a parameter couple of age and inheritance, which displays the minimum chi-squared value (Figures S2-S5). Results from this model are favored in our interpretations.

- A second model (Model 2), with all three parameters (age, erosion, inheritance) inverted in a specific range of values (see Table 2) and a constant density. In this case, the best solution determined is therefore a combination of age, inheritance, and erosion, which displays the minimum chi-squared value (Figures S6-S9). Results from this model are mostly considered as informative of the space of possible solutions and of possible parameter trade-offs.

- Errors for each parameter are computed for error yields of 2-sigma (blue zone in Figures S2-S9), 1-sigma (green zone in Figures S2-S9) and $\chi^2_{\min}+1$ (yellow zone in Figures S2-S9, a criterium classically used in age inversions (e.g., [Yang et al., 2019])). Additional details on these inversions are illustrated in Figures S2 to S9.

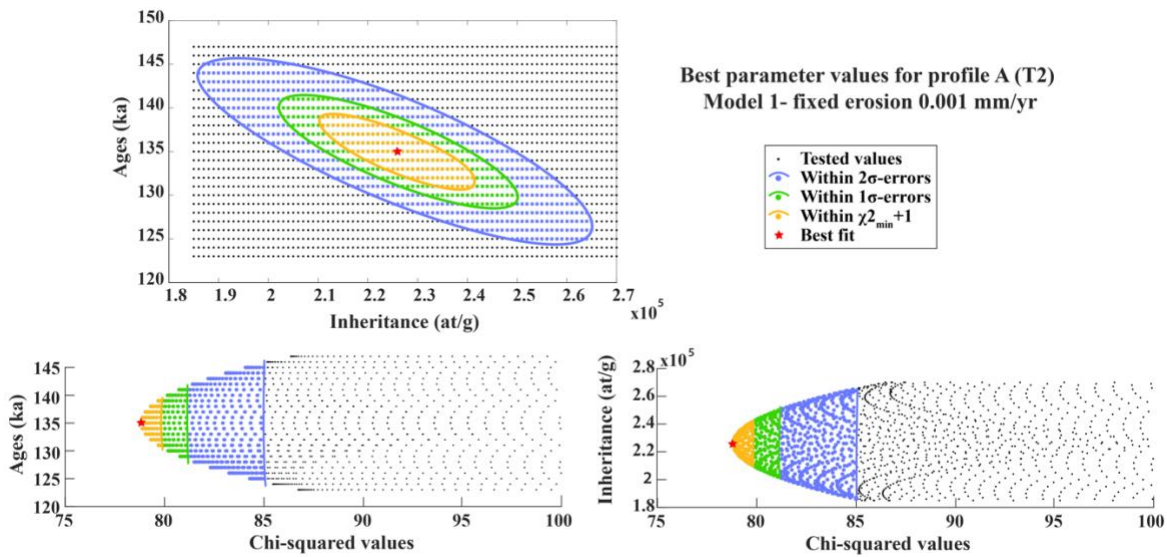


Figure S2. Inversion results for depth-profile A (interpreted as related to terrace level T2), in the case of Model 1 (fixed erosion). Age and inheritance are inverted and the solution is represented by the red star. Various confidence intervals are represented by different colored areas. Top left: trade-off between age and inheritance. Bottom: calculated χ^2 for the explored range of ages (left) and inheritance (right).

