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Shortening Rates and Recurrence of Large Earthquakes From Folded and Uplifted Terraces in the Western Danghe Nan Shan Foreland, North Tibet

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► **To cite this version:**

Yanxiu Shao, Jerome van der Woerd, Jing Liu-Zeng, Daoyang Yuan, Yunsheng Yao, et al.. Shortening Rates and Recurrence of Large Earthquakes From Folded and Uplifted Terraces in the Western Danghe Nan Shan Foreland, North Tibet. *Journal of Geophysical Research: Solid Earth*, 2023, 128 (1), 10.1029/2021JB023736 . hal-04049807

HAL Id: hal-04049807

<https://hal.science/hal-04049807v1>

Submitted on 27 Nov 2023

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1 Shortening rates and recurrence of large earthquakes from
2 folded and uplifted terraces in the Western Danghe Nan Shan
3 foreland, north Tibet

4

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13

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16

17 **Key Points:**

- 18 • Combined cosmogenic nuclides, OSL, and ¹⁴C dating methods constrain the
19 deposition age of four inset folded terraces over the last 130 ka.
- 20 • The shortening rate of 0.8 ± 0.2 mm/yr across the western Danghe Nan Shan
21 thrusts is coeval with left-slip decrease along Altyn Tagh fault.
- 22 • Concurrent rupture of thrusts and strike-slip fault strands could result in a Mw 7+
23 earthquake southwest of Subei town.

24

25 Abstract

26 Understanding the three-dimensional structure, segmentation, and kinematic of
27 complex fault systems are essential to assessing the size of potential earthquakes and
28 related seismic hazards. The Danghe Nan Shan, a major thrust splay of the Altyn
29 Tagh fault in north Tibet, is one of these complex faults junction. Near the town of
30 Subei, the western Danghe Nan Shan thrust is composed of two left-stepping faults
31 outlined by fault scarps in front of folded and uplifted alluvial fans and terraces. Age
32 constraints and 2D reconstructions of the accumulated slip above a transient base
33 level of four terraces standing 7 to 60 m above the present stream bed determine the
34 shortening and vertical uplift rates on the southern thrust of 0.5 ± 0.1 mm/yr and $1.1 \pm$
35 0.3 mm/yr, respectively, over the last 130 ka. Along the northern thrust, vertical
36 terrace offsets of 1.5 to 3.6 m and horizontal slip of 4.5 m documented in a
37 paleoseismological trench occurred after 12 ± 4 ka, constraining coeval rates of $0.3 \pm$
38 0.1 mm/yr for uplift and shortening. Overall, 1.4 ± 0.4 mm/yr terrace uplift and $0.8 \pm$
39 0.2 mm/yr shortening rates are determined, in agreement with late Miocene long-term
40 exhumation rate estimates. Our fault mapping together with geomorphic and
41 structural observations imply that the western Danghe Nan Shan accommodates slip
42 transfer from the Altyn Tagh fault to the west to thrusting and shortening farther east
43 in the Qilian Shan region. Considering the size of the scarps, their lateral extent, the
44 geometry of the faults at depth, and their slip-rate, we suggest the possible occurrence
45 of M_w 7+ earthquakes near Subei.

46

47 Plain Language Summary

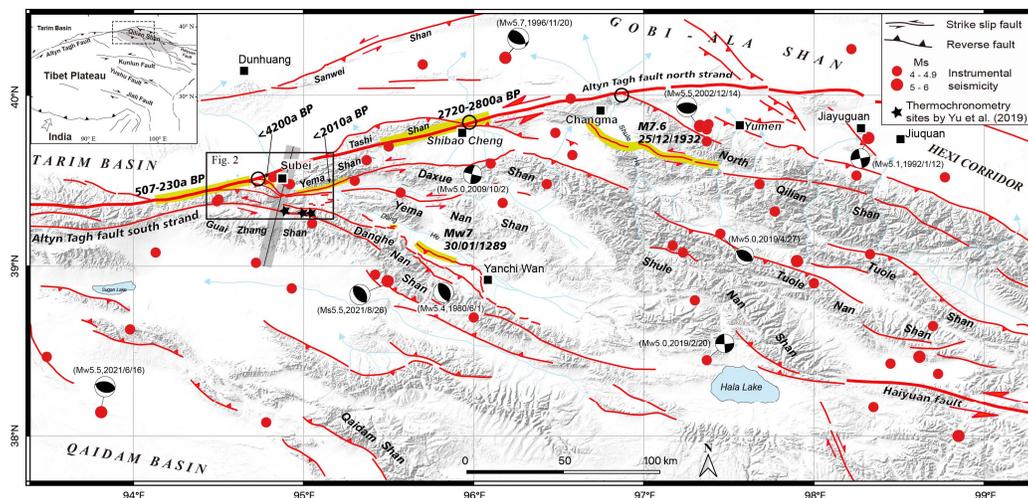
48 How faults slip is essential to understanding continental deformation and seismic
49 hazards. In the western Qilian Shan region, the Danghe Nan Shan is one of the largest
50 ranges bounded by active thrust faults. West and south of Subei, these thrusts can be
51 followed for several kilometers along meter-high scarps and folded terraces.
52 Quantifying the cumulative deformation of the folded and uplifted terraces and fans
53 enables to constrain shortening rates of about 1 mm/yr over the last 130 ka. Evidence
54 from surface geomorphology and sub-surface trenching across the faults indicates the
55 occurrence of large magnitude potentially up to M_w 7 seismic events in the last 16 ka.
56 Integrating age-controlled cumulative geological observations with field-documented
57 co-seismic deformation contributes to increasing the knowledge of slip rate and co-
58 seismic displacement for seismic hazards.

60 **1. Introduction**

61 The repetitive nature of earthquake occurrence is a characteristic of faulting
62 behavior, allowing for seismic hazard assessment. While crustal strain can be
63 concentrated and localized along a single fault, some fault systems are highly
64 segmented and distributed, making it more difficult to constrain each fault's behavior.
65 For instance, it has been shown that the Carrizo plain segment of the San Andreas
66 fault concentrates most of the tectonic plate displacement and with 15 recurrent events
67 in the last 1500 years (e.g., Sieh and Jahns, 1984; Weldon et al., 2004; Liu-Zeng et al.,
68 2006). Strike-slip faults produce long single-trace ruptures, like, for example, along
69 the Kunlun fault, with the Manyi (Mw7.6 1997, Peltzer et al., 1999) and Kokoxili
70 (Mw7.9 14/11/2001, Klinger et al., 2005) events, or along the North Anatolian fault
71 (Stein et al., 1997). Thrust or combined strike-slip/thrust events are sometimes more
72 complex, like for instance, the Wenchuan earthquake (Mw7.9 2008; Xu et al., 2009;
73 Hubbard et al., 2010) with two obliquely slipping major thrusts, or the Kaikoura
74 earthquake (Mw7.8 2016; Shi et al., 2019) with a complex combination of thrusts and
75 strike-slip faults or may even stay blind as the recent Himalaya Gorkha earthquake
76 (Mw7.8 2015; Hubbard et al., 2016). It thus requires quantification of fault geometry,
77 activity, and rupture history along adjacent and intersecting fault networks to
78 understand how such complex rupture occurs.

79 The Qilian Shan region has long been recognized as a crustal-scale restraining
80 step-over between the two northernmost major strike-slip faults at the margin of the
81 Tibetan plateau (Tapponnier and Molnar, 1977; Meyer et al., 1998; Zuza et al., 2018).
82 Left-lateral slip along the N70°E-striking Altyn Tagh fault (ATF) is transferred to the
83 N100°E-striking Haiyuan fault through a set of thrusts and strike-slip faults
84 distributed within the Qilian Shan region. The complex links between strike-slip and
85 thrust faults have been highlighted in some historical earthquakes, such as the
86 Changma event of 1932 (Meyer, 1991; Xu et al., 2010) or the Gulang earthquake of
87 1927 (Gaudemer et al., 1995; Guo et al., 2020). How thrust faults connect at depth to
88 accommodate strain responsible for folding and mountain building may allow us to
89 assess the rupture area at depth better and thus the magnitude of past and future
90 earthquakes (e.g., Stockmeyer et al., 2014 ; Shi et al., 2019).

91

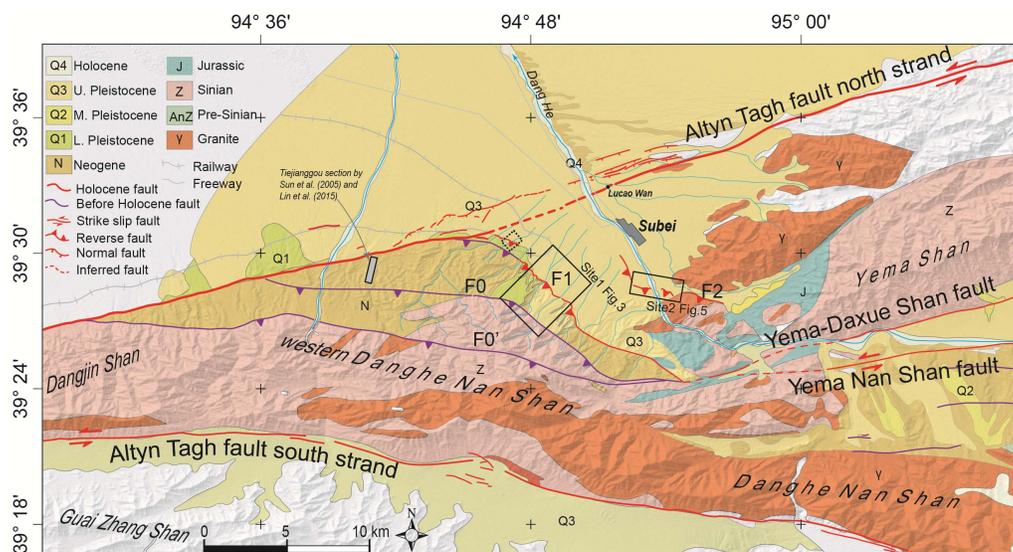


92
 93 Figure 1. Active faults map of northwestern Qilian Shan region (modified from
 94 Tapponnier et al., 1990; Meyer et al., 1998; Van der Woerd et al., 2001). Inset shows
 95 location of area in simplified tectonic map of Tibet. Seismicity (magnitude >4) from
 96 China Earthquake Networks Center (CENC) catalog (1970 – 2021). Focal
 97 mechanisms from GCMT (1970 – 2021). Yellow shaded fault traces with dates are
 98 historical and most recent paleo-earthquake rupture extensions from Meyer (1991),
 99 Xu et al. (2010), Li et al. (2016), Shao et al. (2017), Wu et al. (2017), He et al. (2018)
 100 and Luo et al. (2019). Gray shaded area with black line is position of swath profile in
 101 Figure 12. Black circles are major triple junctions along Altyn Tagh fault, Subei,
 102 Shibao Cheng/Changma, and Qilian junctions, from west to east.

103

104 Along the northwestern Qilian Shan region margin, splays from the Altyn Tagh
 105 fault veer eastward and connect with the ranges of Danghe Nan Shan (DHNS), Daxue
 106 Shan, and north Qilian Shan, in strike-slip/thrust fault triple junctions (e.g., Meyer et
 107 al., 1998; van der Woerd et al., 2001; Xu et al., 2005; Figure 1). The 250 km-long and
 108 20 km-wide DHNS with peaks reaching 5500 m asl, i.e., 4000 m above the Tarim
 109 basin to the west and 2000 m above the average elevation of the Qaidam basin, is one
 110 of the largest ranges of the Qilian mountains. Its western junction with the ATF, west
 111 of the town of Subei, is particularly complex and involves the north and south strands
 112 of ATF and several active thrust faults along the northern piedmont of DHNS (Figure
 113 2).

114



115
 116 Figure 2. Active fault map at Subei junction from field observations, satellite image
 117 interpretation, and topographic data (modified from van der Woerd et al., 2001; see
 118 Figure S1). Background is simplified geology from Gansu Bureau of Geology and
 119 Mineral Resources of Gansu Province (1989). Main thrusts F0', F0, F1, and F2 are
 120 responsible for western Danghenan Shan growth southwest of Subei basin. Open
 121 rectangles locate study sites 1 and 2 (Figures 4 and 8). Dashed line rectangle is site 3
 122 of van der Woerd et al. (2001). Grey rectangular box locates palaeomagnetic section
 123 in Sun et al. (2005) and Lin et al. (2015).
 124

125 In this study, we investigate the geometry and the kinematics of active faults at 2
 126 sites along the ~20 km-long western front of the DHNS (sites 1 and 2, Figure 2).
 127 Based on topographic profiles of deformed and uplifted terraces, we constrain the
 128 sub-surface geometry of the thrust faults and estimate the amounts of uplift and
 129 shortening using 2D area balance. We combine several geochronological tools, such
 130 as ^{14}C , optically stimulated luminescence (OSL), and terrestrial cosmogenic nuclide
 131 (TCN) to date the various terrace levels. Finally, seismic scarp heights and co-seismic
 132 slip in trenches, combined with the inferred thrust geometry at depth make it possible
 133 to estimate the magnitude of surface rupturing earthquakes generated by the thrust
 134 faults.

135

136 2. Tectonic setting

137 The Qilian Shan region at the northern margin of the Tibetan Plateau (Figure 1).
 138 is characterized by NW-SE trending thrust and fold belts. These ranges accommodate
 139 shortening and confine intermontane basins leading to mountain growth and elevation
 140 increase (Burchfiel et al., 1989; Métivier et al., 1998; Meyer et al., 1998; Tapponnier

141 et al., 1990). Deformation in the Qilian Shan in the Cenozoic results from the
142 collision between Eurasia and India, according to geological structures,
143 palaeomagnetic data and sedimentary records in the Qaidam-Qilian area (e.g., Dai et
144 al., 2006; Dupont-Nivet et al., 2004; Horton et al., 2004; Yin et al., 2002; Yin et al.,
145 2008; Zhou et al., 2006; Zhuang et al., 2011). The Qilian Shan grew during late
146 Oligocene – middle Miocene time (e.g., George et al., 2001; Gilder et al., 2001;
147 Métivier et al., 1998; Zheng et al., 2010) as suggested by magnetostratigraphy, apatite
148 fission-track (AFT) and (U-Th)/He cooling data, and sedimentary mass balance. This
149 growth accelerated during the Pliocene (Métivier et al., 1998) and continued until
150 Quaternary time as evidenced by active thrusting (e.g., Tapponnier et al., 1990; Meyer
151 et al., 1998; Zheng et al., 2013; Hetzel et al., 2004, 2019; Hu et al., 2015, 2021; Du et
152 al., 2020; Xu et al., 2021).

153 In the western DHNS piedmont, a conglomerate coarsening in the Tiejianggou
154 Formation occurred at 13.7 Ma as determined from palaeomagnetic data (Sun et al.,
155 2005 ; Figure 2). Both changes in sedimentary facies and variations in depositional
156 rates suggest increased tectonic uplift between 13.7 and 9 Ma (Sun et al., 2005).
157 Similarly, the AFT and U-Th/He data obtained from sections of Cenozoic terrestrial
158 clastic sediments and bedrock of Paleozoic granodiorites along the western DHNS
159 (see location in Figures 1 and 2) are consistent with a change in exhumation rate
160 around 14 Ma, implying intensification of uplift at that time (Lin et al., 2015; Yu et al.,
161 2019).

162 Thrust-fault scarps in the Subei and Yanchi Wan basins, along the western and
163 central DHNS piedmont, respectively, testify to active shortening and surface
164 rupturing earthquakes (Meyer et al., 1998; van der Woerd et al., 2001; Shao et al.,
165 2017; Xu et al., 2021; Figure 1). Thrusting along the western DHNS southwest of
166 Subei is kinematically compatible with a compressive right step between the north
167 strand of ATF and the Yema-Daxue Shan and Yema Nan Shan left-lateral strike-slip
168 faults (Figures 1 and 2). The kinematics of triple junctions predicts that the ATF slip-
169 rate decreases northeastward past the different junctions (e.g., Meyer et al., 1998) as
170 observed by geological and geodetic slip rates distribution along the ATF (Figure S2;
171 Mériaux et al., 2005; Xu et al., 2005; Zhang et al., 2007; Seong et al., 2011; Xiao et
172 al., 2017; Elliott et al., 2018; Liu et al., 2020). Locally, other rates have been
173 determined on thrust or strike-slip faults (van der Woerd et al., 2001; Luo et al., 2015;
174 Zheng et al., 2013; Xu et al., 2021).

175 A few paleoseismic trenches have been dug along the DHNS thrust and adjacent
176 strike-slip faults. West of the Subei junction, Li et al. (2016) described evidence for
177 four events from a trench dug across the Aksay segment of the ATF. They constrained
178 the ages of two events at 1180 a B.P. and 507 – 230 a B.P. To the east, three events
179 are identified in trenches near Shibao Cheng over the last 7000 years, with the most
180 recent one between 2720 and 2800 a B.P. (Luo et al., 2019). Along the western DHNS
181 (site 3 in Figure 2), a thrust rupture with a vertical offset of ~ 1.7 m after 4300 a B.P.
182 is documented (Wu et al., 2017). Along the central DHNS in the Yanchi Wan basin
183 (Figure 1), Shao et al. (2017) used high-resolution topographic data and trenches to
184 describe a late event that ruptured between A.D. 1289 and 1395, with horizontal
185 shortening of ~ 1.3 m, inferred to be the A.D. 1289 event that occurred during the
186 Yuan Dynasty according to historical records. Along the Yema-Daxue Shan fault, He
187 et al. (2018) excavated three paleoseismic trenches and found evidence of three events
188 since 7000 a B.P., with the latest event occurring after 2010 ± 30 a B.P.

189 Recent seismicity in the western Qilian Shan region is mostly characterized by
190 moderate earthquakes (Figure 1). For instance, nine earthquakes with magnitudes of
191 $M4 - 6$ have occurred along the DHNS since 1960. The focal mechanisms of these
192 earthquakes are both thrust and strike-slip. A $M5.5$ earthquake occurred recently on
193 26 Aug. 2021 with an epicenter along the southern piedmont of DHNS. Its focal
194 mechanism is similar to the $Mw5.4$ 1980 event and, given the uncertainty in
195 hypocenter depth and location, compatible with faulting along north-dipping thrusts
196 subparallel to the range. The largest known event in the area is the 25 December 1932
197 $M7.6$ Changma earthquake (Figure 1). The surface rupture of this earthquake extends
198 for 120 km with coseismic horizontal displacements up to 5.5 m (Du et al., 2020;
199 Meyer, 1991; Peltzer et al., 1988). The potential for a large earthquake is suspected in
200 the Subei basin along the DHNS as large several-meter-high seismic scarps are
201 evidenced (see below).

202

203 **3. Methods**

204 3.1. Shortening constraints from uplifted area balancing

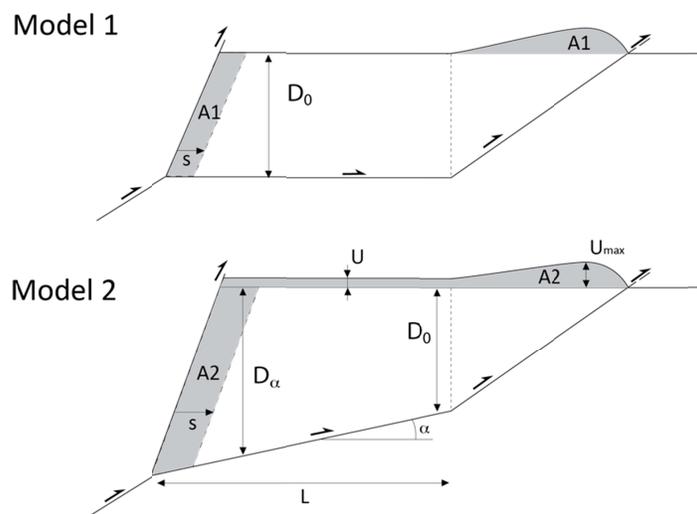
205 Abandoned, uplifted, and folded terraces may be used to constrain the kinematics
206 of shortening over the time span of the terrace ages (e.g., Meyer et al., 1998;
207 Rockwell et al., 1998; Lavé & Avouac, 2000; van der Woerd et al., 2001; Thompson
208 et al., 2002; Daëron et al., 2007; Simoes et al., 2007; Hu et al., 2015, 2021; Guilbaud

209 et al., 2017; Xu et al., 2021). Provided that the base levels can be reconstructed, the
210 present height of the strath terraces above their base level is an estimate of the
211 tectonic throw on the underlying thrust fault (e.g., Bull, 1991; Merritts et al., 1994;
212 Lavé and Avouac, 2000; van der Woerd et al., 2001; Thompson et al., 2002; Wang et
213 al., 2020). Using a mass conservation model (Figure 3) and assuming negligible
214 erosion of the terrace, it is possible to estimate the amount of corresponding slip on
215 the thrust fault at depth (S) to account for uplift and folding of the strath level (areas
216 A1 and A2).

217 We infer the geometry of the thrust fault at depth from the surface geology and
218 structural data, such as bedding of the sedimentary beds or fault dip, similarly to
219 Meyer et al. (1998), van der Woerd (1998), or Hu et al. (2015). We consider a simple
220 geometry of the thrust fault at depth made of two connected planes, a sub-horizontal
221 décollement, and a ramp (Figure 3). The dip of the frontal ramp without surface
222 observation of the emerging thrust fault, may range between $45\pm 15^\circ$ as commonly
223 observed or inferred from structural data, i.e., bedding of the thrust sediment layers
224 (e.g., van der Woerd et al., 2001; Gilder et al., 2001). The wavelength of the
225 antiformal folding of the terraces may be used to constrain the depth where the ramp
226 and the décollement meet (D_0 ; Figure 3, e.g., Meyer et al., 1998). The precise dip (α)
227 or rooting depth (D_α) of the décollement is unknown, thus we use two end-member
228 models, where the dip of the “flat” changes from near 0° (Model 1) to a certain dip
229 (Model 2, $0 < \alpha < 45^\circ$; Figure 3). In Model 1, this depth remains constant until it
230 eventually joins with other faults at depth. In Model 2, the décollement dips with an
231 angle α and thus connects with other faults deeper (Figure 3).

232 In the case where several terrace levels are incrementally uplifted and folded, and
233 without independent structural constraints on the geometry of the décollement, it is
234 possible to adjust the dip angle α and thus depth D_α by minimizing the shortening rate
235 variations among the terraces. A sloping décollement has two consequences. First, the
236 dipping décollement acts like a shallowly dipping ramp with uniform uplift (U in
237 Model 2, Figure 3) above its entire length (e.g., Meyer et al., 1998; Hu et al., 2015).
238 Second, increasing D_α reduces the slip amount (S) necessary to account for the
239 uplifted area (A2) above base level.

240



242

245 Figure 3. Simplified geometry of ramp and décollement in models 1 and 2 with
 246 corresponding parameters (see text). Décollement remains flat in Model 1 while it
 247 dips with angle α in Model 2.
 246

247 3.2. Topographic survey of uplifted landforms

257 We used different tools to survey the topography of the terrace surfaces and
 258 associated riverbeds. At site 1, we levelled 5 – 1600 m-long differential global
 259 position system (DGPS) profiles. The rover of the GPS instrument records one
 260 position every second. After post-processing, the horizontal and vertical accuracies of
 261 the position are approximately 5 and 10 mm, respectively. These profiles are
 262 combined with 4 profiles surveyed in 1996 with a digital-recording theodolite-
 263 distancemeter with similar sub-cm precision (see details in van der Woerd et al., 2001;
 264 [Figures 2, 4 and 5](#)). Given the precisions of both instruments, the roughness and
 265 irregularities of the several-meter loess-covered terrace surfaces or cobble-rich
 266 riverbeds are the primary sources of errors.

262 We used a drone-based laser scanning to survey the topography at site 2 ([Figures](#)
 263 [6 and 7](#)). The drone has eight rotors with a VLP16 Velodyne[®] laser sensor flying at
 264 ~ 50 m above the surface. Millions of cloud points are processed to produce a 0.5 m-
 265 resolution digital elevation model (DEM). Vegetation was removed by point
 266 classification ([Figure 6c](#)).

265 Chinese ZY-3 satellite stereo-pair images, with a resolution of ~ 2.5 m (acquired
 266 in December 2017), were used to generate a DEM with a horizontal and vertical
 267 resolution of ~ 6 m across the Subei basin ([Figure S1](#)).

266 3.3. Paleoseismic trenching

266 To document the subsurface fault geometry and rupture history of the seismic
267 scarp at site 2, we investigated the subsurface sedimentation and structure by opening
268 a trench across the main fault scarp (Figures 6 and 7). The exposed walls in the trench
269 are cleaned and photographed with a digital camera (Figure 8). We used the structure-
270 from-motion method to mosaic the photos (e.g. Bemis et al., 2014; Reitman et al.,
271 2015) before mapping the deformed and offset sedimentary units onto the orthophotos.

272 3.4. Sampling and dating methods

273 To constrain the age of the deformed terraces and fans, we combined three
274 different dating techniques: terrestrial cosmogenic nuclide (TCN) dating, accelerator
275 mass spectrometry for radiocarbon samples (AMS ^{14}C) and optically stimulated
276 luminescence dating (OSL).

277 AMS ^{14}C and OSL were used to constrain the ages of sandy or silty sediments
278 found either capping the coarser conglomeratic material of the alluvial deposits or
279 sandy lenses in coarser sediments. The radiocarbon samples were processed and dated
280 at Beta Analytic. Ages were calibrated with IntCal13 (Reimer et al., 2013). All OSL
281 samples were processed and measured following the procedure described in Shao et al.
282 (2017). Fine-grained (4 – 11 μm) quartz fractions were prepared and treated with 30%
283 hydrogen peroxide (H_2O_2) and 10% hydrochloric acid (HCl) to remove organic
284 materials and carbonates, respectively. The purity of the prepared quartz was checked
285 by the ratio of infrared-stimulated luminescence (IRSL) to blue light-stimulated
286 luminescence (BLSL) (<5%). We used a Risø TL/OSL-DA-20 reader equipped with a
287 $^{90}\text{Sr}/^{90}\text{Y}$ beta source (~ 0.12 Gy/s), blue LEDs (470 \pm 20 nm, ~ 80 mW/cm 2), and two
288 Hoya U340 filters (290 – 370 nm) to conduct irradiation, heating, and luminescence
289 measurements. A simplified multiple aliquot regenerative dose (SMAR) was used to
290 measure the quartz aliquots.

291 At site 1, depth profiles are dug into the loess covered surfaces of the terraces in
292 relatively flat places away from any visible surface erosion or incised channels
293 (Figure 4), except for one terrace, where the depth profile was sampled by refreshing
294 its steep eastern riser (Figure 9). From each profile, we collected 6 – 8 samples of
295 quartz-rich gravels in the top 2 m of the several meter-thick gravels/cobbles deposits
296 (Figures 9 and 10). The age of abandonment of the underlying conglomerate is then
297 determined by ^{10}Be depth profile modeling considering shielding of the overlying
298 loess cover, which age is constrained with ^{14}C and OSL dating, following the
299 approach developed in Shao et al. (2021). At site 2, only surface samples were

300 collected from a boulder fan surface. Six 2.5 cm-thick rock samples were collected
301 with a chisel and hammer from the top of six embedded granitic boulders (size > 0.5
302 m in diameter) (Figure 11).

303 The TCN samples were crushed and sieved, allowing the 250 – 500 μm fraction
304 to be treated by sequential ultrasonic leaches to obtain pure quartz. Beryllium oxide
305 was extracted by chromatography from the sample solution and inserted into target
306 holders at the Cosmogenic Isotope Laboratory of Institut Terre et Environnement de
307 Strasbourg (ITES UMR7063). The isotopic ratios were determined at the ASTER-
308 AMS facility at the Centre Européen de Recherche et d'Enseignement des
309 Géosciences de l'Environnement (CEREGE, Aix-en-Provence, France). We used the
310 online calculator of CRONUS-Earth calculator Version 3 (Balco et al., 2008) to
311 calculate the exposure ages of surface samples with the time-dependent scaling model
312 of Lal (1991)/Stone (2000). For subsurface samples, we modeled the ages based on
313 the depth dependence of the concentration (e.g., Braucher et al., 2003; Gosse &
314 Phillips, 2001; Lal, 1991; Yang et al., 2019; Shao et al., 2021). The terrace
315 conglomerate densities, which depend on composition, grain size and consolidation of
316 terrace sediment were estimated from the exponential best fit of the depth profiles.
317 We performed forward modeling to estimate the shielding effect of loess considering
318 a density of 1.4 g/cm^3 and constant accumulation rates calculated by linearly fitting
319 the age/depth relationship from up to three OSL and/or ^{14}C dates (similarly to other
320 studies in the Qilian Shan area, e.g., Hetzel et al., 2004, 2019; Shao et al. 2021).

321

322 4. Results

323 The study sites 1 and 2 are situated in the DHNS bajada south of Subei along the
324 main thrusts F1 and F2 (Figure 2). The first site (site 1) is located in the middle of the
325 uplifted ledge at a place already described in van der Woerd et al. (2001) and
326 composed of a set of four main uplifted and folded terrace levels by thrust F1 (Figure
327 4). The second site (site 2) is located 10 km to the east along the fault scarp of F2
328 crossed by the Dang He river.

329

330 4.1. The geometry of uplifted and folded alluvial surfaces

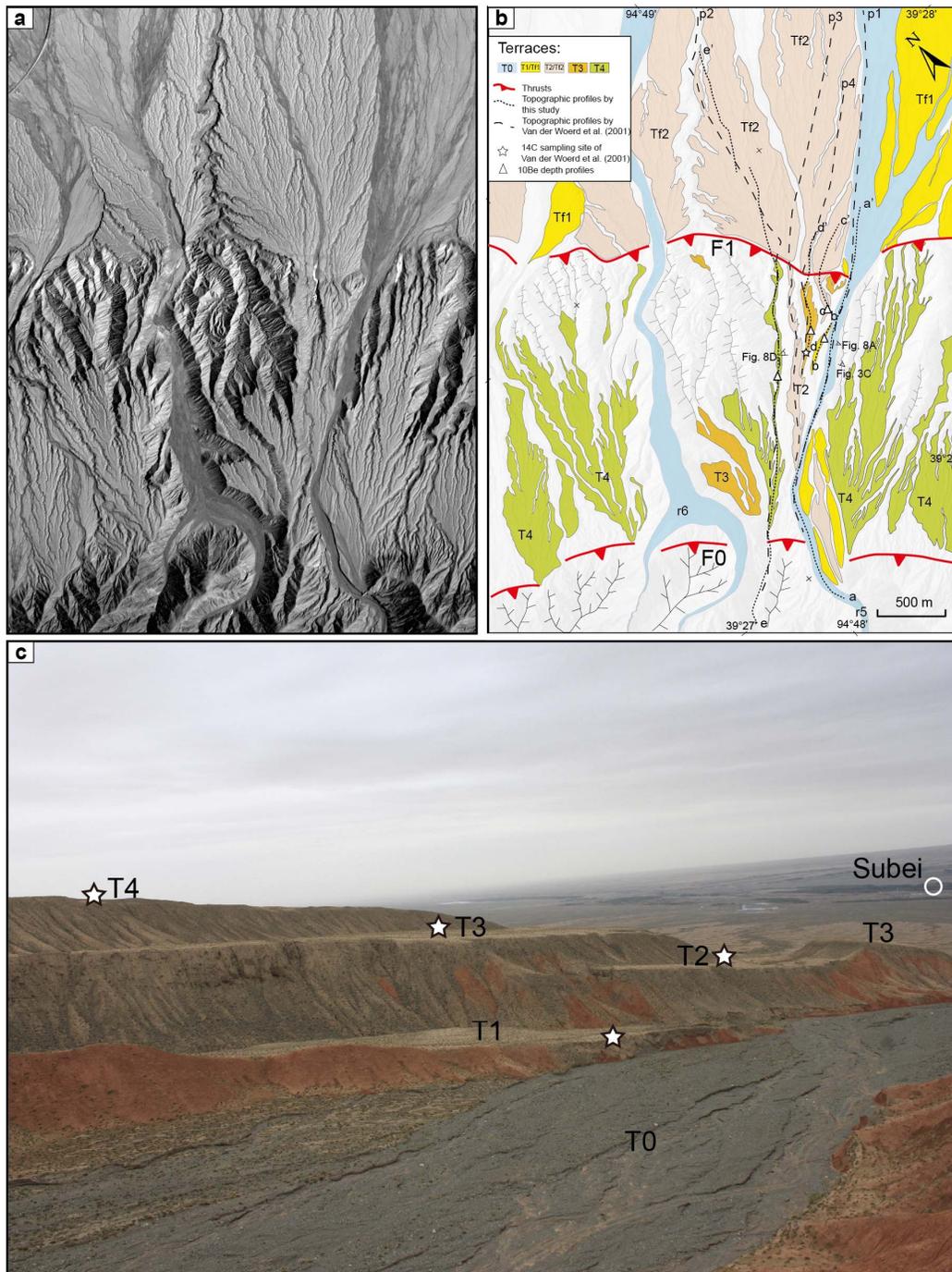
331 4.1.1. Site 1

332 Southwest of Subei, thrust F1 sharply disrupts the bajada to form a NW-SE
333 trending topographic scarp. Upstream of the scarp, uplifted reaches of late Pleistocene

334 strath terraces are deposited unconformably above Tertiary reddish sandstones and
335 conglomerates. Variably thick loess covers the sediments of the terraces. Two to three
336 kilometers further southwest, the terraces are overthrust by the steeply faceted
337 Paleozoic bedrock of the DHNS along thrust F0 (Figure 2; van der Woerd et al.,
338 2001). As a response to recent and active uplift, numerous streams have grown into
339 subparallel catchments of relatively small size, deeply incising across the uplifted
340 ledge and depositing alluvial fans in front of thrust F1. While older deposits are
341 preserved upstream as narrow terrace remnants, downstream, younger deposits have
342 buried the older sediments (van der Woerd et al., 2001).

343 Site 1 is located in the central part of thrust F1 along a stream of one of the
344 largest catchments along the range front (stream r5 in van der Woerd et al., 2001). A
345 set of four terraces (T1 to T4) are preserved upstream of F1, whereas on the
346 downstream side, only one main fan level (Tf2) is present (Figure 4). To constrain the
347 geometry of uplifted and folded terraces upstream of F1, a set of topographic profiles
348 has been leveled along different terraces on the western bank of stream r5 (Figure 4).
349 All profiles are projected in a N45°E direction perpendicular to the average N135°E
350 strike of thrust F1 in Figure 5.

351

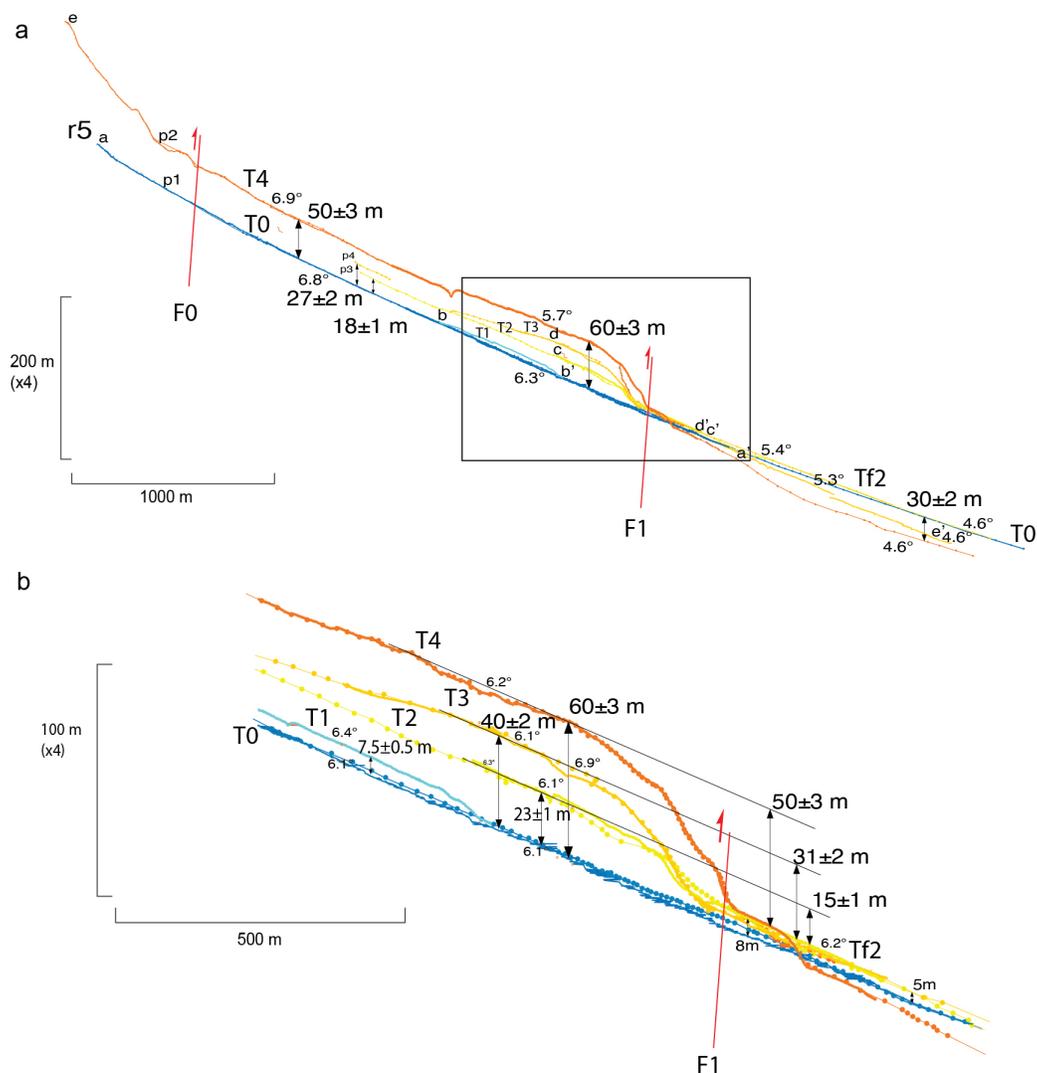


352

353 Figure 4. Uplifted and folded terraces along fault F1 at site 1. (a) ZY3 satellite image
 354 of site 1. (b) Geomorphic interpretation from image in A and field observations. South
 355 of main thrust fault F1, four main terrace levels T1 to T4 are abandoned by stream
 356 r5. Dashed and dotted lines are topographic profile locations acquired with total station
 357 (van der Woerd et al., 2001) and DGPS (this study), respectively (see Figure 5).
 358 Triangles are pit locations for cosmogenic isotope depth profile dating. Star is
 359 charcoal sample location in van der Woerd et al. (2001). (c) Field view to north of
 360 uplifted and folded terraces at site 1 south of Subei. Terraces are loess covered
 361 several-meter-thick greyish conglomerate unconformable on red-orange south dipping

362 Tertiary sandstones and conglomerate. Stars locate sampling pits in terraces T1, T2,
363 T3 and T4.
364

365 From the range front in the south to the bajada in the north, i.e., between F0 and
366 F1 (Figure 4), all terraces show a similar pattern of uplift and deformation. Upstream
367 they stand above the present stream bed with a slightly steeper slope as T0, while they
368 form a clear antiform near their northern end (Figure 5). This shape is particularly
369 clear for T4 and T2, which treads extend largely upstream, while for T1 and T3 only
370 the folded bulge to the north remains. Minimum and maximum heights of terraces T1,
371 T2, T3 and T4 above the present-day stream bed are 0 , 18 ± 1 , 27 ± 2 , and 50 ± 3 m
372 and $7. \pm 0.5$, 23 ± 1 , 40 ± 2 , and 60 ± 3 m, respectively (Figure 5; Table 1). From our
373 mapping (Figure 2), the bajada in front of the fold are fan deposits (Tf2)
374 corresponding mostly to terrace level T2. The corresponding maximum offset of the
375 surface T2 relative to the Tf2 bajada is 15 ± 1 m (Table 1).