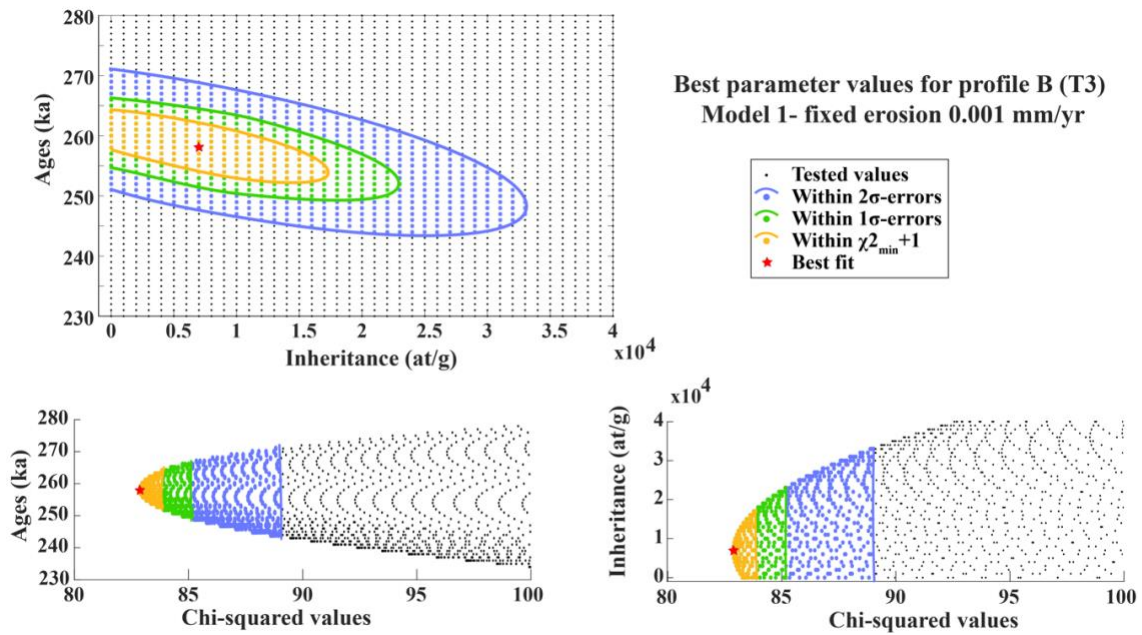


Figure S3. Inversion results for depth-profile B (interpreted as related to terrace level T3), in the case of Model 1 (fixed erosion). Age and inheritance are inverted and the solution is represented by the red star. Various confidence intervals are represented by different colored areas. Top left: trade-off between age and inheritance. Bottom: calculated χ^2 for the explored range of ages (left) and inheritance (right).

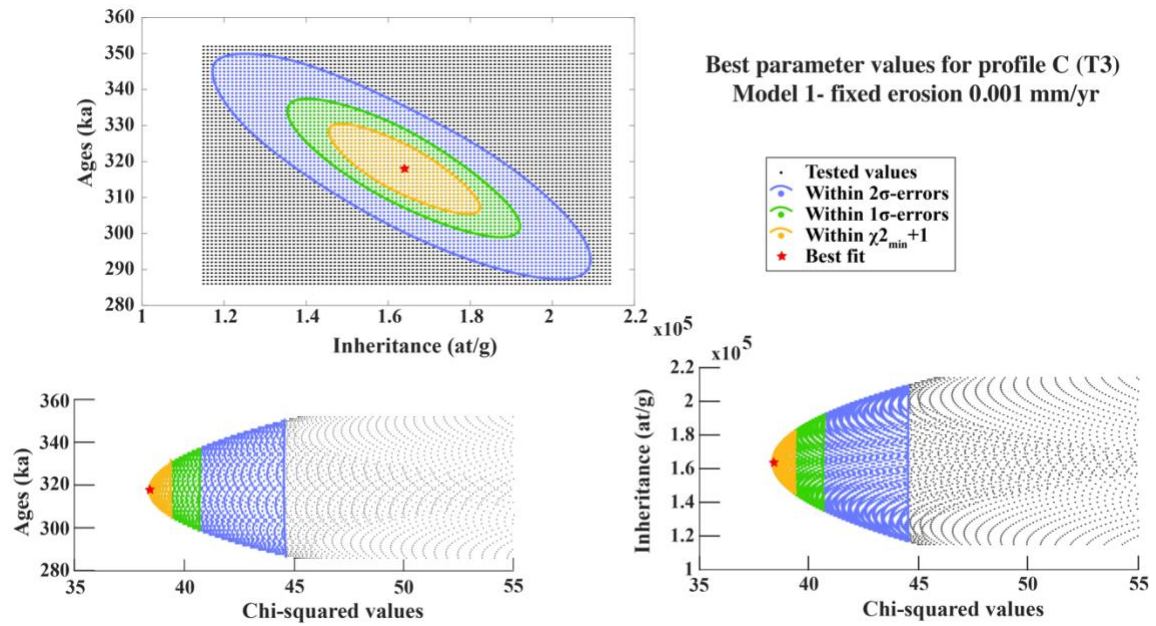


Figure S4. Inversion results for depth-profile C (interpreted as related to terrace level T3), in the case of Model 1 (fixed erosion). Age and inheritance are inverted and the solution is represented by the red star. Various confidence intervals are represented by different colored areas. Top left: trade-off between age and inheritance. Bottom: calculated χ^2 for the explored range of ages (left) and inheritance (right).

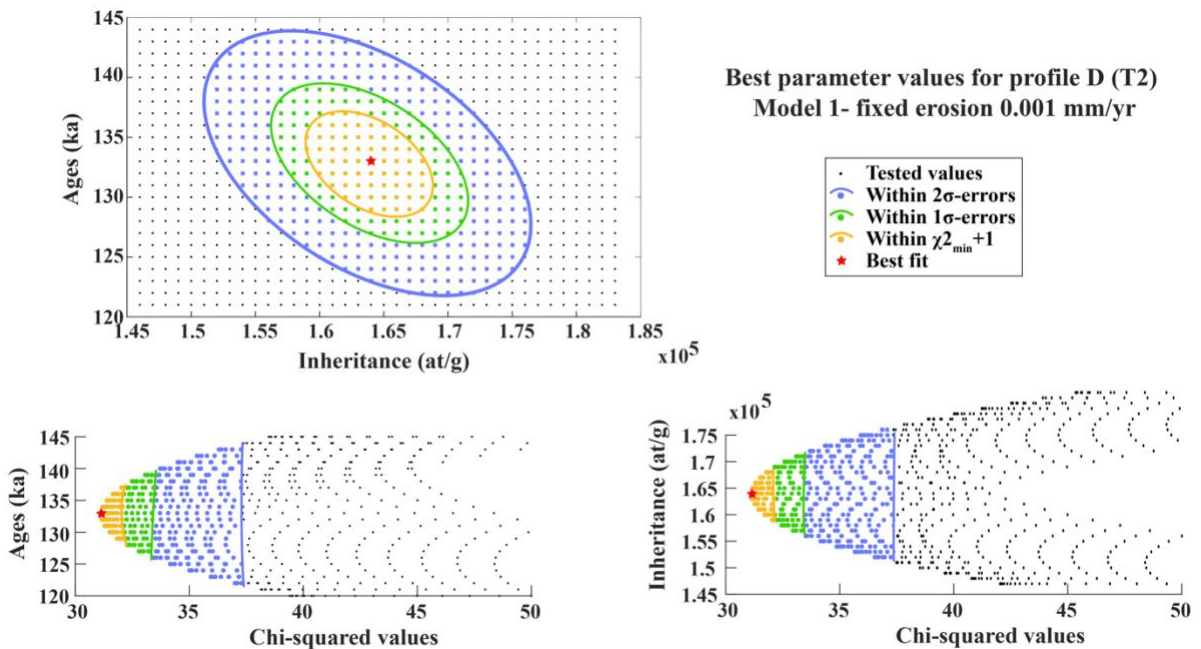


Figure S5. Inversion results for depth-profile D (interpreted as related to terrace level T2), in the case of Model 1 (fixed erosion). Age and inheritance are inverted and the

solution is represented by the red star. Various confidence intervals are represented by different colored areas. Top left: trade-off between age and inheritance. Bottom: calculated χ^2 for the explored range of ages (left) and inheritance (right).