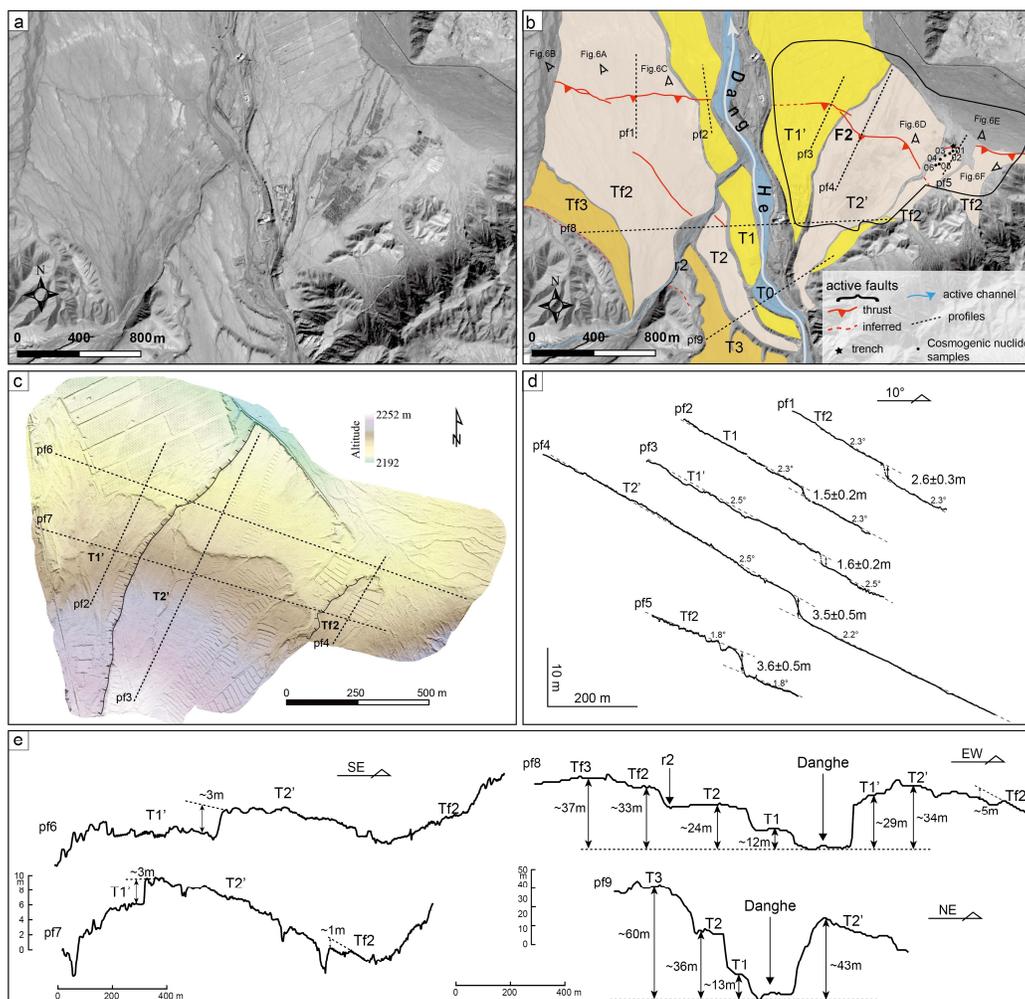
376

377 Figure 5. Uplift measurements at site 1. (a) N45°E projected topographic profiles (see
 378 location in Figure 4, same labels). Terrace heights measured above active stream bed
 379 T0. Main thrust position inferred. (b) Detail of folded terraces at ledge front. Fan Tf2
 380 in footwall corresponds to T2 in the hanging wall. Terrace maximum heights
 381 measured above active stream bed T0 in hanging wall and above Tf2 in footwall.
 382

383 4.1.2. Site 2

384 Site 2 is located at the bottom of the Subei basin along the Dang He river-bed,
 385 approximately 7 km east of site 1 (Figures 2, 5, 6). There, thrust F2 can be traced for
 386 approximately 10 km outlined by a series of half-meter- to several-meter-high scarps.
 387 To the northwest, the scarps, across the distal fans of the main bajada, are subparallel
 388 to F1 for 2 km (Figure 2). The scarps then veer eastward to strike east-west on
 389 average across a set of fans abandoned by the Dang He and a major tributary. Site 2 is
 390 located where the scarps reach up to several meters high. Farther east, the scarps

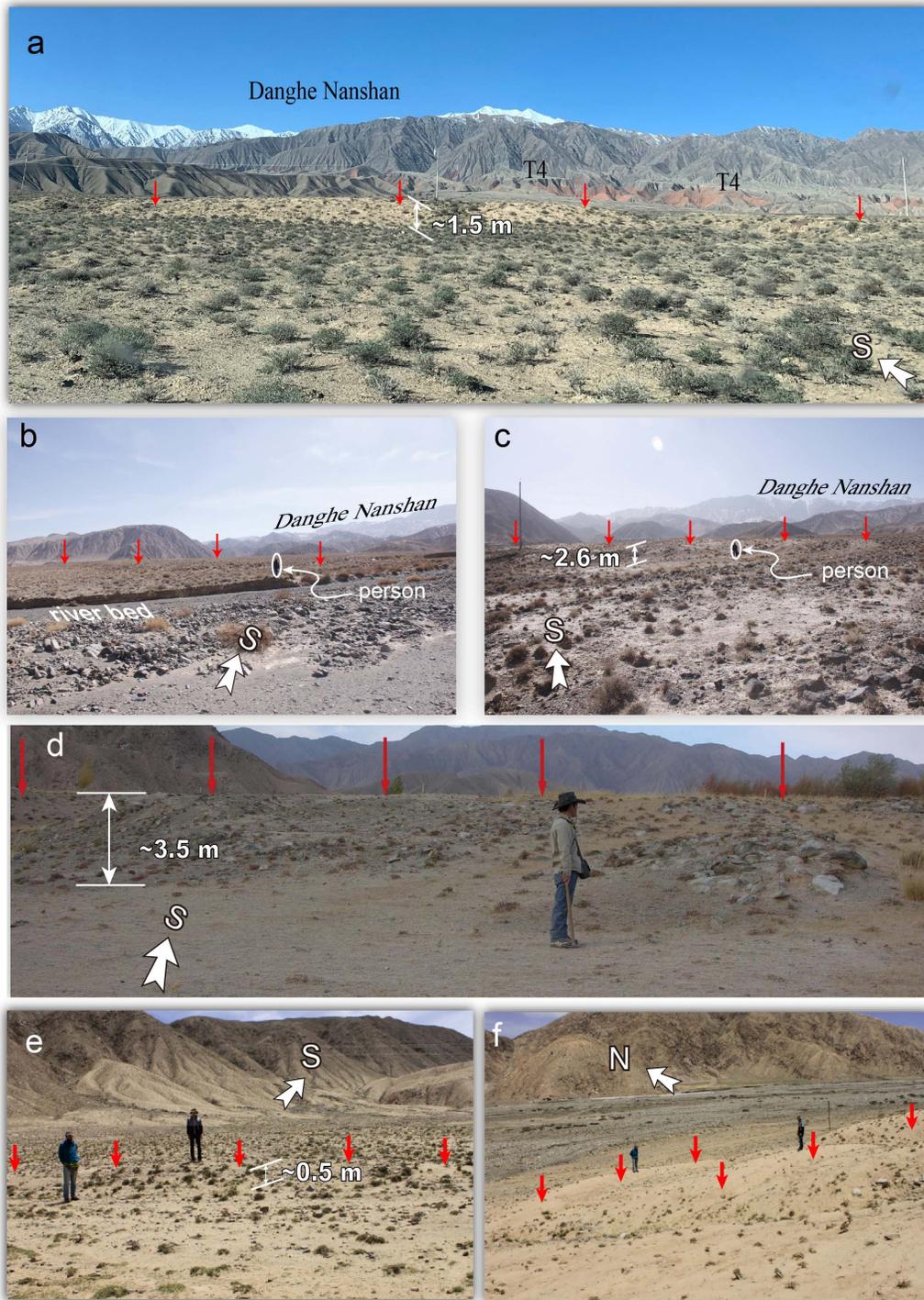
391 continue along the southern rim of a major valley within the Yema Shan, and their
 392 traces become unclear in the high relief eastward.



393
 394 Figure 6. Geomorphic interpretation of site 2. (a) ZY3 satellite image of site 2. (b)
 395 Geomorphic map. Black outline is drone based Lidar survey in (c). Dang He has
 396 abandoned a set of terraces T1, T1', T2, T2' and T3 as it leaves its gorge across
 397 southwest Yema Shan. Tf2 and Tf3 are alluvial fans from smaller tributaries. Terraces
 398 and fans are cut and offset by active thrust F2 (in red). (c) High resolution (0.5 m)
 399 DEM from drone based Lidar survey. Dashed lines are topographic profiles in (d) and
 400 (e). Solid lines represent terrace riser tops, ticks are on lower side. (d) Topographic
 401 profile across F2 scarps. Piercing lines (dashed) fit far field slopes of surfaces. (e) Left,
 402 topographic profiles extracted from ~6 m resolution DEM, and right, from drone
 403 DEM, located in (b) and (c), respectively.

405 The Dang He and tributaries at site 2 incise in a set of imbricated fans, the oldest
 406 and highest fans (T3, Tf3) standing up to 60 m above the riverbed near their apex or
 407 where the Dang He leaves the gorge to the east (profiles pf8, pf9; Figure 6). Their
 408 slopes decrease progressively from the oldest to youngest. Presently, F2 thrust scarps

409 cut through the lower parts of Tf3, Tf2, T1 and T2'. Across Tf2 and T2', the scarps
 410 heights range from 2.6 to 3.5 m, while across T1, the scarp is approximately 1.6 m
 411 high (Figure 6D).



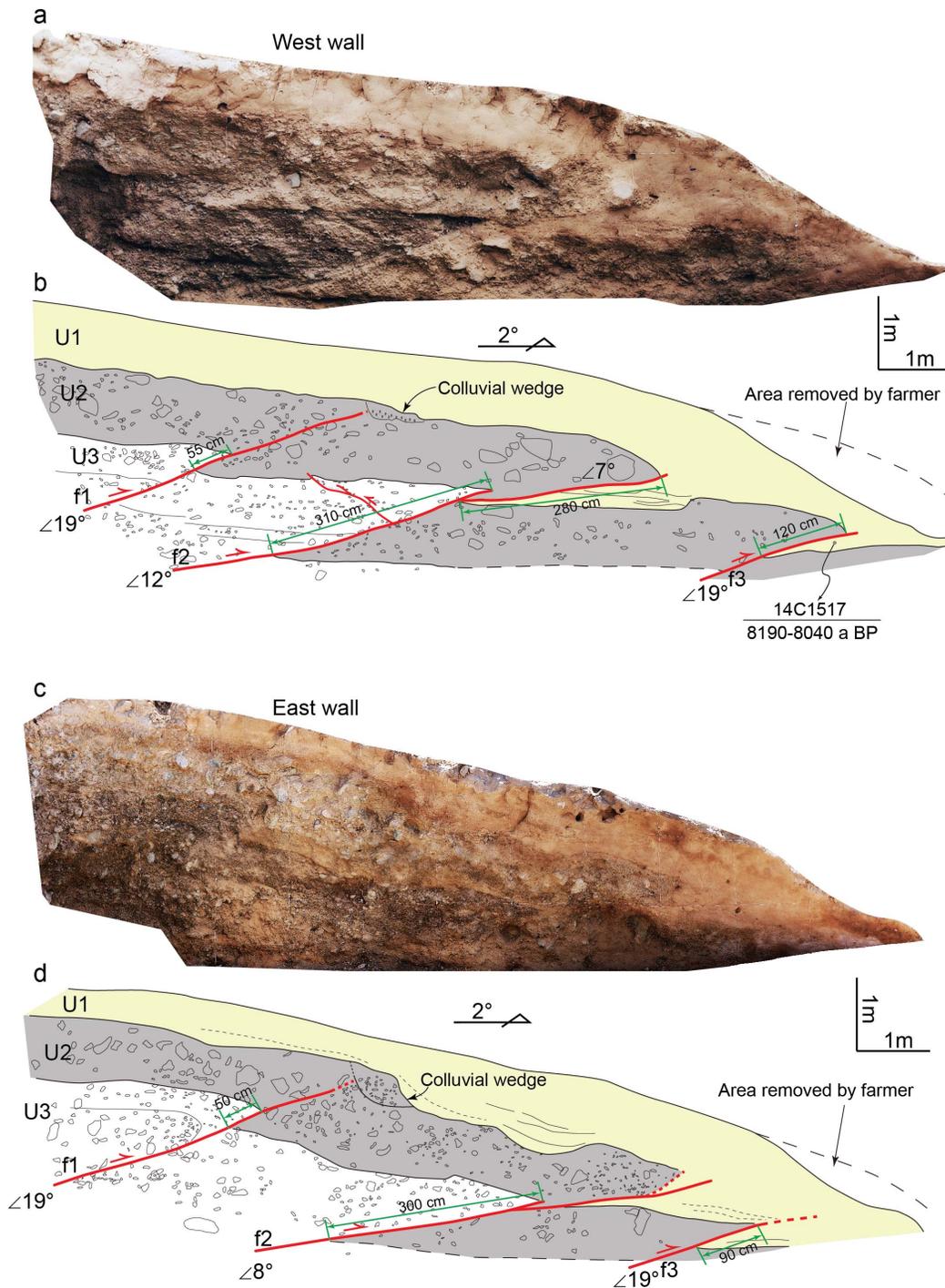
412
 413 Figure 7. Field views of faulted alluvial fans and terraces by thrust F2 at site 2 (See
 414 Figure 6 for locations). Small arrows indicate top of fault scarp. Person is ~1.7 m tall..
 415 (a)-(c) ~1 – 2 m-high scarp west of Dang He. (c) ~2.6 m-high scarp across Tf2. (d)

416 ~3.5 m-high scarp across Tf2. (e) ~0.5 m-high scarp across Tf2. (f) Multiple scarps
417 across 4 m-high faulted and folded terrace front.
418

419 The frontal 3.5 ± 0.2 m-high scarp on surface Tf2 (profile pf5; Figure 6D) was
420 targeted for trenching (Figure 8). Orthophotos of the two 13 m-long and 3.5 m-high
421 trench walls exhibit similar units and faulting evidence. Three main units can be
422 separated. Unit 1 (U1) is the top layer of approximately 1 m-thick silty loess thinning
423 upstream, underlain by a 3 – 5 cm thick bed of reddish coarse sand at the bottom. Unit
424 2 (U2) consists of approximately 1.5 m-thick light gray poorly sorted angular sandy
425 gravels. Unit 3 (U3) is a layer of gray, poorly sorted angular sandy gravels.

426 Three main shallow-dipping (7° to 19°) thrust faults (f1-f3) can be seen on both
427 walls cutting through units U1 to U3, with a minor back thrust splaying from f2
428 (Figure 7c). All three faults offset the top of unit U2; faults f2 and f3 also offset the
429 loess unit U1. As a result, the bottom of U2 is displaced ~50 – 55 cm and ~300 – 310
430 cm by f1 and f2, respectively. The top of U2 is displaced ~50 cm, ~280 cm, and ~90 –
431 120 cm by f1, f2, and f3, respectively.

432 Trench exposures show evidence of two events. The older event is indicated by a
433 gravelly colluvial wedge in the footwall of fault f1, occurring before the deposition of
434 unit U1. The younger event is manifested by faults f2 and f3, which offset the base of
435 unit U1. The upward extending traces of faults f2 and f3 are unclear in the
436 homogeneous loess unit. Therefore, it is difficult to determine whether there is more
437 than one event after the deposition of the lower part of unit U1.



438

439 Figure 8. Trench walls photo mosaic (a, c) and interpretation (b, d) at Site 2 (see
 440 location in Figure 5). East wall (c and d) is mirrored. Three main units U1, U2 and U3
 441 are faulted. Solid and dashed thin black lines are minor layers. Red lines f1, f2 and f3
 442 are shallow dipping faults, dashed when inferred. Offsets of units are indicated in
 443 green. Charcoal sample and calibrated age is indicated (Table 4). See text for unit
 444 details.
 445

446 Taking a 20° dip, the summed dip slip of approximately 4.9 ± 0.2 m of the thrusts

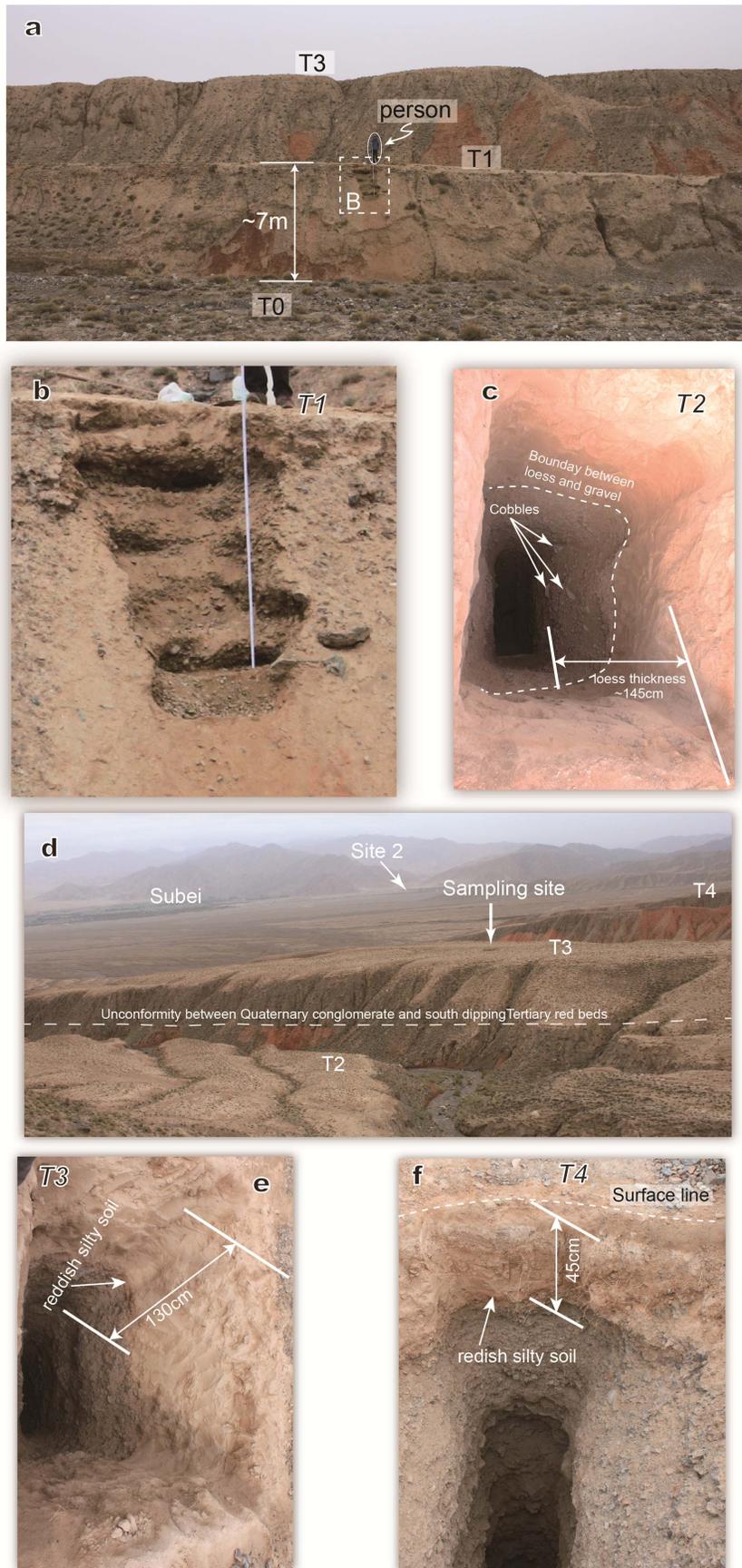
447 in the trench would account for 1.7 ± 0.2 m of uplift. This value is less than half of the
448 3.6 ± 0.2 m height of the frontal scarp of terrace Tf2 (profile pf5 in Figure 6). For
449 these faults and slip amounts to be responsible for the 3.6 ± 0.2 m-high cumulative
450 scarp of Tf2, they must steepen rapidly south of the trench to approximately 47° .

451

452 4.2. Terrace age determination

453 At site 1, we model the exponential depth distribution of ^{10}Be concentrations of
454 amalgamated samples to determine the exposure ages. To average the variable
455 inheritance among the individual rock pieces, our samples include a minimum of 60
456 distinct pebbles, a number usually considered large enough for this purpose (e.g.,
457 Anderson et al., 1996; Hancock et al., 1999; Matrau et al., 2019; Shao et al., 2021).
458 Nevertheless, our depth profiles show scattered ^{10}Be concentrations with depth not
459 simply modeled by an exponential fit. With scattered values largely above analytical
460 errors and in the absence of vertical mixing, several reasons could be responsible for
461 the dispersion observed, such as sequential deposits or not well-averaged inheritance
462 (e.g., Perrineau et al., 2011; Le Dortz et al., 2012). There are no field evidence for
463 sequential deposits or unconformities within the conglomerates. Consequently, in the
464 case of not well-mixed inheritance, the less concentrated samples in the profile are
465 likely closer to the actual terrace exposure time (e.g., Le Dortz et al., 2012). Figure 10
466 shows modeling fitting the less concentrated samples. Model solutions showing an
467 average fit of all samples are also presented in Figures S3 – S6. Both models give
468 similar results with differences up to 10%, which remain small with no systematic
469 bias. We thus present our results below with our preferred modeling considering not
470 well averaged inheritance.

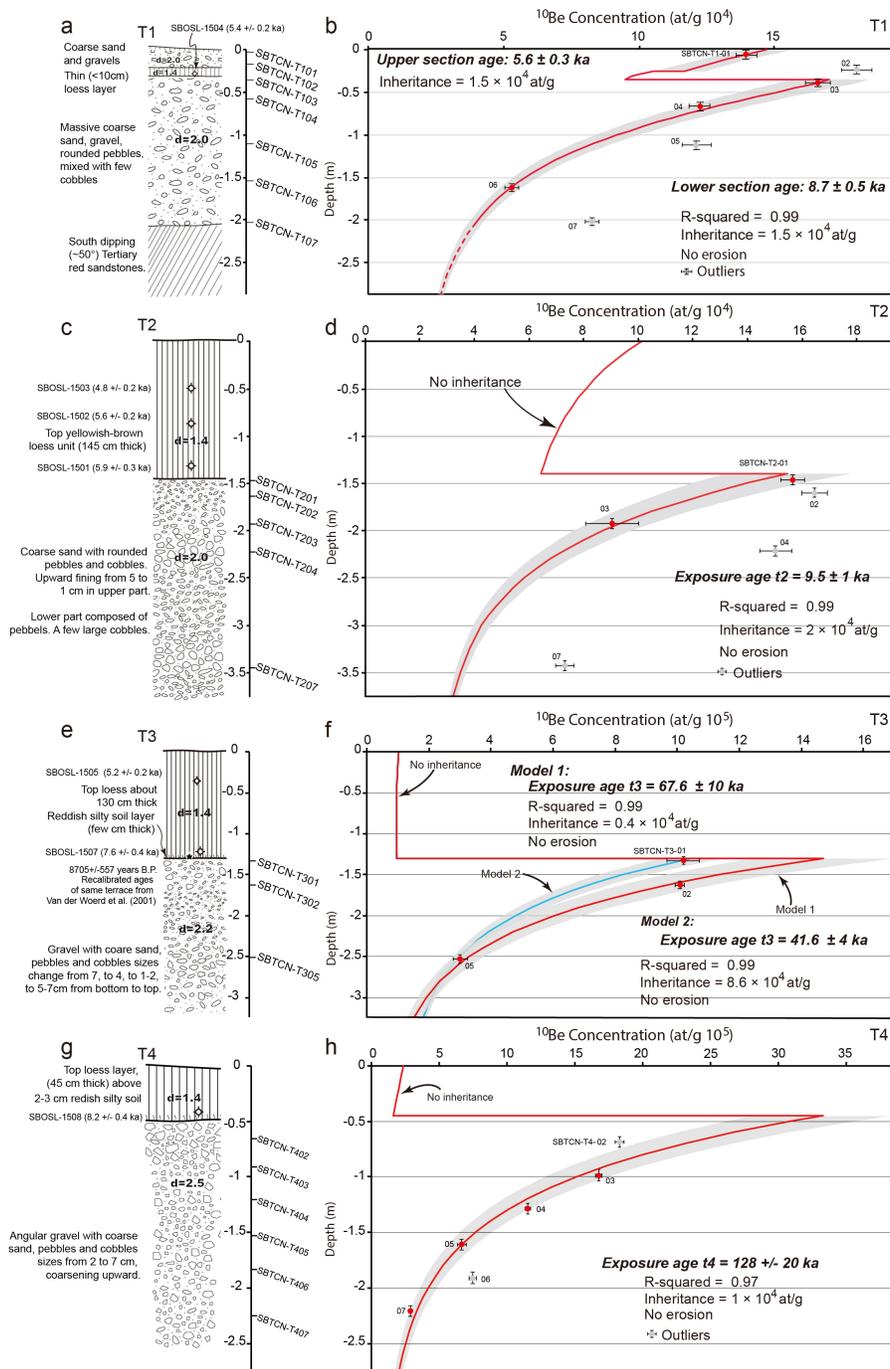
471



473 Figure 9. Field views of terraces and sampling sites at site 1 (locations of views and
474 pits in Figure 4). (a) Westward view across T1 and T3. Position of depth profile in top
475 2 m of a loose conglomerate in T1, 7 m above the river bed. Orange colored outcrops
476 below unconformable grey surface conglomerate are south-dipping Tertiary red beds.
477 (b) Close-up view of depth profile in T1. Sandy gravels and pebbles are collected
478 every 20 cm along the profile. (c) Pit dug in terrace T2. Conglomerate is capped by
479 145 cm-thick loess deposit. (d) Landscape view taken from upper terrace T4 eastward
480 across site 1, Subei county in background left, and site 2 to the right. Location of pit
481 dug in T3 indicated. (e) View of sampling pit in T3 with top 130 cm-thick loess,
482 separated from bottom conglomerate by a reddish silty soil layer. (f) 3 m-deep pit in
483 terrace T4. The top 45 cm-thick layer is made of loess covering a few cm-thick
484 reddish silty soil.
485

486 The depth profile in T1 includes 7 samples in the 2 m-thick deposit, which consist
487 of a 20 cm-thick gravel deposit on top of a 10 cm-thick loess layer above a
488 homogenous 1.7 m-thick conglomerate (Figure 10a). An OSL sample from the loess
489 layer yields a date of 5.4 ± 0.2 ka, which we interpret as a minimum age of the terrace
490 (Table 2). The best fit to the ^{10}Be concentration with depth explains four out of seven
491 samples ($R^2 = 0.99$) and yields a model age of 8.7 ± 0.5 ka. The modeling suggests
492 two depositional phases: a first phase of conglomerate accumulation since 8.7 ± 0.5
493 ka; after a hiatus, a second phase of 10 cm-thick loess accumulation since 5.4 ± 0.2 ka,
494 and the gravel layer on top (Figure 10b). In the following, we use the 8.7 ± 0.5 ka as
495 the age of T1 as it represents the major phase of terrace formation and because the top
496 loess and gravel layer may post-date the onset of folding and could be derived from
497 post-depositional surface reworking.

498 A ~3.6 m-deep pit in T2 reveals a loess layer of 145 cm above the conglomerate
499 of the terrace (Figure 10c). Three OSL samples at depths of 50, 90, and 135 cm yield
500 dates of 4.8 ± 0.2 , 5.6 ± 0.2 , and 5.9 ± 0.2 ka, respectively (Table 2). The samples are
501 stratigraphically ordered, which implies that the 145 cm-thick loess deposits
502 accumulated between 6.1 ka and 4.3 ka at a rate of ~72 cm/kyr. Because the ^{10}Be
503 concentrations of 5 conglomerate samples distributed between 1.5 and 3.5 m depths
504 are scattered, we fit only the 2 less concentrated samples at 1.5 and 2 m depths, thus
505 yielding an exposure age of 9.5 ± 1 ka (Figure 10d).



506

507 Figure 10. Depth profile description, sample type and position, and ^{10}Be age models
508 for T1, T2, T3, and T4 at site 1. (a), (c), (e), and (g) are simplified stratigraphic logs
509 with sample positions and average layer densities. (b), (d), (f), and (h) are ^{10}Be depth
510 profile models “d” stands for density in g/cm^3 . See also alternative models in Figures
511 S3 – S6.

512

513 The 3 m-deep depth profile in T3 reveals a ~130 cm-thick loess layer on the top
514 with a thin reddish silty soil layer at its base, indicating weathering and pedogenesis

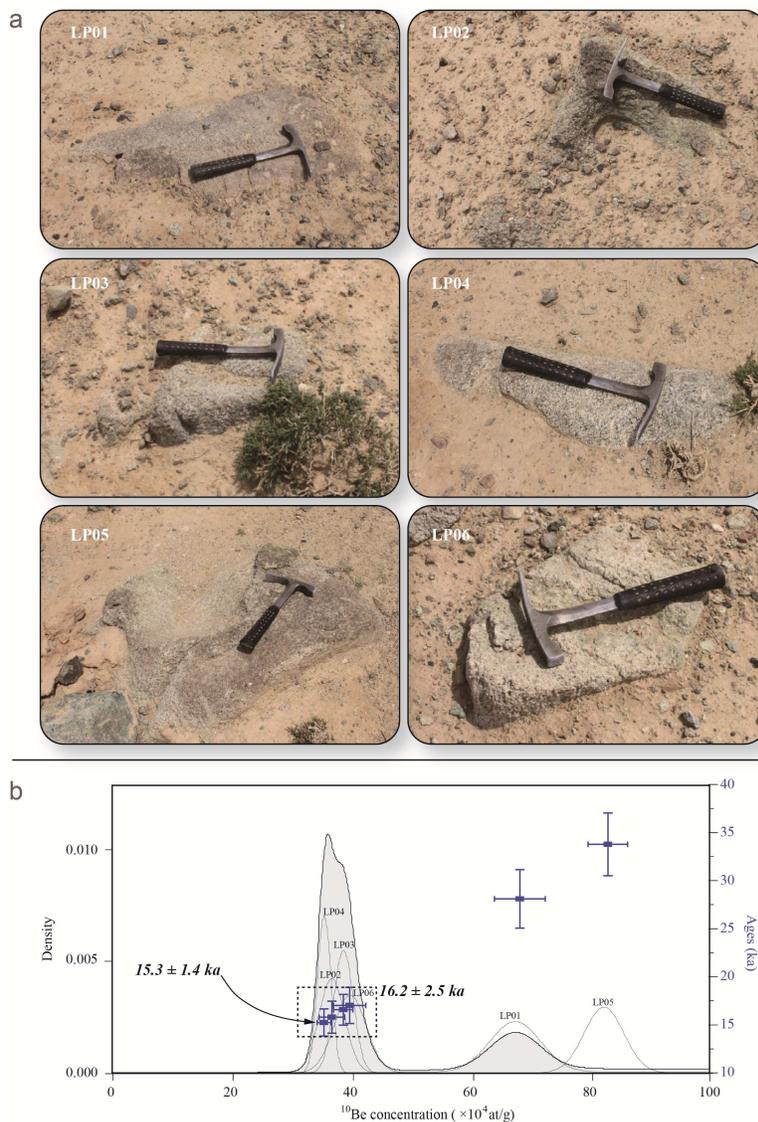
515 (Figure 10e). Two OSL samples at depths of ~40 and ~130 cm yield burial ages of 7.6
516 ± 0.4 ka and 5.2 ± 0.2 ka, respectively (Table 2), indicating loess deposition since
517 ~7.6 ka at a rate of 37 cm/kyr. Three charcoal samples collected ~150 m to the
518 southwest of the depth profile at the conglomerate-loess interface of T3 (see position
519 in Figure 4; van der Woerd et al., 2001) yield consistent maximum ^{14}C AMS ages of
520 8.2 – 9.2 kyr B.P. for the base of the loess (Table 4). Three conglomerate samples
521 from depths of 1.3 to 2.5 m are used to constrain the exposure age of T3. We suggest
522 two model scenarios that alternatively fit the two top-most samples (Figure 10f).
523 Model 1 treats sample SBTCN-T3-01 as a reworked surface sample, whereas Model 2
524 treats sample SBTCN-T3-02 as an outlier with larger inheritance. Model 1 yields an
525 exposure age of 67.6 ± 10 ka and Model 2 yields an exposure age of 41.6 ± 4 ka
526 (Figure 10f). In any case, the considerable time lag implied by the TCN modeling (30
527 to 50 kyr) between the conglomerate deposit and the overlying loess cover indicates
528 that the top terrace remained devoid of loess cover most of its time, consistent with
529 the observation of soil development at its top (Figures 9 and 10e). Given the
530 significant uncertainty, the two models provide the upper and lower bounds on the age
531 of T3 (41 – 68 ka or 51.5 ± 13.5 ka).

532 The 2.8 m-deep pit in T4 reveals a ~45 cm thick loess layer with a thin reddish
533 silty soil at its base above the terrace conglomerate (Figure 9). One OSL sample at the
534 bottom of the loess yields a burial age of 8.2 ± 0.4 ka (Figure 10g, Table 2). The
535 model yields an exposure age of 128 ± 20 ka by fitting four out of six conglomerate
536 samples (Figure 10h). Samples SBTCN-T4-02 and SBTCN-T4-06 are considered
537 outliers. Sample SBTCN-T4-02 is close to the top of the terrace deposit and may have
538 been bio- or cryo-perturbed. Including sample SBTCN-T4-02 results in a similar
539 age with larger uncertainties (Figure S6). Similar to terrace T3, the time lag between
540 the deposition of the conglomeratic strath and the overlying loess cover (~100 kyr) is
541 consistent with soil development at the top of the conglomerate and indicates no or
542 minimal surface erosion before loess mantled its surface.

543 At site 2, six embedded granitic boulders from surface Tf2 were targeted for TCN
544 dating (Figure 11). Four (LP02, 03, 04, 06) samples are clustered at approximately 16
545 ka, with an average age of 16.2 ± 2.5 ka (youngest age of 15.3 ± 1.4 ka) (Table 2;
546 Figure 11). Two samples yielded older dates than the rest of the samples (~25 – 35
547 ka), probably owing to longer exposure in the catchment upstream resulting in larger
548 inheritance and are thus considered outliers. Since the ages of the four youngest

549 samples are well-clustered, erosion and inheritance may be considered negligible (e.g.,
550 van der Woerd et al., 1998; Ryerson et al., 2006; Matrau et al., 2019). The position of
551 Tf2 relative to the other terrace implies that the date of 16.2 ± 2.5 ka is a maximum
552 emplacement age of the fan surfaces at site 2, indicating that these terraces and fans
553 formed at the end of the last glacial maximum with the return of a more humid
554 climate (e.g., van der Woerd et al., 2002; Hetzel, 2013).

555 In the trench, one radiocarbon sample collected at the bottom of the upper loess
556 unit U1 in the west wall was dated at 8040 – 8190 years B.P.. This age is overall
557 concordant with the base of the loess units at site 1 and even farther west (this study;
558 [Tables 2 and 4](#); see also van der Woerd et al., 2001). When we consider the bottom
559 burial age of the loess covering the terrace conglomerate at ~ 8 ka, we can bracket the
560 age of Tf2 between 8 and 16.2 ka (12 ± 4 ka).



561

562 Figure 11. ^{10}Be dating of surface samples at site 2. (a) Samples were collected
 563 from the surface of imbedded granitic boulders. Hammer for scale. See Figure 8 and
 564 Table 3 for location details. (b) ^{10}Be cosmogenic nuclide concentration and age
 565 distributions of TCN samples at site 2 (Table 3). Thin curves are single sample
 566 density distribution, and gray shaded area is cumulative density distribution. Four
 567 youngest ages cluster with overlapping error bars and constrain alluvial fan (Tf2)
 568 abandonment age.
 569

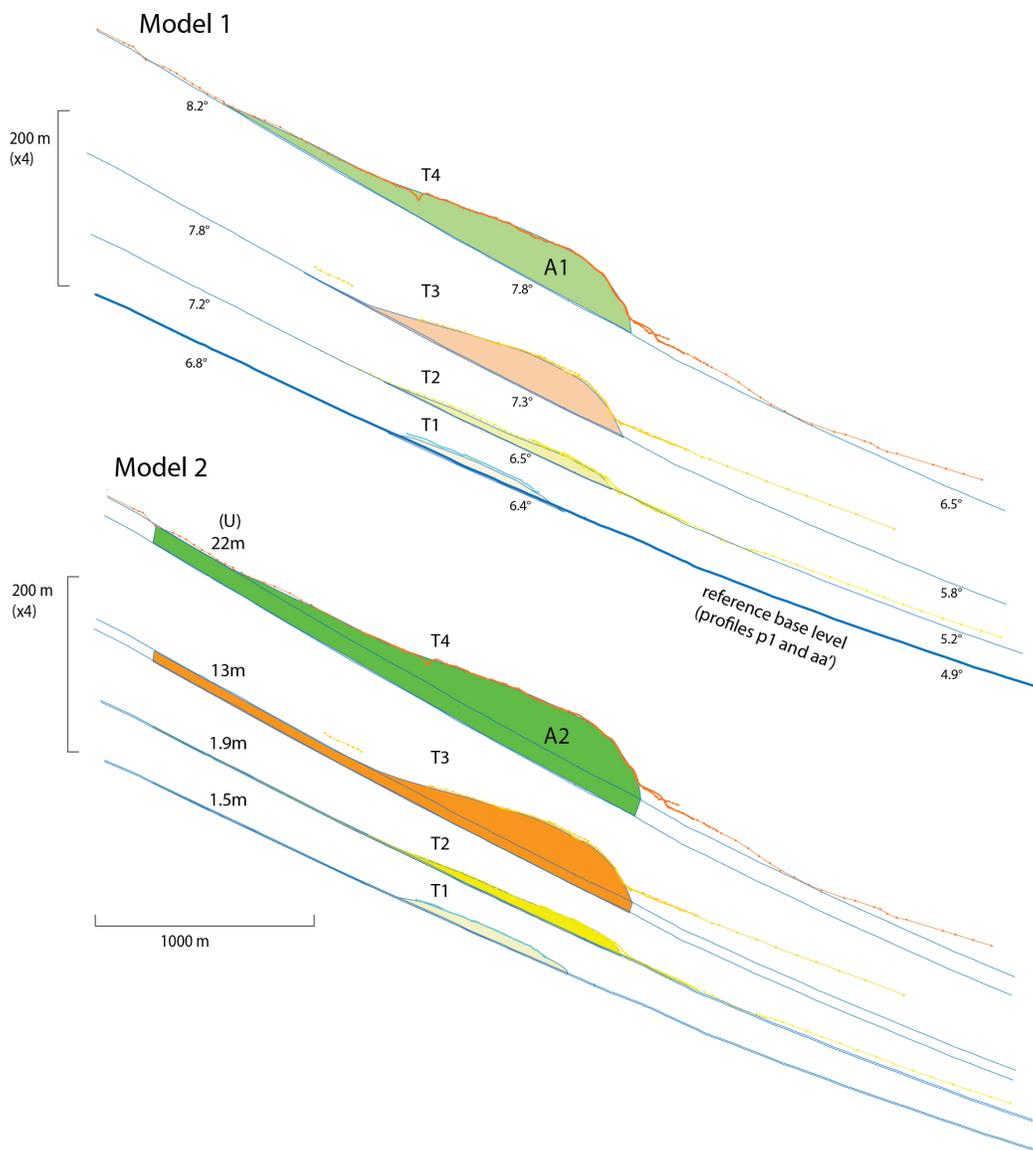
570 5. Uplift and shortening rates estimates

571 Site 1 along catchment r5 has been targeted because it is one of the above average
 572 sized catchments along the western DHNS range with moderate incision, which has
 573 preserved a set of four major terrace levels (Figure 4; e.g., van der Woerd et al., 2001).
 574 The tops of terraces T1 to T4 stand 7 to 60 m at their highest point above the present

575 stream bed (Figure 6). The highest terrace T4, which can be mapped up to the range
576 front, has a slightly steeper slope upstream ($7 - 8^\circ$) than the present stream bed ($6 -$
577 7°), indicating progressing slope lowering with time. In addition, sedimentation at the
578 front of the fold and burial of the older fan deposits also need to be considered when
579 reconstructing older base levels (van der Woerd et al., 2001). Indeed, the present
580 bajada is dominated by deposits related to T2 (see Figure 4), i.e., Holocene fan
581 deposits, and by the present-day active stream outlets and fans that develop farther
582 downslope (Figures 2 and 4; see also van der Woerd et al., 2001). Because the stream
583 gradient changes across the length of the terraces (2° between F0 and F1, i.e.,
584 approximately 3 km) and because the past base level geometry is not available, we
585 take the present-day stream bed river profile and its gradient variations as a template
586 for reconstructing past base levels. Given the steeper slope of upstream terrace T4, we
587 thus rotate the present-day stream bed profile to fit the steeper gradient of terrace T4.
588 For terrace T3, we choose an intermediate gradient between that of T4 and the
589 present-day gradient. For T2 and T1, the base level is that of the present-day stream r5
590 (Figure 12).

591 The frontal ramp is considered to reach the surface with a dip close to the bedding
592 of the frontal red beds, i.e., approximately $40 - 50^\circ$ (van der Woerd et al., 2001). The
593 wavelength of the antiformal folding of the terraces averages approximately 1000 m,
594 thus constraining the depth where the ramp and the décollement meet, i.e., at 1000 m
595 depth (D_0 ; Figures 3 and 12, Table 6). For Model 1, this depth then remains constant
596 until it joins with thrust F0. For Model 2, the décollement dips with an angle α and
597 thus joins F0 at a deeper depth (approximately 1900 m; Figure 12, Table 6).

598 Shortening results for models 1 and 2 are shown in Table 6. The areas considered
599 are measured between the top present-day folded and uplifted surface of the terraces
600 and the reconstructed base level (Figure 12), neglecting the impact of loess cover and
601 thickness of the conglomerates as their thickness remains small compared to the
602 amounts of incision. Overall, the thickness of the conglomeratic deposit above the
603 strath is approximately 2 – 4 m for T1 and T2 and reaches 5 – 10 m for T3 and T4.



604
 605 Figure 12. Reconstructed uplifted areas A1 and A2 for the 4 terraces at Site 1
 606 considering end-member models with (Model 2) and without (Model 1) dip of the
 607 décollement. Blue lines are reference base levels above which uplifted area is outlined
 608 by terrace surface profiles (Figures 4 and 5). See text and [Table 6](#) for details.
 609

610 In Model 1 shortening is absorbed solely by the warped frontal part of the
 611 terraces, and the upstream parts of the terraces are considered to have little or no uplift.
 612 The incision of the river is thus driven mainly by progressive base level adjustment
 613 (see van der Woerd et al., 2001). For this model, the shortening rate varies from 0.8 –
 614 1.0 mm/yr for T1 and T2 to 0.5 mm/yr for T3 and T4 ([Table 6](#)).