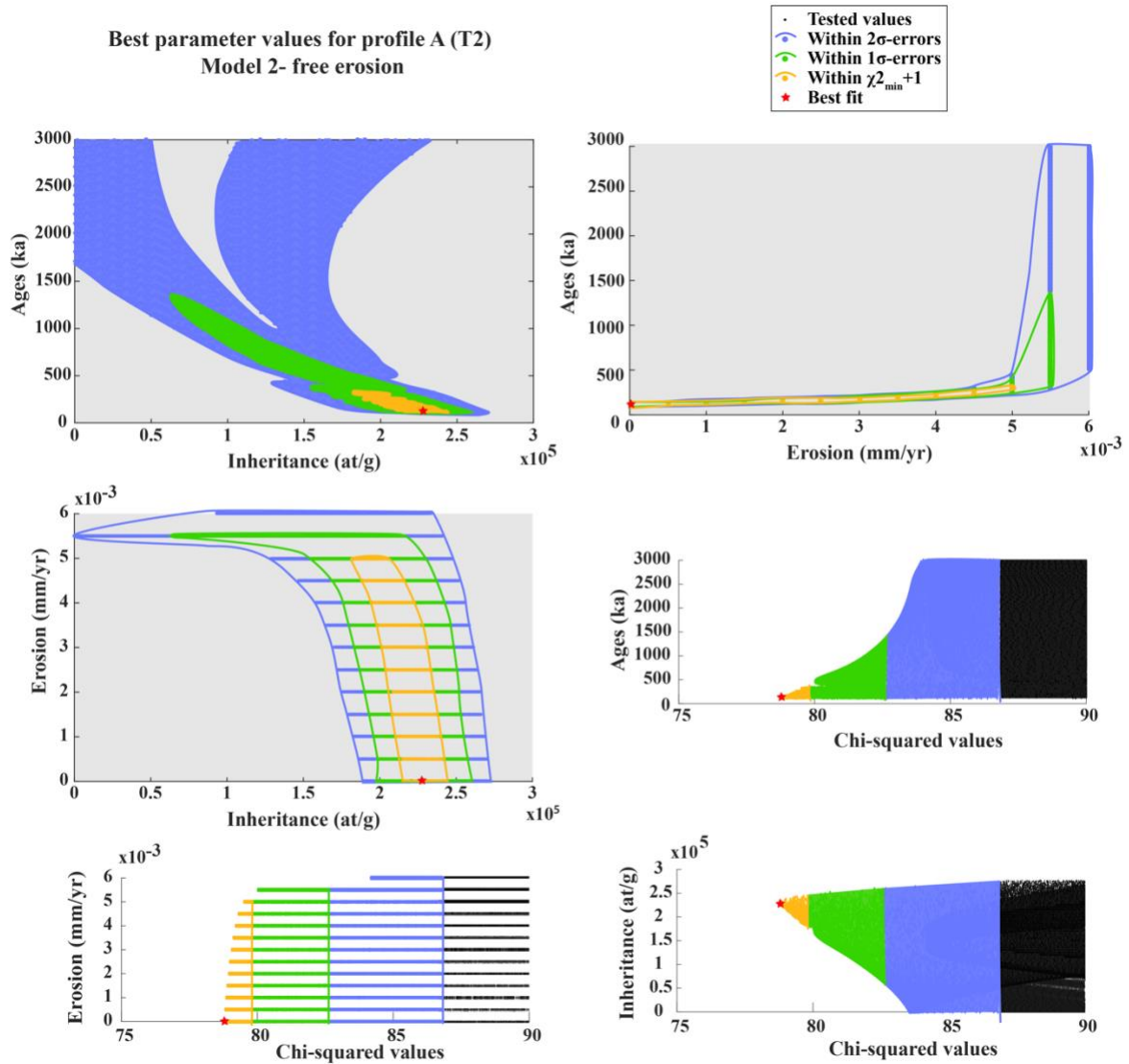


Figure S6. Inversion results for depth-profile A (interpreted as related to terrace level T2), in the case of Model 2. Erosion, age, and inheritance are inverted, and the best solution is represented by the red star. Various confidence intervals are represented by different colored areas. From top to bottom, and from left to right:

- * Top left panels: trade-offs between parameters, as illustrated in 2-dimensions for an easier reading (age vs. inheritance, age vs. erosion, and erosion vs. inheritance).
- * Bottom right panels: calculated χ^2 for the explored range of ages, erosion, and inheritance.