615 Model 2 considers a dipping décollement before it ramps up as fault F1. In this
 616 case, shortening is absorbed by both the warped frontal part of the terraces and their
 617 uplifted upstream part. The incision of the river is now a combination of both

618 progressive base level adjustment and overall uplift and folding of the terraces
619 between F0 and F1. With this model, adjusting the dip of the décollement allows to
620 minimize shortening rate differences throughout the set of terraces. An almost
621 constant shortening rate of 0.5 ± 0.1 mm/yr for all terraces can be calculated with a
622 dip of 23.5° , the corresponding uplift rate is 1.1 ± 0.3 mm/yr (Table 6).

623 At site 2, the 4.9 ± 0.5 m total slip on all branches of fault F2, as revealed from
624 trenching may correspond to slip on a 47° dipping ramp to account for the 3.6 ± 0.2 m
625 cumulative uplift and approximately 3.4 ± 0.2 m of shortening of Tf2. Dividing the
626 slip values by the age of Tf2 (12 ± 4 ka) yields uplift and shortening rates of 0.5 ± 0.1
627 mm/yr on fault F2.

628 Considering the less well constrained shortening rate along thrust F2, and
629 accounting for some shortening accommodated along thrust F0, total uplift and
630 shortening rates along the western Danghe Nan Shan range front, south of Subei, for
631 the last 130 ka could reach 1.4 ± 0.4 mm/yr and 0.8 ± 0.2 mm/yr, respectively.

632

633 **6. Discussion**

634 6.1. Shortening rate bias due to unsteady base level

635 Using terrace heights in the hanging wall of active thrusts to estimate shortening
636 rate requires knowledge of a reference base level (e.g., Merritts et al., 1994; Guilbaud
637 et al., 2017; Hetzel et al., 2004; Lavé & Avouac, 2000; van der Woerd et al., 2001;
638 Daeron et al., 2007; Hu et al., 2021). Although changes with time of the base level are
639 almost always a dynamic adjustment to the sediment supply or the transport capacities
640 of a stream, they can be triggered by various causes, such as climate change, stream
641 capture, subsidence, uplift, and earthquake-triggered changes in sediment supply
642 (Tapponnier et al., 1990; Merritts et al., 1994; van der Woerd et al., 2002; Bull, 2008;
643 Poisson & Avouac, 2004; Rizza et al., 2019). Long-term tectonic uplift is a cause of
644 changes in base level when erosion and incision are not able to compensate for uplift.
645 This is the situation along the western DHNS range front, where small and steep
646 catchments are uplifted (van der Woerd et al., 2001). As a consequence, aggradation
647 occurs in the bajada south of Subei, as shown by the youth of the fan deposits (fan Tf2
648 of Holocene age) burying older deposits.

649 We have no subsurface data of the deposits in the footwall of fault F1. However,
650 according to our mapping (Figure 3), fan Tf2 deposits in the footwall correspond to
651 terrace T2 in the hanging wall, which indicates that the pre-Holocene corresponding

652 footwall terrace deposits are buried. This is considered in Model 2 (Figure 12).
653 Therefore, the heights of terraces T3 and T4 above their respective base levels are
654 possibly greater than 13 and 22 m, as suggested in Model 2 (Figure 12; Table 6).
655 Consequently, the estimated shortening is minimized.

656 6.2. Shortening model implementation

657 Contrary to large mountains belt incised by large rivers (>50 km long), like in the
658 North Qilian Shan, the Western Kunlun Shan, the Tian Shan, the catchment lengths
659 west of Subei barely exceed about 5 km. As a consequence, the intermittent streams
660 that incise the western DHNS are far from being stabilized graded rivers over long
661 time periods. Shortening models as proposed, for instance, by Hu et al. (2015),
662 Thompson et al. (2002), are well adapted for the development of folds along stable
663 rivers with almost no large base level changes and negligible slope changes. This is
664 not the case west of Subei, where, as explained above, we still fail to reconstruct a
665 reliable base level throughout the deposition time span (130 ka) of all the terraces. We
666 thus choose to adapt our simplified structural model to this situation by considering
667 two end-members, one without dipping décollement (Model 1) and another with a
668 dipping décollement (Model 2) (Figure 3). Both situations are equally possible as
669 illustrated by the variety of well documented fold and thrust belts across central Asia
670 (e.g., Tapponnier et al., 1990; Avouac and Tapponnier, 1993; Meyer et al., 1998; van
671 der Woerd, 1998; van der Woerd et al., 2001; Poisson and Avouac, 2004; Hubert-
672 Ferrari et al., 2007; Guilbaud et al., 2017).

673 Making the right assumptions for assigning parameters to estimate shortening rate
674 is challenging (Figure 3; see also, for instance, Suppe and Medwedeff, 1990; Lave
675 and Avouac, 2000; Thompson et al., 2002; Hu et al., 2015). Here, we use the
676 topographic profiles (Figure 5) and ages (Figure 10) of 4 incrementally folded terraces
677 together with both a changing base level with time (slope decrease) and a changing
678 décollement dip (from flat to dipping) (Figures 3 and 12). The 2 proposed end-
679 member models capture a large range of the variability in shortening rate estimates
680 (0.4 to 1 mm/yr, Table 6) although, as explained above, these rates might still be
681 minimum rates (Figures 2, 13, 14).

682 6.3. Link with long-term cumulative shortening and exhumation

683 Accelerated range growth between 13.7 Ma and 9 Ma inferred from
684 magnetostratigraphy (Sun et al., 2005) is correlated with the deposition of coarse
685 conglomerate deposits along Tiejangou (see location in Figure 2). Provenance change

686 and lag-time variation of detrital apatite fission-track data in the same section
687 corroborate a tectonic pulse at 14 Ma (Lin et al., 2015). Consistently, apatite fission-
688 track and apatite (U-Th)/He data from three bedrock transects in the western DHNS
689 (see location in [Figure 2](#)) suggest a period of rapid exhumation of approximately 0.1
690 to 0.3 mm/yr from 16 to 9 Ma (Yu et al., 2019).

691 Balanced geological sections indicate cumulative shortening of 10 – 20 km by
692 thrusts F0 and F1, based on structural constraints along several river cuts with an
693 onset of shortening likely at 9 – 14 Ma based on the age of the uppermost coarse
694 conglomerate (Gilder et al., 2001; van der Woerd et al., 2001). The average
695 shortening rate is thus estimated at 1.5 ± 0.7 mm/yr since the late Miocene. This
696 estimate may indicate a near-constant shortening rate since 14 Ma when compared to
697 our late Pleistocene rate. In addition, it indicates that F1 and F2 accommodate most of
698 the shortening responsible for the present-day uplift of the western DHNS.
699 Conversely, it may suggest that F0 or the southern DHNS thrust do not accommodate
700 much shortening, although more work is probably needed on these mostly bedrock
701 faults to assess their slip history.

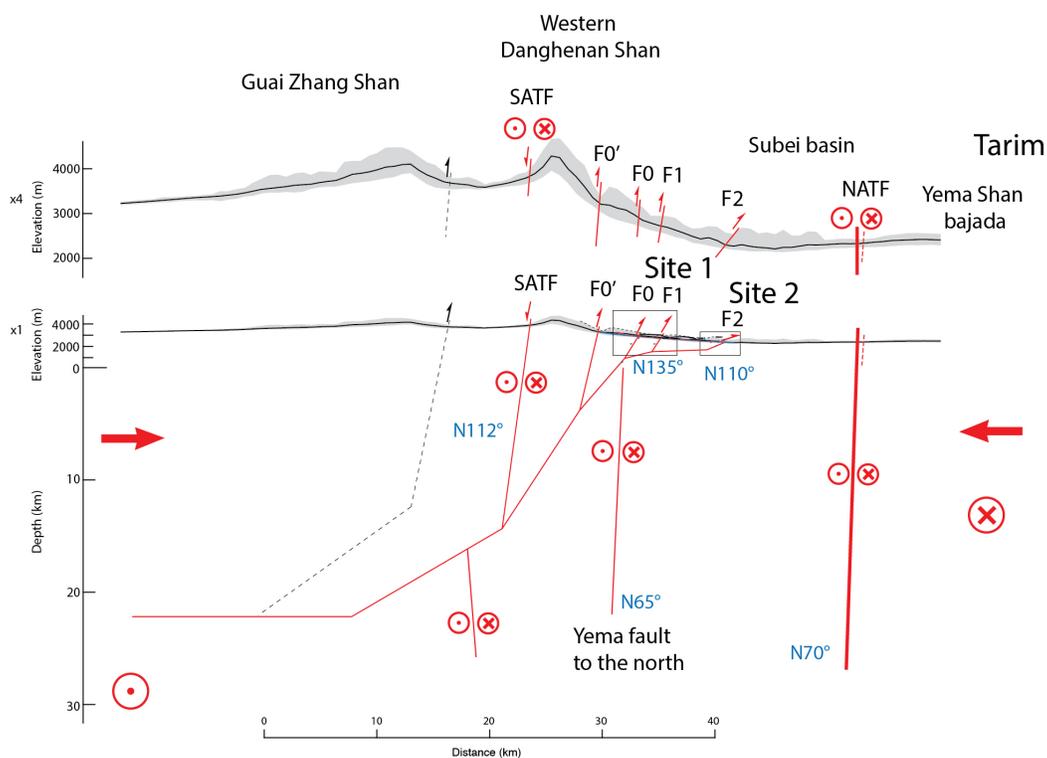
702 6.4. Slip accommodation at the Subei junction: geometric and kinematic 3D
703 complexities

704 6.4.1 *The fault geometry in 3D*

705 The Subei junction is particularly complex and involves numerous active faults,
706 including strike-slip faults, thrusts, and normal faults (e.g., Meyer et al., 1998). The
707 N70°E striking north strand of the ATF (NATF) connects the Aksay section with the
708 Shibao Cheng section to the northeast ([Figure 1](#)). West of Subei, the N135°E-striking
709 and south-dipping thrusts of the western DHNS veer to the west and connect with the
710 NATF. Although the NATF can be traced continuously, thrusting also occurs north of
711 it but west of the junction (e.g., van der Woerd et al., 2001). East of the junction and
712 north of NATF, normal faults run parallel to the ATF for 20 km across the Subei
713 basin and Dang He ([Figures 2 and S2](#)). Slip transfer from NATF to the DHNS, and
714 rigid block kinematics explain the extensional component northeast of the junction
715 (e.g., Meyer et al., 1998). In the long term, these kinematic interactions also explain
716 the long-lived Subei trough along the NATF.

717 The geological and topographical continuity of the arcuate shape of the western
718 DHNS range, underlined by the continuous trace of the south strand of the ATF
719 (SATF), testifies to the time-progressive accumulation of deformation both by strike-

720 slip faulting along the ATF and shortening across the DHNS range. This long-term
 721 coupling of strike-slip faulting and thrusting accommodated the strain at the margin of
 722 the Tibet plateau contributing to continental plateau growth and widening (e.g., Meyer
 723 et al., 1998; Tapponnier et al., 2001; van der Woerd, 1998; van der Woerd et al.,
 724 2001). The two sub-parallel, arcuate ranges of Guai Zhang Shan (GZS) and DHNS,
 725 with their contrasted morphology, further depict this long-term coupling. GZS
 726 summits are only slightly below the peaks of DHNS, although it is not bounded by
 727 active thrust faults. On the contrary, valleys and mountain slopes merge gently with
 728 the surrounding sedimentary basin fills, indicating no tectonic activity but passive
 729 erosion. We thus infer that GZS grew in the early Miocene and stopped growing as
 730 strain stepped northeastward along the ATF (Figure 12). Describing the full 3D
 731 geometry of the Subei junction is beyond the scope of this study, and we only stress
 732 part of this complexity to address the difficulty of balancing the kinematics across the
 733 junction.



projection N15°

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Figure 13. Simplified tectonic section across western Danghe Nan Shan with possible fault geometry at depth (see location of faults and section in Figures 1 and 2). Thrust fault north of Guai Zhang Shan (black dashed line) is inferred. Blue labels present fault strikes.

741 Although the structural link in map view is clear between the thrusts F0', F0 and
742 F1 with the N70°E striking ATF, how thrusts faults that root onto shallow dipping
743 ramps and décollement at the crustal level (e.g., Meyer et al., 1998) are linked with
744 subvertical lithospheric scale strike-slip fault (e.g., Wittlinger et al., 2004) is not
745 straightforward. Our mapping (Figures 1 and 2) and section (Figure 13) suggest a
746 transition from strike-slip to thrusting as faults progressively deviate from the N70°E
747 direction of the main ATF and become perpendicular to the main stress direction
748 (N45°E) as depicted in thrust focal mechanisms (Figure 1) or the average geodetic
749 displacement field direction (Wang and Shen, 2020). Southeast of Subei, other strike-
750 slip faults emerge (YDS or YNS) at the eastern fold tip implying kinematic links and
751 slip transfer from thrusting to strike-slip faulting (Figures 2, 13 and 14).

752

753 *6.4.2 Slip accommodation by the western Danghe Nan Shan*

754 If the amount of shortening along the DHNS results entirely from slip transfer
755 from the north strand of ATF, then a similar amount of left-slip rate is lost for the
756 north strand of ATF eastward from the Subei junction. Our estimated shortening rate
757 on fault F1 is approximately 0.5 mm/yr (Table 6), and the total, including fault F2 and
758 possible additional slip on F0, could amount to 1 mm/yr. There is no doubt that the
759 rate of slip of the north strand of the ATF decreases toward the east (e.g., Xu et al.,
760 2005; Wang and Shen, 2020; Xu. et al., 2021; Figure S2), however, where and how
761 the slip is distributed remains difficult to constrain.

762 If all the left-slip displacement along the ATF is absorbed in the Qilian Shan
763 region in shortening or left-lateral faulting (Tapponnier and Molnar, 1977; Peltzer et
764 al., 1989; Burchfiel et al., 1989; Meyer et al., 1998; Tapponnier et al. 2001) and the
765 knowledge of slip rate on each fault is known, one should be able to build an
766 equilibrated kinematic budget between slip on the ATF and the other faults to the
767 southeast (Figure 14). However, slip rates on faults are variably documented in the
768 Qilian Shan region (Figure S2, Table S1).

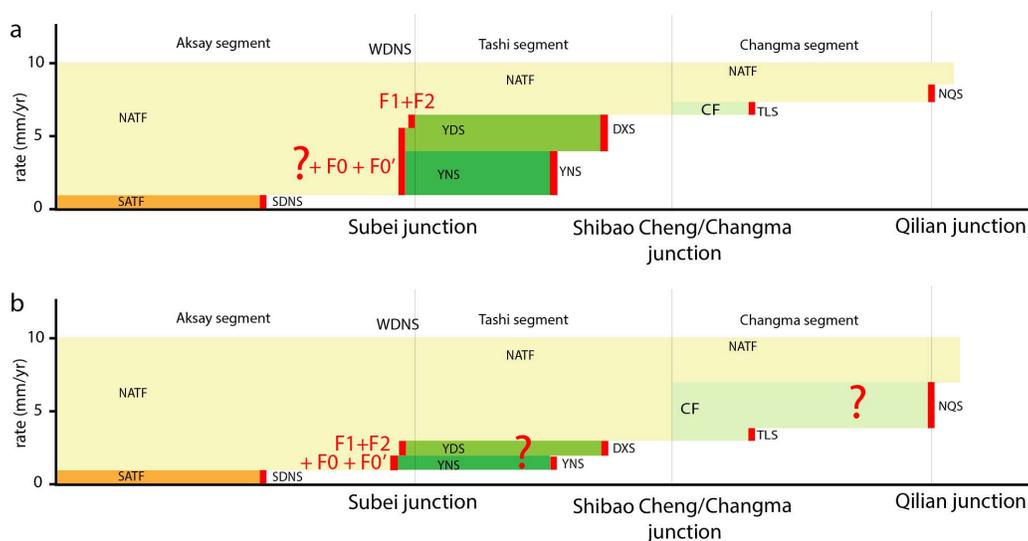
769 Figure 14 suggests an overall slip-rate decrease along ATF from about 10 mm/yr
770 west of Aksay to a few mm/yr east of the Qilian Shan junction at 97°E. At the Subei
771 junction, our shortening rates and published rates on adjacent faults may point to
772 kinematic incompatibilities. On one hand (Figure 14a), if indeed YDS and YNS faults
773 connect southeast of thrusts F1 and F2, our maximum total shortening of 1 mm/yr is
774 low compared to the published rates on YDS (2 – 3 mm/yr; Luo et al., 2015) and YNS

775 (3 mm/yr; Zheng et al., 2013), as well as shortening along DXS (1-3 mm/yr; Meyer et
776 al., 1998; Luo et al., 2015) (Figure 14a). The rates on YDS and YNS may suggest
777 additional connection to the ATF in addition to thrust F1 and F2, which might be
778 found in thrust faults within the western DHNS, mapped as faults F0' and F0 (Figures
779 2 and 13). Thrusts faults F0' and F0 cut through western DHNS bedrock and may
780 have been overlooked. On the other hand (Figure 14b), rates along YDS and YNS
781 may be overestimated and no additional slip transfer from faults F1 and F2 is needed.
782 In this case, the ATF slip-rate east of Subei remains larger and more slip is transferred
783 in the junctions eastward, at Shibao Cheng/Changma and Qilian (Figures 1 and 14).

784 Complex fault systems are subject to slip variations as faults interact in time and
785 space during their seismic cycle (e.g., Dolan et al., 2016; Matrau et al., 2019; Bennet
786 et al., 2004). Building a kinematic budget as proposed in Figure 14 implies that rates
787 are known over similar temporal periods or that their rates remain constant over long-
788 time scale. Very few rates are determined for time scales larger than the last 10 ka
789 (Table S1). Knowing that the return time of large magnitude 7 – 8 earthquake in the
790 region is on the order of a few 1000 years (Figure 1; e.g., Luo et al., 2019), makes
791 these rates highly dependent on their seismic cycle. Discussing variations in slip-rate
792 in space and time, even along the ATF that is the most documented fault in the area,
793 remains challenging.

794 Overall, the rate we found along the western DHNS near Subei is comparable to
795 the rate estimated eastward along the same range in Yanchi Wan (Figure 1; Xu et al.,
796 2021). It is also comparable to other rates along the northern Qilian Shan using
797 similar approaches (e.g., Tapponnier et al., 1990; Hetzel et al., 2004; Hu et al., 2015;
798 Yang et al., 2018; Wang et al., 2020), which indicates that the NE-SW directed
799 shortening in the DHNS, together with other thrust faults, contributes to overall
800 crustal shortening in the Qilian Shan region.

801



802

803 Figure 14. Schematic kinematic budget along ATF (see Figure S2, and Table S1).
 804 Width of color band is rate. Red bars are shortening rates connected to strike-slip
 805 faults. (a) Published rates on YDS and YNS are larger than our determined rates on
 806 F1 and F2. Possibly, rates of F0 and F0' are underestimated. (b) Rates on YDS and
 807 YNS are overestimated as well as the shortening rates along DXS and YNS. Strike-
 808 slip rate is larger on ATF east of Subei, and shortening rates larger on TLS and QS.
 809 NATF: north strand of Altyn Tagh fault, SATF: south strand of Altyn Tagh fault,
 810 SDNS: south Danghe Nan Shan thrust, WDNS: western Danghe Nan Shan thrust, CF:
 811 Changma fault, TLS: Tuolai Shan thrust, NQS: Northern Qilian Shan thrust, DXS:
 812 Daxue Shan thrust, YNS: Yema Nan Shan thrust.

813

814 6.5. Seismic hazards

815 Evaluating seismic hazards in the Subei area becomes critical due to population
 816 growth and the new traffic line across the Danghe Nan Shan thrust and the Altyn
 817 Tagh fault (Figure 2). There is no historical record of large earthquakes in the Subei
 818 basin related to faulting along the northern DHNS thrust. However, multiple 0.6 to 2.7
 819 m-high fault scarps across the fans in front of the frontal folded hills (e.g., site 3 of
 820 van der Woerd et al., 2001; site 2 of this study; Figure 2) suggest several surface
 821 rupturing events in the Holocene. A paleoseismic trench across one of these scarps on
 822 a lower terrace by Wu et al. (2017) shows evidence of slip on a 20 – 25° dipping fault
 823 plane and ~1.7 m vertical offset after ~4.2 ka B.P. The ~1.5 m high scarp across the
 824 same terrace has been suggested to result from one seismic event, while the 2.6 – 3.3
 825 m cumulative vertical offset of this terrace, approximately twice the height of the
 826 scarp, implies a record of two earthquakes (van der Woerd et al., 2001). Although no
 827 evidence of a penultimate event was described in the trench (Wu et al., 2017), we
 828 cannot rule out the possibility of the occurrence of one older event in the frontal ramp

829 and folding in the hanging wall.

830 At site 2, similar 1.5 – 3.6 m scarp heights are observed across various terraces
831 (Figures 5, 6). The young terraces (pf2 and pf3) show smaller offsets (1.5 m) than the
832 older terraces (pf4-pf5 with offsets up to 3.5 – 3.6 m), suggesting the occurrence of
833 two events, with the latest event occurring after emplacement of the youngest terrace.
834 The ages of these two events are loosely constrained by limiting ages at site 2. In the
835 trench, a radiocarbon sample under the younger event horizon provides an age of 8.1
836 ka for the loess unit (U1). The exposure age of the terrace defines a maximum age of
837 16.2 for the gravels unit (U2). We then can bracket the occurrence time of the older
838 event between 16.2 and 8.1 ka and the younger event after 8.1 ka. According to the
839 ages of these events, it is possible that the younger event at site 2 correlates with the
840 recent event at site 3.

841 Earthquake magnitude can be estimated by the empirical relationship of rupture
842 parameters (e.g., Wells & Coppersmith, 1994). There is no strong evidence indicating
843 that sites 2 and 3 (15 km apart) along two different structures (F1 and F2) have
844 ruptured together during the same seismic events. The sizes of the minimum slip
845 events at both sites, however, are compatible with concomitant 20 km-long ruptures.
846 It is thus tempting to suggest a minimum magnitude of 6.5 – 7 (1 – 2 m of slip over 20
847 km).

848 We suggest that both thrusts F1 and F2 merge into one ramp and décollement at
849 depth (Figure 13). Mapping of the surface trace of the thrusts and the deformed
850 terraces and fans in their hanging wall constrain in part the geometry of the ramp and
851 décollement at depth (e.g., Meyer et al., 1998; Stockmeyer et al., 2014). The
852 combination of co-seismic thrust scarp measurements and the cumulative deformation
853 of the thrusts makes it possible to suggest the occurrence of events with multi-fault
854 ruptures. Taking the length of the thrust system (20 km), the total width of the ramps
855 and décollement above a depth of 15 km (20 km), and a total slip at depth of 4 m
856 distributed on F1 and F2 would result in a seismic moment of 7.9×10^{19} N.m with the
857 shear modulus of 3.3×10^{10} N/m². Such a moment is compatible with a moment
858 magnitude of $M_w=7.2$ (e.g., Kanamori, 1977).

859 Complex ruptures on multiple faults are not uncommon, as described above. In
860 Subei, thrust faults F1 and F2 are connected with strike-slip faults to the west and to
861 the east (see Figures 2 and S2). These connected faults could rupture together
862 similarly to the 2016 $M_w7.8$ Kaikōura complex earthquake rupture (e.g., Klinger et

863 al., 2018). Previous paleoseismic investigations show that temporally consistent events
864 ruptured on these faults (Xu et al., 2007; He et al., 2018; Luo et al., 2019). For
865 instance, both the Subei-Old Aksay section of the north strand of Altyn Tagh fault, the
866 Yema-Daxue Shan fault ruptured in 6.3 – 6.8 ka (Xu et al., 2007; He et al., 2018), and
867 the east section (east of Subei) of the north strand of Altyn Tagh fault also ruptured in
868 6.4 – 7.2 ka (Luo et al., 2019). The section from Old Aksay to Shibao Cheng of the
869 north strand of Altyn Tagh fault ruptured in 2.6 – 2.8 ka, which is consistent with the
870 most recent event of the Danghe Nan Shan thrust. While the date uncertainties may
871 only indicate close in time ruptures, it is not possible to rule out simultaneous ruptures
872 on several faults. Thus, if combined ruptures involving both the thrust system and
873 strike-slip faults are considered, then our magnitude estimation above is a minimum.
874 One such earthquake could potentially damage local infrastructure. More structural
875 and paleoseismic investigations involving all the active faults, are needed to assess the
876 seismic hazard of the complex fault system in and around the Subei basin.

877 **7. Conclusion**

878 Based on field investigations and high-accuracy topographic surveys of multiple
879 folded terraces along the western DHNS, we constrain the cumulative and recent
880 activity of two thrusts. Considering 2D mass balance across the fold and thrust belt,
881 with either a flat décollement or a dipping décollement allows us to constrain the
882 amount of accumulated shortening. Integrating the terrace ages constrained by
883 cosmogenic nuclides, OSL, and ^{14}C dating methods, we estimate a minimum
884 shortening rate of 0.8 ± 0.2 mm/yr for the northern thrusts of the western DHNS for
885 the last 130 ka. This rate is also compatible with the long-term shortening estimate,
886 which implies that the shortening rate across the DHNS has remained constant since
887 the upper Miocene. Our results indicate that the NE-SW-directed shortening along the
888 western DHNS thrusts contributes to slip-rate decrease along the main ATF and slip-
889 transfer to other thrust and strike-slip faults to the east. Evidence of surface seismic
890 ruptures and large earthquakes is clear from scarps and trenches along the western
891 DHNS. We suggest the possible occurrence of Mw7+ earthquake southwest of Subei.
892 Considering the structural links between the thrust system of the western DHNS and
893 the strike-slip faults of ATF to the northwest or Yema-Daxue Shan to the southeast,
894 earthquake ruptures may involve both strike-slip and thrust faults allowing for larger
895 magnitude events.

896

897 **Data Availability Statement**

898 The historical earthquakes are from the China Earthquake Data Center (CEDC,
899 <http://data.earthquake.cn/index.html>). The raw datasets of DEM derived from satellite
900 images (<https://doi.org/10.5281/zenodo.7045176>) and LiDAR point cloud
901 (<https://doi.org/10.5281/zenodo.7045169>) are shared in zenodo.org, and as well as the
902 terrace survey data of site 1 (<https://doi.org/10.5281/zenodo.7070457>). The focal
903 mechanisms come from the Global Centroid Moment Tensor project (CMT;
904 <http://www.globalcmt.org>).

905

906 **Acknowledgments**

907 This work is jointly funded by the National Key Research and Development Program
908 of China (2017YFC1500100), the Second Tibetan Plateau Scientific Expedition and
909 Research Program (STEP) (2019QZKK0901), and the National Natural Science
910 Foundation of China (41802228, U1839203), the Basic Research Project of Lanzhou
911 Institute of Seismology (2018IESLZ01). YS has been partly supported by a grant
912 from K.C. Wong Foundation and CNRS. We thank ASTER Team (D. Bourlès, G.
913 Aumaître, Vincent Godard, Régis Braucher and K. Keddadouche) at ASTER-
914 CEREGE for ¹⁰Be AMS measurements. ASTER AMS national facility (CEREGE,
915 Aix-en-Provence, France) is supported by INSU/CNRS, ANR through “Projets
916 thématiques d’excellence” program for “Equipements d’excellence” ASTER-
917 CEREGE action, and IRD.

918

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1316 Table 1. Terrace heights and vertical offset estimates

Terrace	Profile	Height above river bed upstream (m)	Maximum height above river bed (m)	Vertical offset relative to downstream fan (m)
Site 1				Relative to Tf2
T1	b-b'		7.5±0.5	-
T2	p3/c-c'	18±1	23±1	15±1
T3	p4/d-d'	27±2	40±2	31±2
T4	P2/e-e'	50±3	60±3	50±3
Site 2				
T1	pf2			1.5±0.2
T1'	pf3			1.6±0.2
T2'	pf4			3.5±0.5
Tf2	pf1			2.6±0.3
	pf5			3.6±0.5

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1319 Table 2. Optically stimulated luminescence (OSL) analytical data

Sample No	Lab. no ^a	Dept h(m)	Water content ^b (%)	U(ppm)	Th(ppm)	K(%)	Cosmic dose rate(Gy/ka)	grain size (μm)	Dose rate ^c (Gy/ka)	Equivalent dose ^d (Gy)	Age ^e (ka)
T1											
SB-OSL-1504	15-55	0.3	2.09	2.48±0.1	10.6±0.3	1.73±0.06	0.31±0.02	90-125	3.31±0.13	17.7±0.4	5.4±0.2
T2											
SB-OSL-1501	15-52	1.35	3.51	3.08±0.11	10.4±0.29	1.71±0.05	0.26±0.02	90-125	3.32±0.12	19.6±0.5	5.9±0.3
SB-OSL-1502	15-53	0.9	4.67	2.68±0.1	11±0.31	1.82±0.06	0.28±0.02	90-125	3.35±0.12	18.7±0.4	5.6±0.2
SB-OSL-1503	15-54	0.5	0.64	2.56±0.1	9.85±0.29	1.62±0.05	0.3±0.02	90-125	3.21±0.12	15.3±0.3	4.8±0.2
T3											
SB-OSL-1505	15-56	0.42	1.25	2.33±0.09	10.4±0.29	1.7±0.05	0.3±0.02	90-125	3.26±0.12	17±0.4	5.2±0.2
SB-OSL-1507	15-58	1.3	6.34	2.26±0.09	8.31±0.25	1.7±0.05	0.26±0.02	90-125	2.89±0.11	21.8±0.5	7.6±0.3
T4											
SB-OSL-1508	15-59	0.45	10.02	2.56±0.1	10.9±0.31	1.79±0.06	0.3±0.02	90-125	3.14±0.12	25.8±0.5	8.2±0.4

1320 a: All samples were processed at Zhejiang Zhongke Institute of Luminescence Testing Technology;

1321 b: Measured water content;

1322 c: Concentrations of U, Th and K were measured by neutron activation analysis at China institute of atomic energy;

1323 d: All De values conform to normal distribution;

1324 e: All De values were calculated with central age model (Galbraith et al., 2012).

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1326 Table 3. ^{10}Be analytical data for depth profiles at site 1 and surface boulders at site 2

Samples	Mass quartz (g)	Be Carrier (mg)	Latitude (N)	Longitude (E)	Altitude (m)	Depth (cm)	Thickness (cm)	Sample type	$^{10}\text{Be}/^9\text{Be}$	$^{10}\text{Be}/^9\text{Be}$ error (%)	$^{10}\text{Be} \pm \text{error}$ (atom/g)	Exposure age \pm uncertainty (yr)
Site 1												
T1 depth profile												
T101	19.082	0.259	39.4644	94.8154	2550	4	5	gravel	1.99743E-13	2.84493542	139215 \pm 3966	See model
T102	18.3446	0.2622	39.4644	94.8154	2550	23.5	5	gravel	2.34437E-13	3.16617704	179736 \pm 5698	/
T103	24.6633	0.2619	39.4644	94.8154	2550	38.5	5	gravel	2.79777E-13	2.81553863	165707 \pm 4672	/
T104	23.5114	0.2604	39.4644	94.8154	2550	66.5	5	gravel	2.11103E-13	3.21317004	122006 \pm 3925	/
T105	16.2753	0.2628	39.4644	94.8154	2550	112	5	gravel	1.58258E-13	4.50561624	120855 \pm 5453	/
T106	19.9296	0.2597	39.4644	94.8154	2550	162	5	gravel	1.06914E-13	4.78682755	52823 \pm 2532	/
T107	24.7446	0.2595	39.4644	94.8154	2550	202	5	gravel	1.63803E-13	3.32730129	82378 \pm 2745	/
T2 depth profile												
T201	26.985	0.2572	39.46584	94.81676	2543	146.5	5	gravel	2.88702E-13	2.80433717	156347 \pm 4390	See model
T202	18.6568	0.2644	39.46584	94.81676	2543	160	5	gravel	2.16697E-13	2.79024612	164281 \pm 4590	/
T203	29.7611	0.2625	39.46584	94.81676	2543	192.5	5	gravel	1.96218E-13	10.6339378	90175 \pm 9602	/
T204	24.0209	0.2505	39.46584	94.81676	2543	221.5	5	gravel	2.47250E-13	3.86466734	150008 \pm 5805	/
T207	29.7902	0.2592	39.46584	94.81676	2543	342.5	5	gravel	1.57479E-13	4.58115034	72964 \pm 3347	/
T3 depth profile												
T301	30.002	0.2603	39.46587	94.81474	2571	132.5	5	gravel	1.79830E-12	5.06926063	1019912 \pm 51771	See model
T302	30.1499	0.2621	39.46587	94.81474	2571	162.5	5	gravel	1.77618E-12	1.48973645	1009077 \pm 15053	/
T305	30.1379	0.261	39.46587	94.81474	2571	252.5	5	gravel	5.58156E-13	7.55396228	300382 \pm 22721	/
T4 depth profile												
T402	30.0074	0.265	39.46608	94.81159	2616	68.5	5	gravel	3.13757E-12	1.77849721	1831441 \pm 32615	See model
T403	31.7724	0.2616	39.46608	94.81159	2616	98.5	5	gravel	3.08259E-12	1.3182097	1677233 \pm 22139	/
T404	29.6724	0.2619	39.46608	94.81159	2616	131.5	5	gravel	1.98844E-12	1.77949615	1152668 \pm 20539	/
T405	31.3068	0.2554	39.46608	94.81159	2616	161	5	gravel	1.25525E-12	4.89009314	665693 \pm 32596	/
T406	30.4697	0.2693	39.46608	94.81159	2616	191	5	gravel	1.29302E-12	3.78535521	743514 \pm 28182	/
T407	32.4745	0.2627	39.46608	94.81159	2616	221	5	gravel	5.47390E-13	2.53345878	277465 \pm 7039	/
Site 2 surface												
LP01	30.7098	0.308	39.47123	94.90786	2265	0	2.5	boulder	1.02453E-12	6.32872596	670639 \pm 42500	28133 \pm 3040
LP02	30.2138	0.3154	39.47112	94.90761	2265	0	2.5	boulder	5.48760E-13	5.84434137	366152 \pm 21428	15826 \pm 1661

LP03	27.7759	0.3149	39.47099	94.90705	2266	0	2.5	boulder	5.32490E-13	4.19369166	385332±16181	16603±1605
LP04	19.6304	0.3159	39.4708	94.90692	2267	0	2.5	boulder	3.52189E-13	3.21520471	353072±11367	15268±1416
LP05	23.0298	0.3157	39.47078	94.90681	2269	0	2.5	boulder	9.16409E-13	4.03025204	817601±32995	33799±3257
LP06	24.787	0.3178	39.47128	94.90797	2265	0	2.5	boulder	4.85817E-13	6.75410724	395787±26767	17042±1880

Note: AMS measurements were performed at ASTER in Centre de Recherche et d'Enseignement de Géosciences de l'Environnement (CEREGE), Aix-en-Provence, France. Surface ages at site 2 are calculated with the CRONUS 2.3 calculator (<http://hess.ess.washington.edu/>), with time-dependent Lal (1991) / Stone (2000) scaling scheme for spallation, sample density 2.7 g/cm³, and no erosion. Reference production rate of ¹⁰Be is 4.00±0.29 atoms / gram of quartz /year. (Borchers et al., 2016).

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1333 Table 4. Radiocarbon analytical data

Sample No.	Percent Modern Carbon (pMC)	Fraction Modern	D14C	Conventional age	2 sigma calibration ^c
14C-1517 ^a	40.2 +/-0.2 pMC	0.4015±0.0015	-598.5 +/-1.5	7330 +/-30 BP	Cal BC 6240 to 6090 (Cal BP 8190 to 8040)
STH48c ^b		0.3901±0.0112		7500±240	8473±512 years B.P.
STH48d ^b		0.3605±0.0031	-639.5±3.1	8200±80	9163±158 years B.P.
STH48e ^b		0.3854±0.0044	-614.6±4.4	7660±100	8478±104 years B.P.
					Mean: 8705±557 years B.P.

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a: Sample AMS dated at Beta Analytic;

b: Re-calibrated radiocarbon dates from van der Woerd et al. (2001) measured at Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory (Livermore, California, USA) in 1998.

c Calibration with IntCal13 (Reimer et al., 2013), not including probability distribution of range < 0.1.

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Table 5. ^{10}Be depth profiles modeling parameters

Terrace name	Inheritance (atoms / g)	Stage 1	Stage 2	Stage 3	
T1	150000	8.7 - 6 ka	6 - 5.6 ka	5.6 - 0 ka	
T2	200000	9.5 - 6.1 ka	6.1 - 4.3 ka	4.3 - 0 ka	
T3	Model 1	4000	67.6 - 7.6 ka	7.6 - 4.1 ka	4.1 - 0 ka
	Model 2	86000	41.6 - 7.6 ka	7.6 - 4.1 ka	4.1 - 0 ka
T4	Model 1	10000	128 - 8.2 ka	8.2 - 0 ka	

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Table 6. Shortening estimates from models 1 and 2 at site 1.

	Age (ka)	Height/T0 (m)	Vertical offset (m)	Min height /paleobaselevel (m)	Max height /paleobaselevel (m)	Area (m ²)	Depth (m)	Depth (m)	Ramp (°)	Slip on ramp (m)	Shortening rate (mm/yr)	Vertical rate /T0 (mm/yr)
		U _{max}				S1/A1	D α	D0	α	S		
Model 1												
T1	8.7±0.5	7.5±0.5	na			7138	1000	1000	0	7.1±0.5	0.84±0.1	0.82±0.09
T2	9.5±1	23±1	15±1			9190	1000	1000	0	9.2±0.5	0.97±0.12	2.42±0.11
T3	51.5±13.5	40±2	>31±2			33903	1000	1000	0	33.9±2	0.64±0.27	0.78±0.27
T4	128±20	60±3	>50±3			55316	1000	1000	0	55.3±3	0.51±0.16	0.55±0.16
Model 2				U	U _{max}	S2/A2						
T1	8.7±0.5	7.5±0.5	na	1.5±0.5	8.5±0.5	9933	1889	1889	23.5	3.8±0.5	0.44±0.14	0.82±0.09
T2	9.5±1	23±1	15±1	1.9±0.5	13±1	13817	1889	1889	23.5	4.8±0.5	0.50±0.15	2.42±0.11
T3	51.5±13.5	40±2	>31±2	13±1	56±3	59480	1889	1889	23.5	31.3±2	0.61±0.27	0.78±0.27
T4	128±20	60±3	>50±3	22±1	81±3	104684	1889	1889	23.5	55.2±3	0.43±0.16	0.55±0.16

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Figure 1.

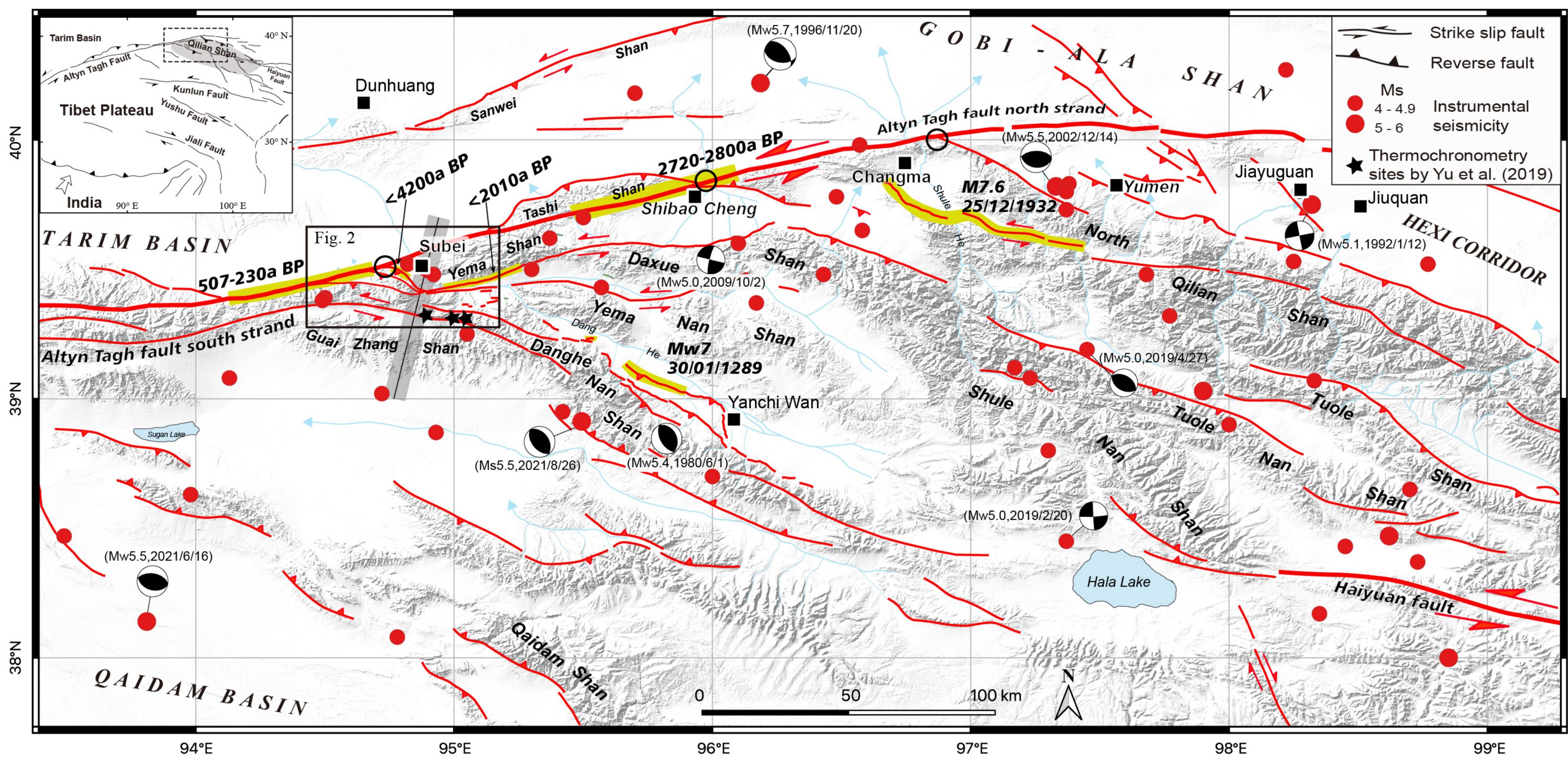


Figure 2.