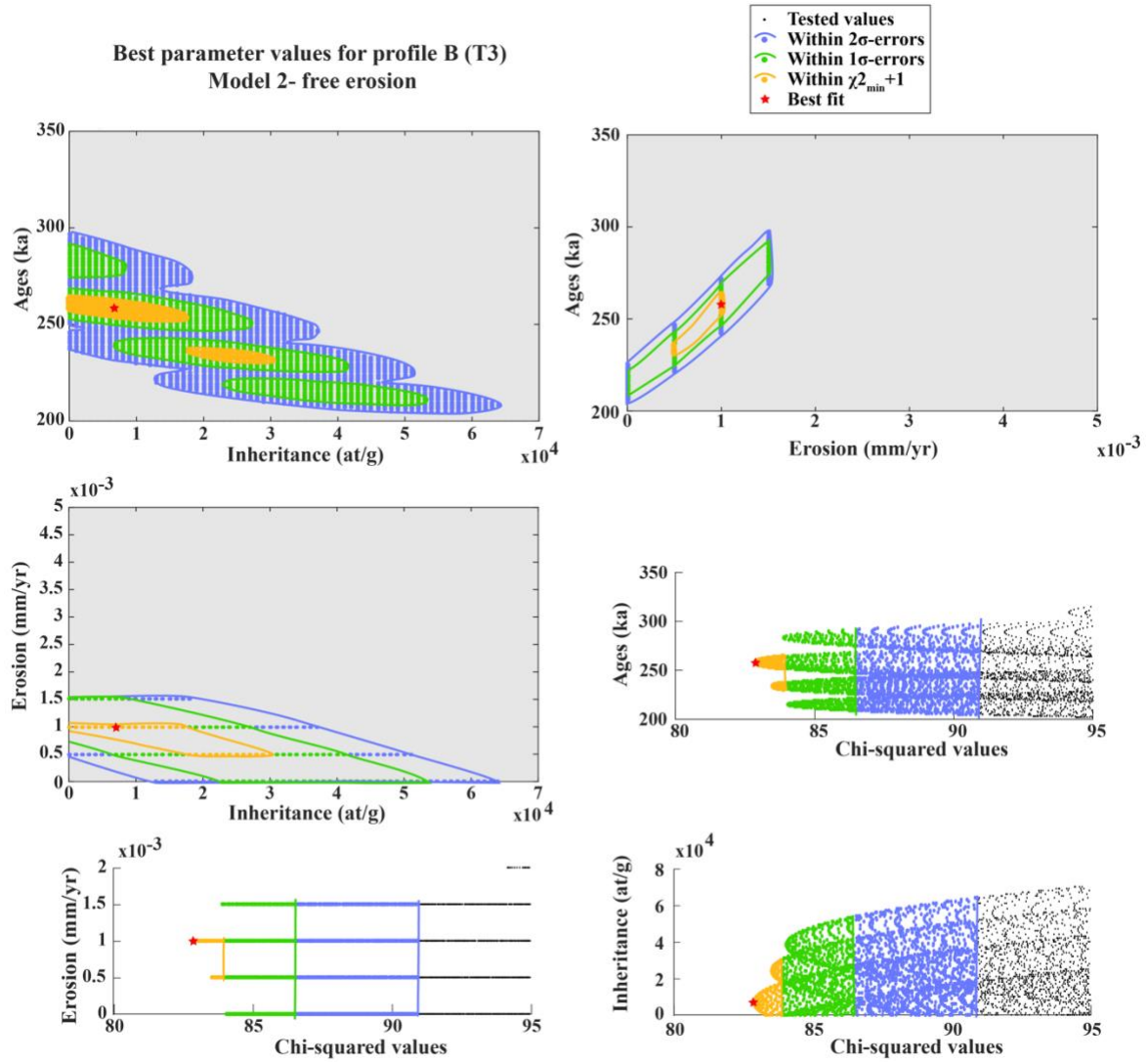


Figure S7. Inversion results for depth-profile B (interpreted as related to terrace level T3), in the case of Model 2. Erosion, age, and inheritance are inverted, and the best solution is represented by the red star. Various confidence intervals are represented by different colored areas. From top to bottom, and from left to right:

* Top left panels: trade-offs between parameters, as illustrated in 2-dimensions for an easier reading (age vs. inheritance, age vs. erosion, and erosion vs. inheritance).

* Bottom right panels: calculated χ^2 for the explored range of ages, erosion, and inheritance.