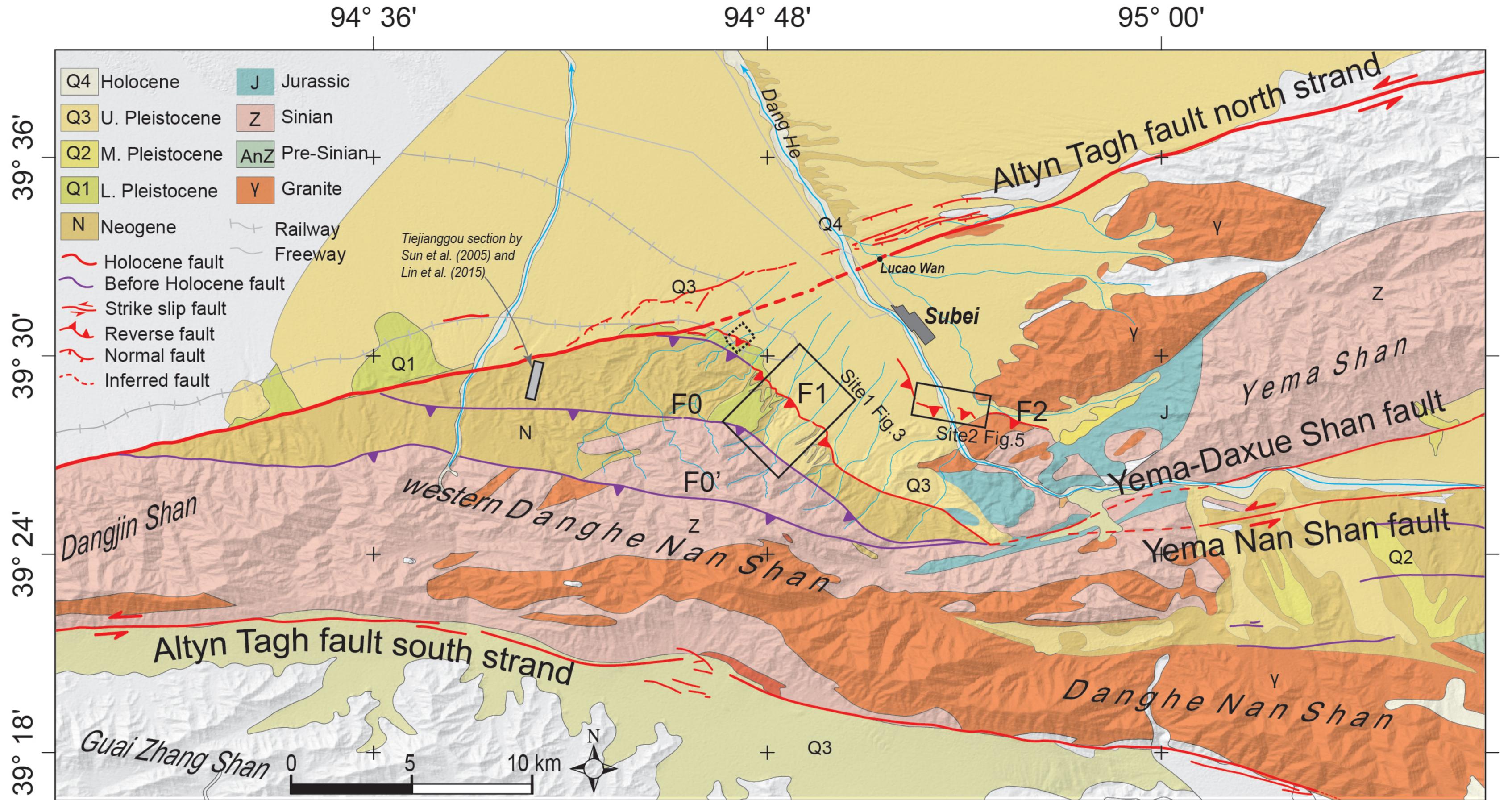
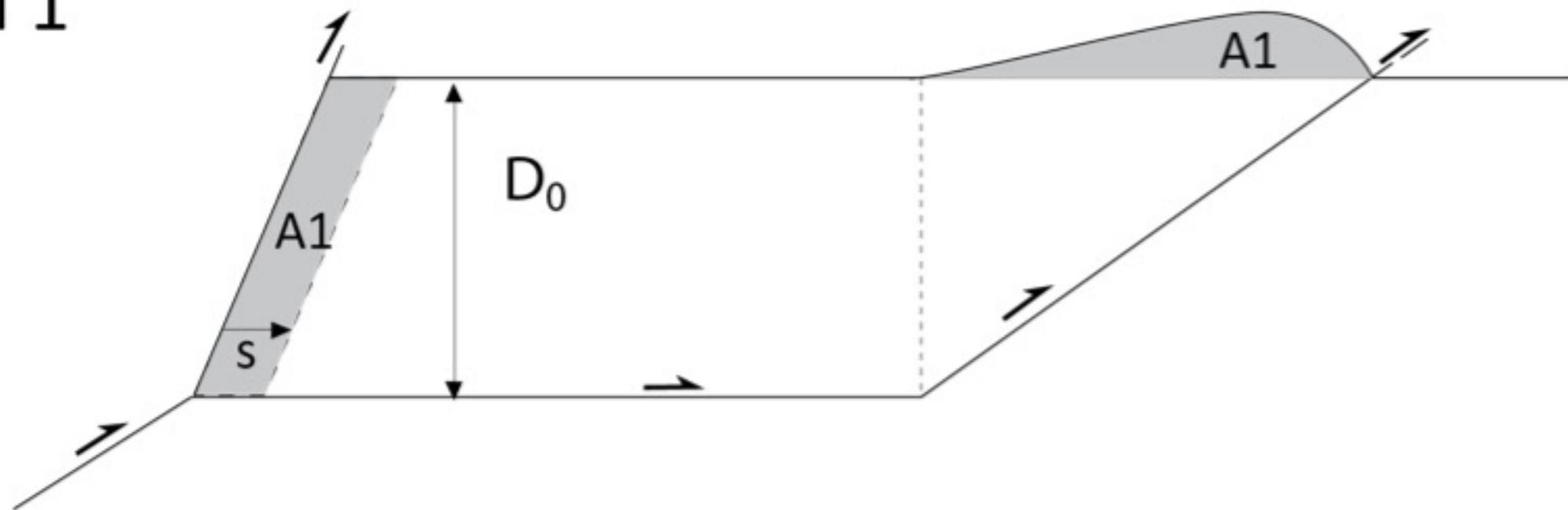


Figure 3.

Model 1



Model 2

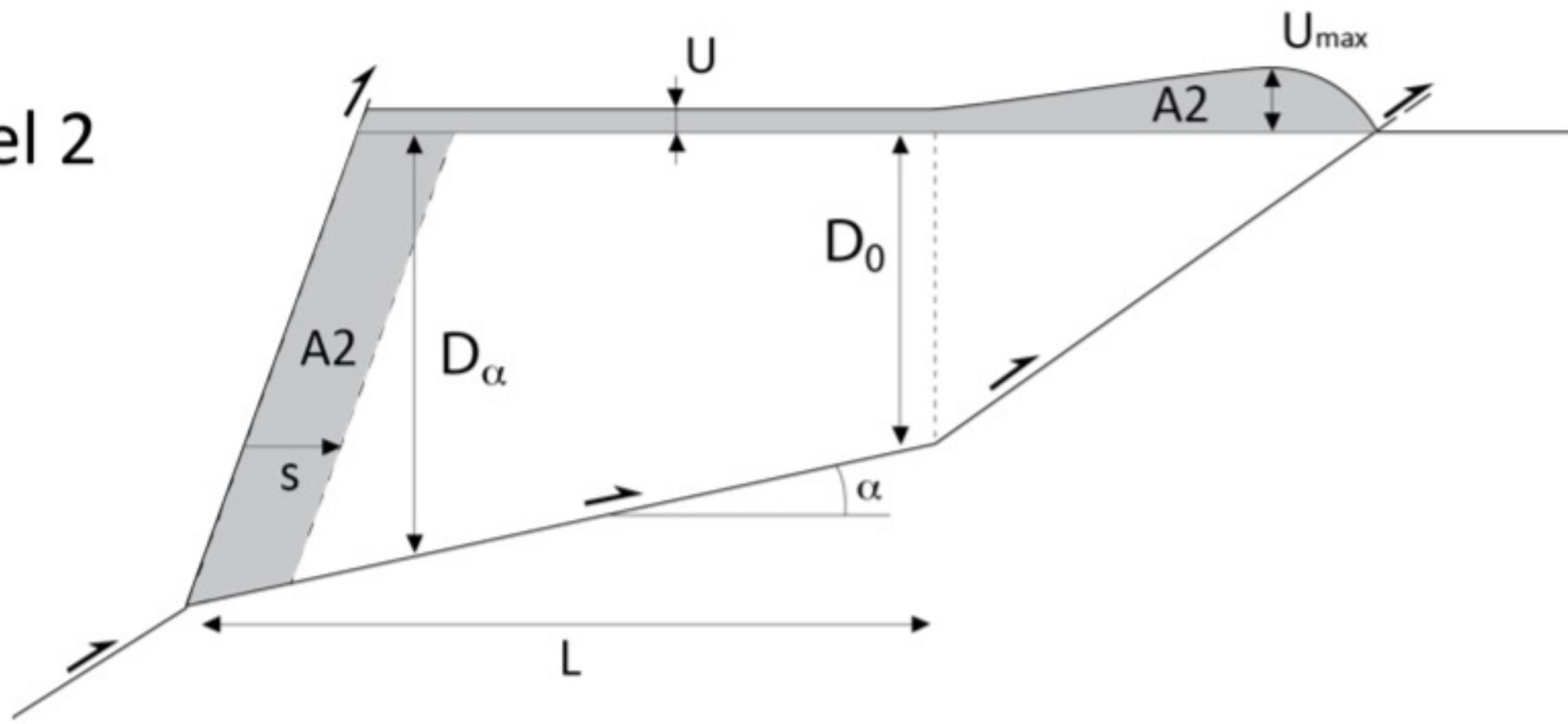


Figure 4.

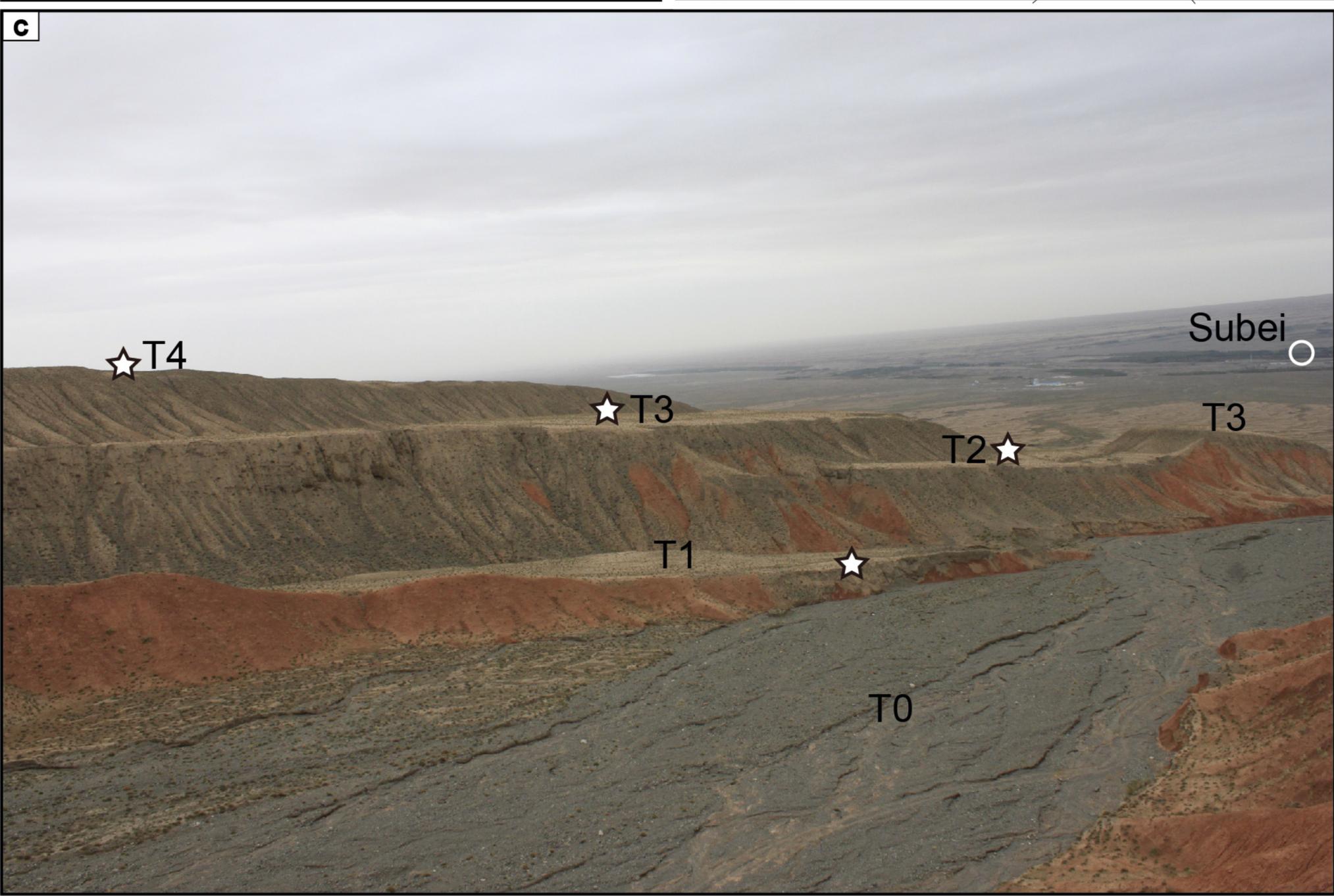
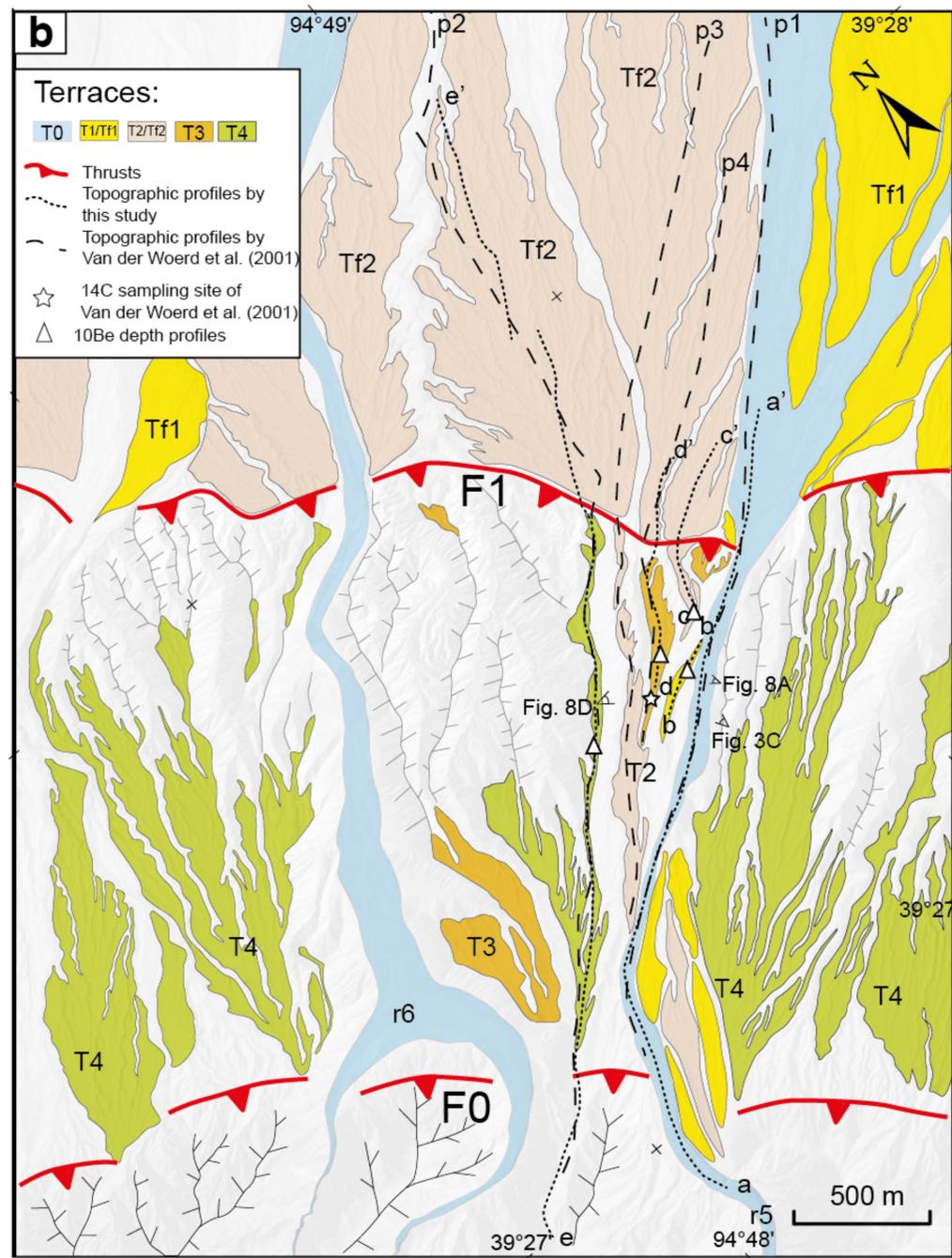
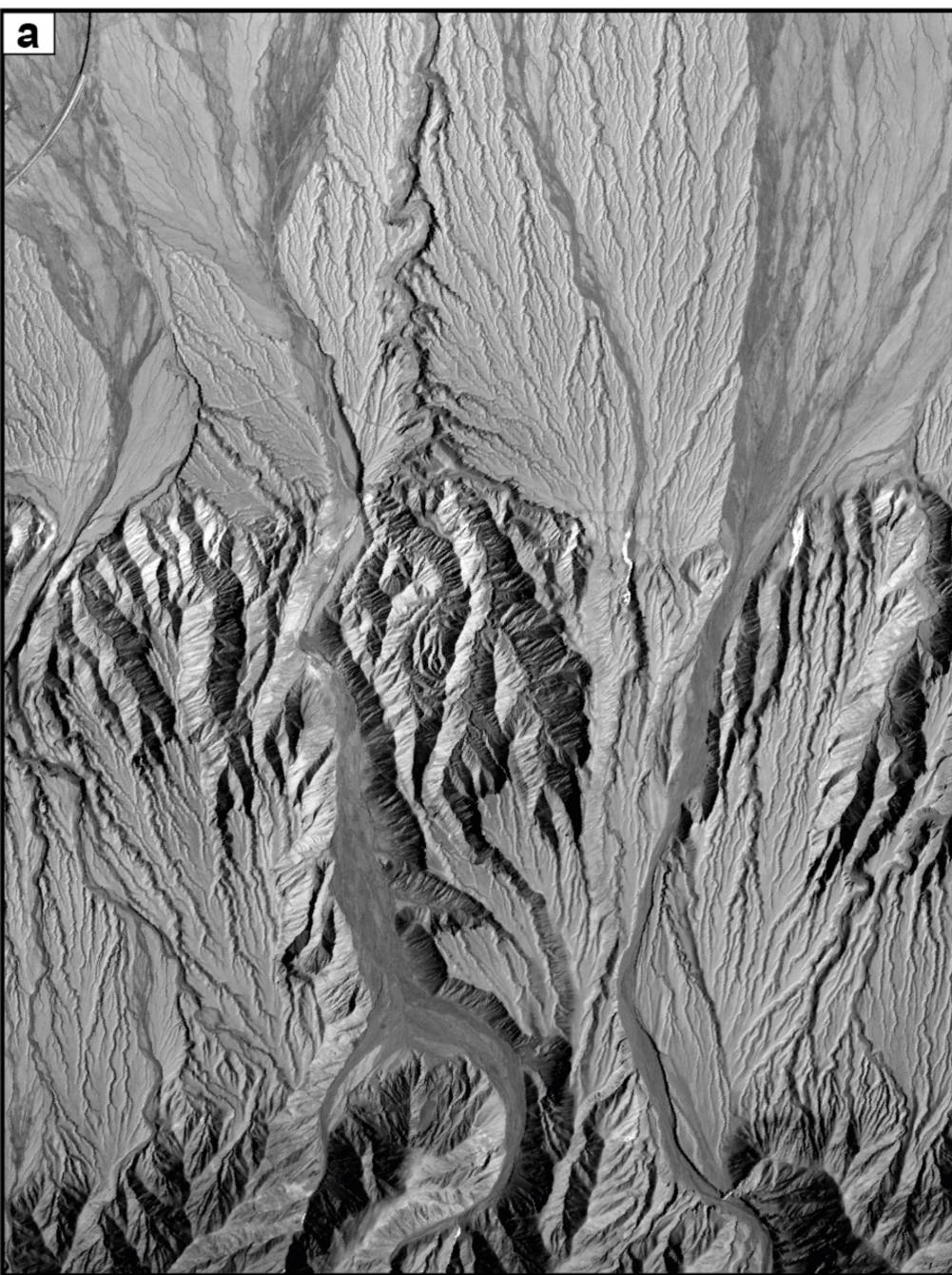


Figure 5.