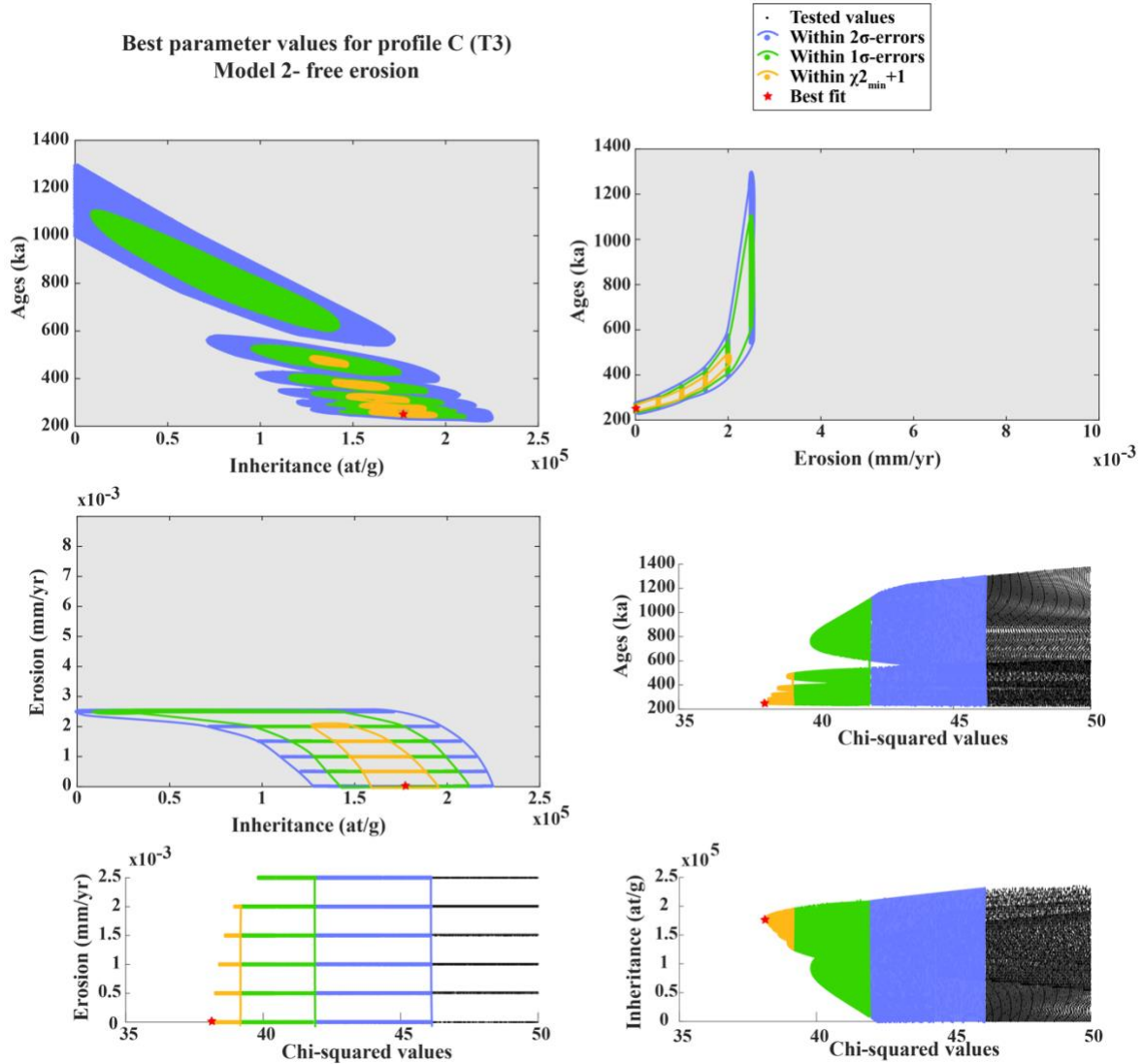


Figure S8. Inversion results for depth-profile C (interpreted as related to terrace level T3), in the case of Model 2. Erosion, age, and inheritance are inverted, and the best solution is represented by the red star. Various confidence intervals are represented by different colored areas. From top to bottom, and from left to right:

- * Top left panels: trade-offs between parameters, as illustrated in 2-dimensions for an easier reading (age vs. inheritance, age vs. erosion, and erosion vs. inheritance).
- * Bottom right panels: calculated χ^2 for the explored range of ages, erosion, and inheritance.

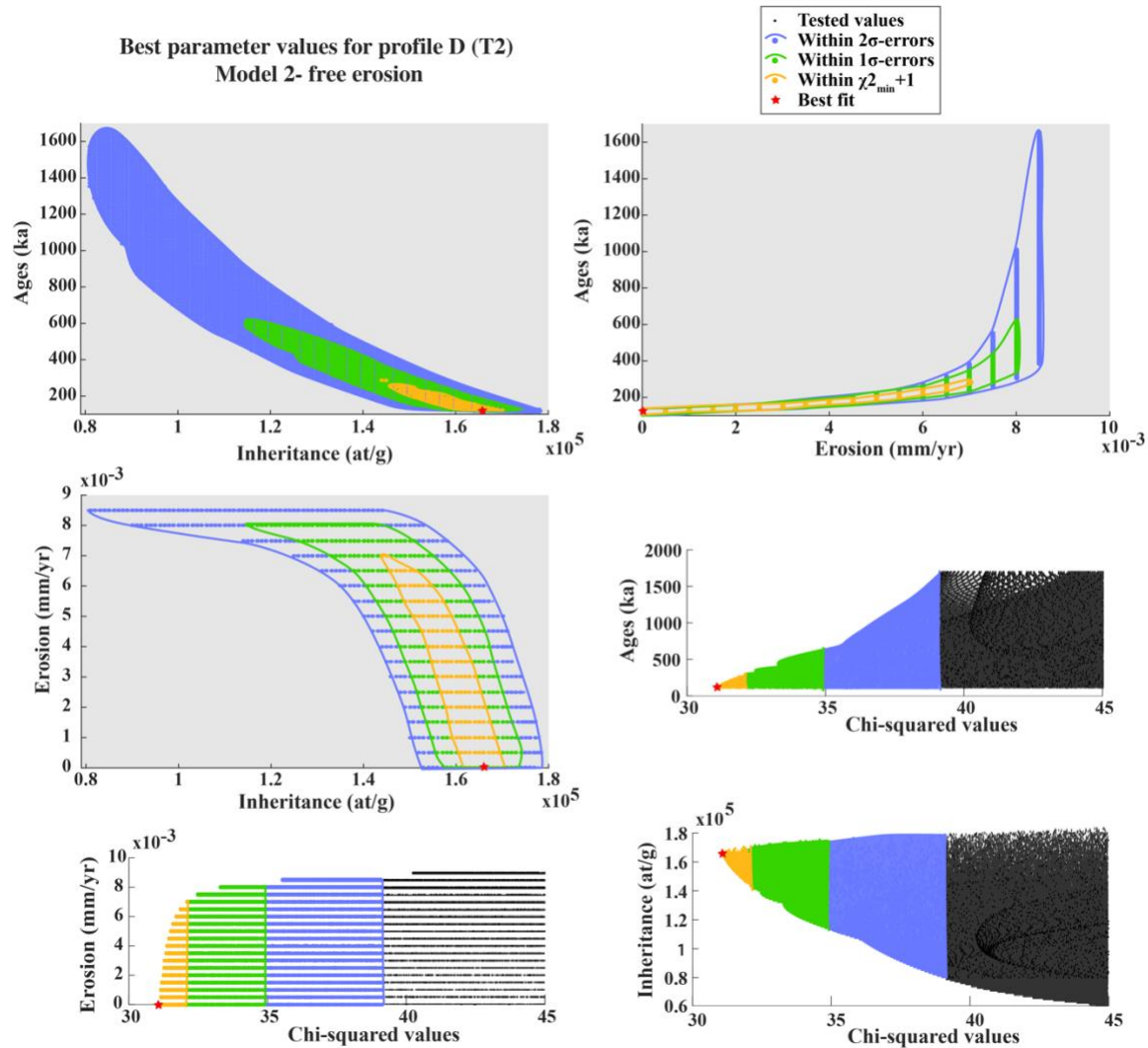


Figure S9. Inversion results for depth-profile D (interpreted as related to terrace level T2), in the case of Model 2. Erosion, age, and inheritance are inverted, and the best solution is represented by the red star. Various confidence intervals are represented by different colored areas. From top to bottom, and from left to right:

- * Top left panels: trade-offs between parameters, as illustrated in 2-dimensions for an easier reading (age vs. inheritance, age vs. erosion, and erosion vs. inheritance).
- * Bottom right panels: calculated χ^2 for the explored range of ages, erosion, and inheritance.

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