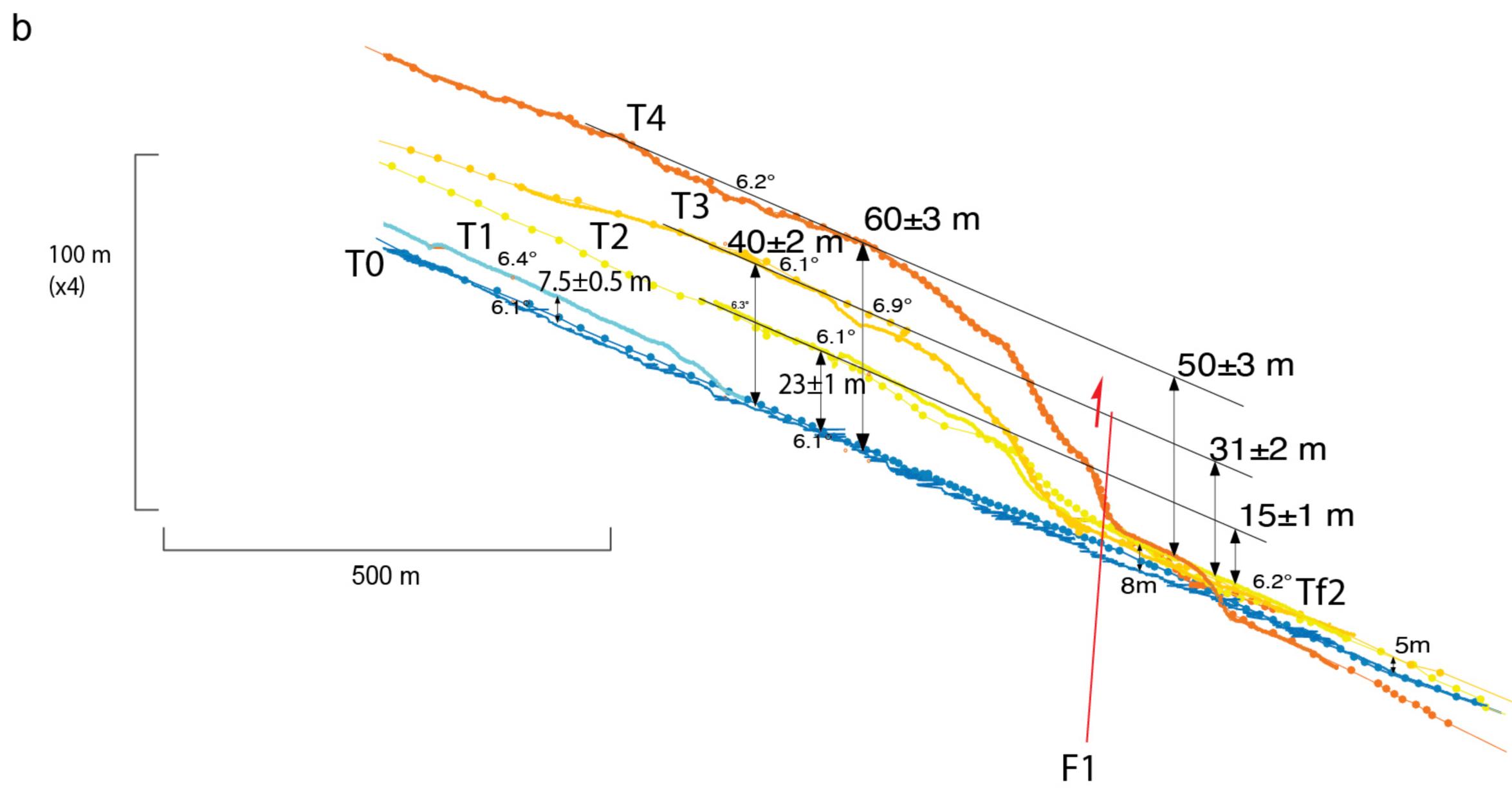
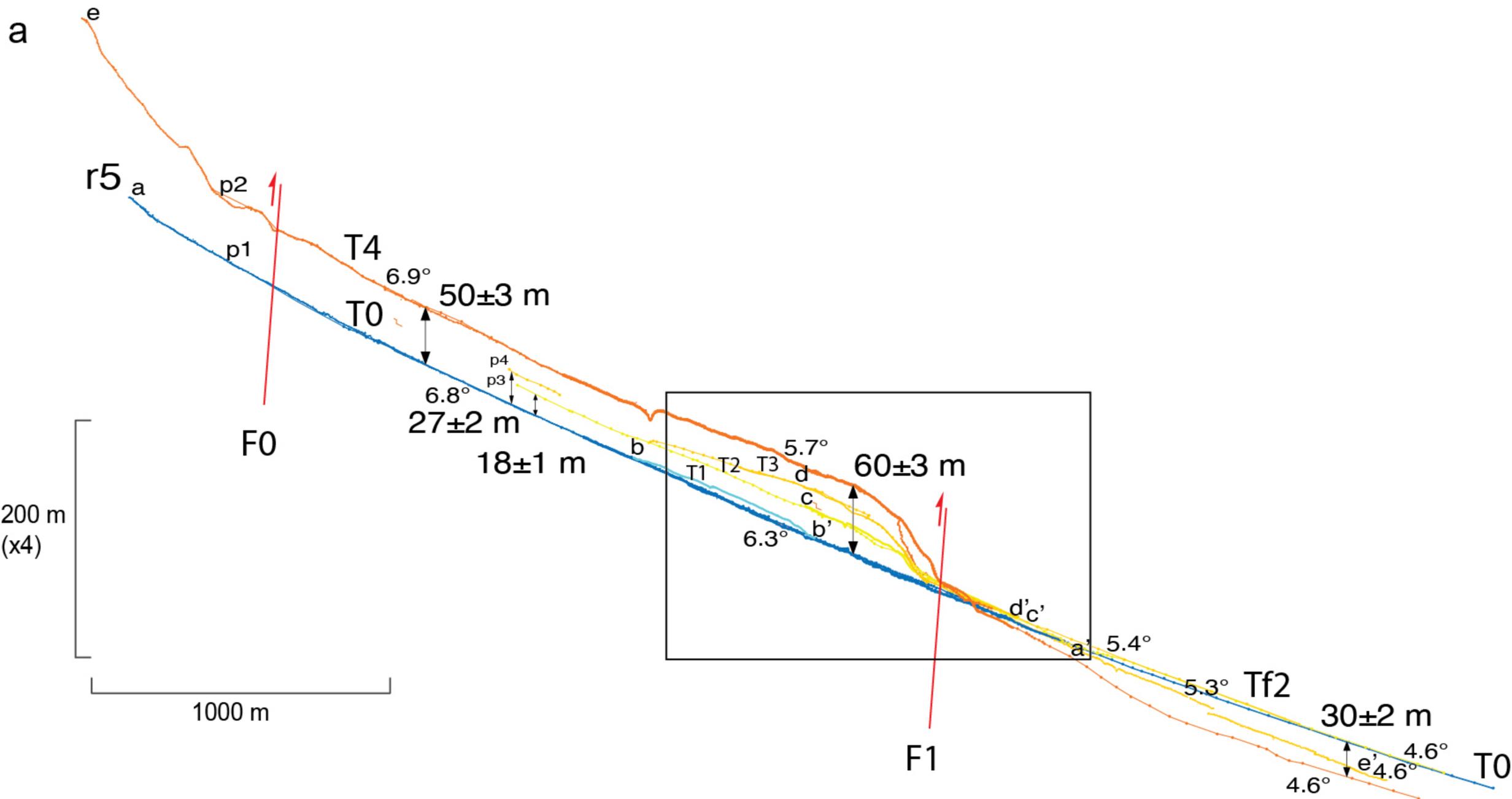


Figure 6.

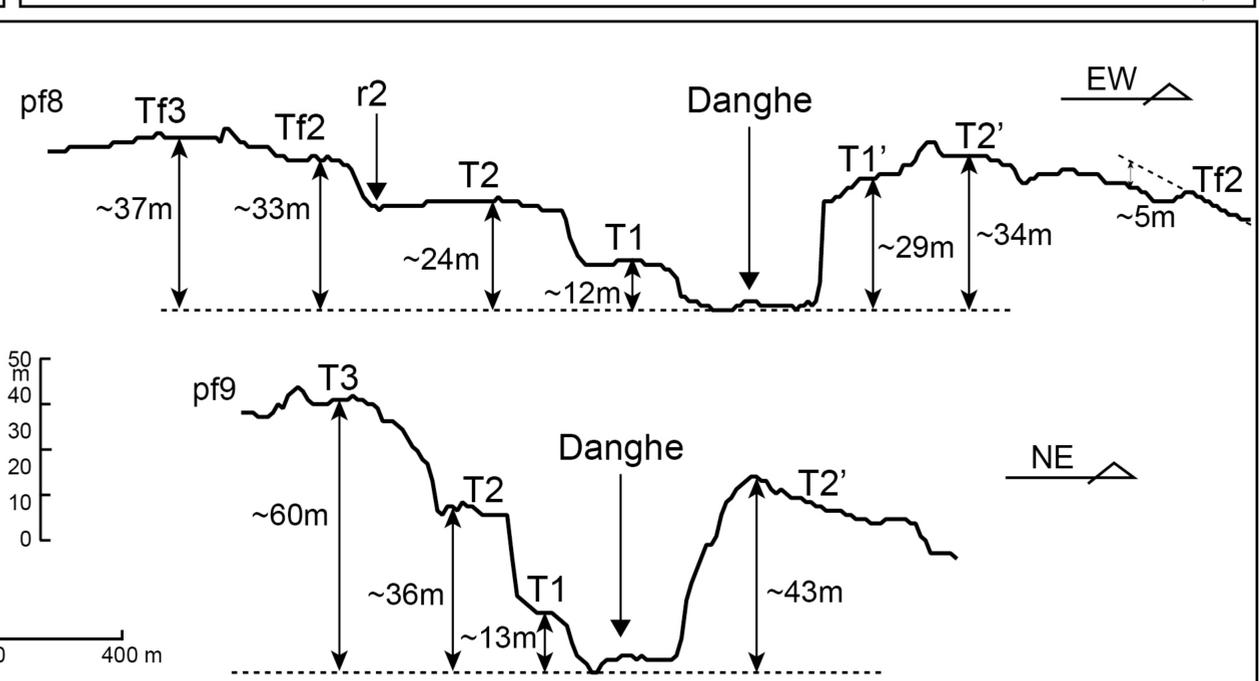
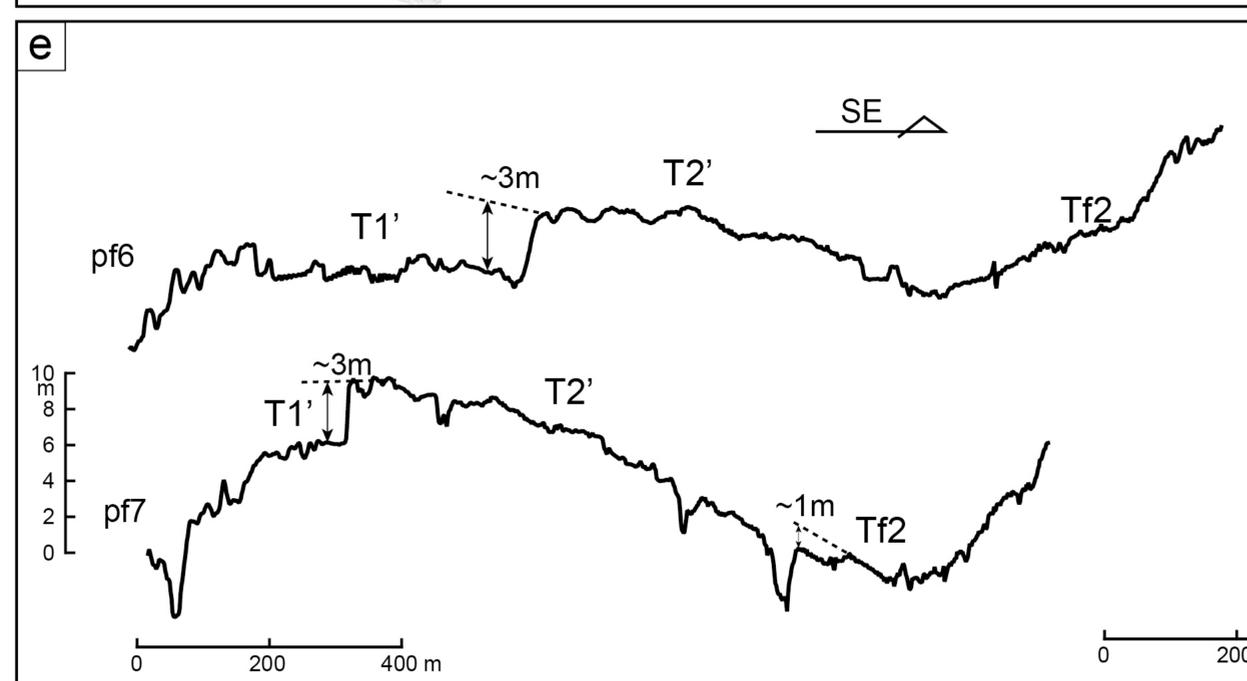
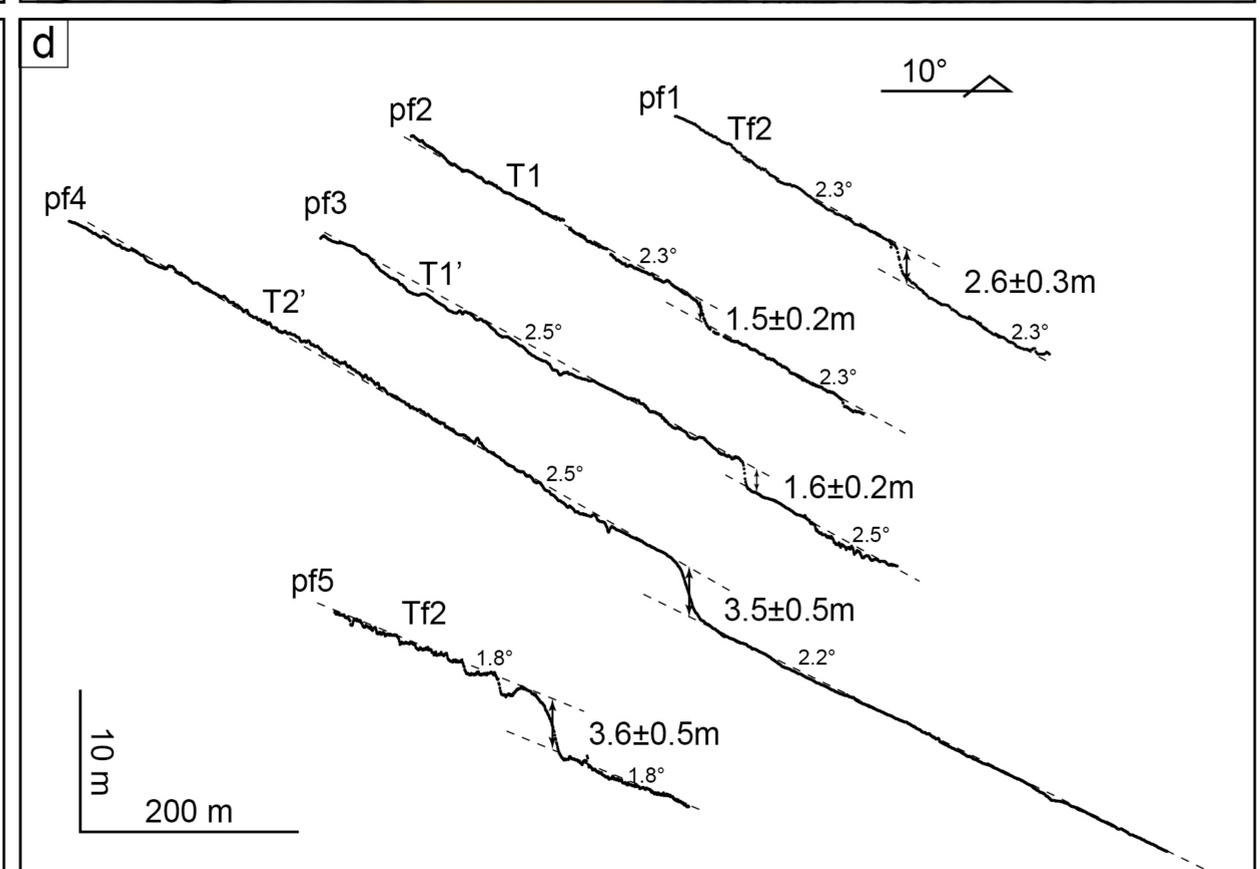
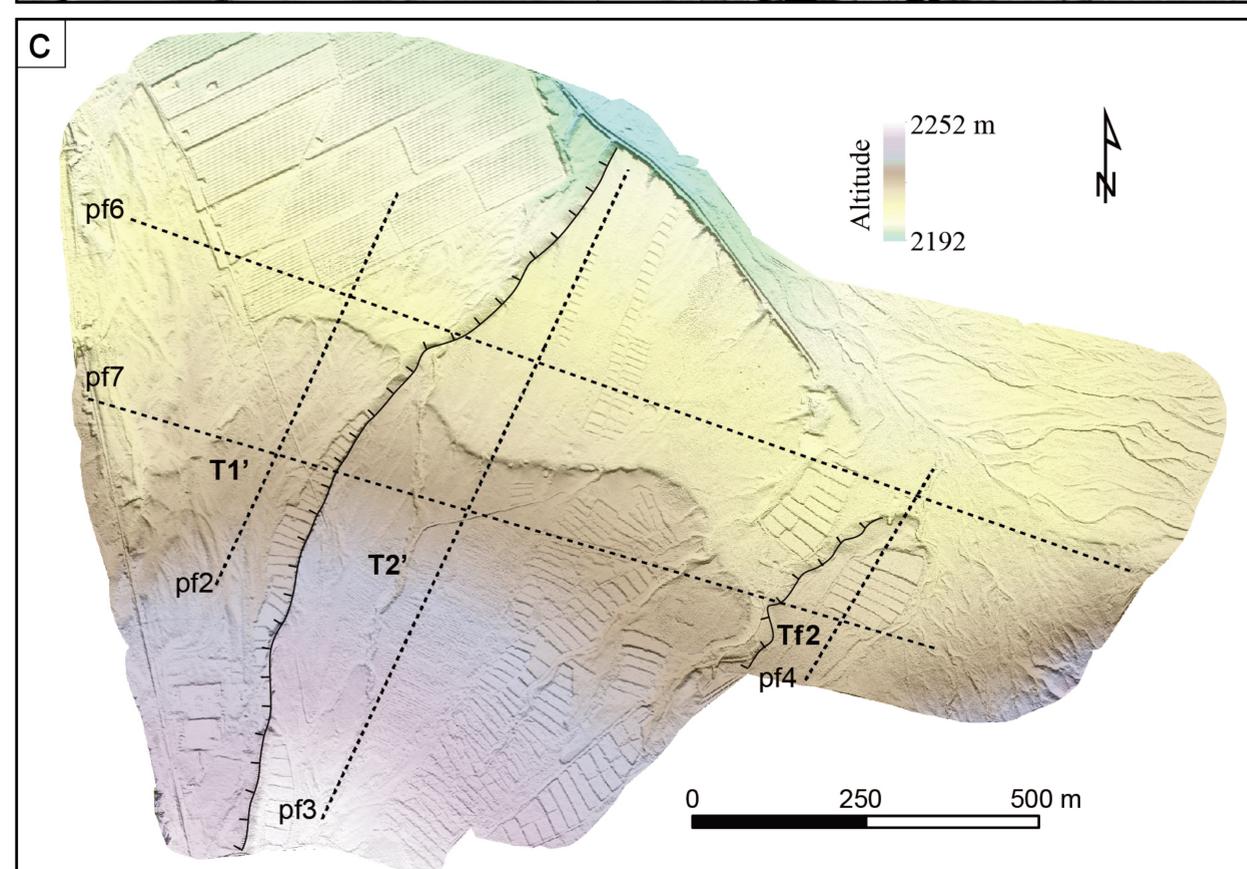
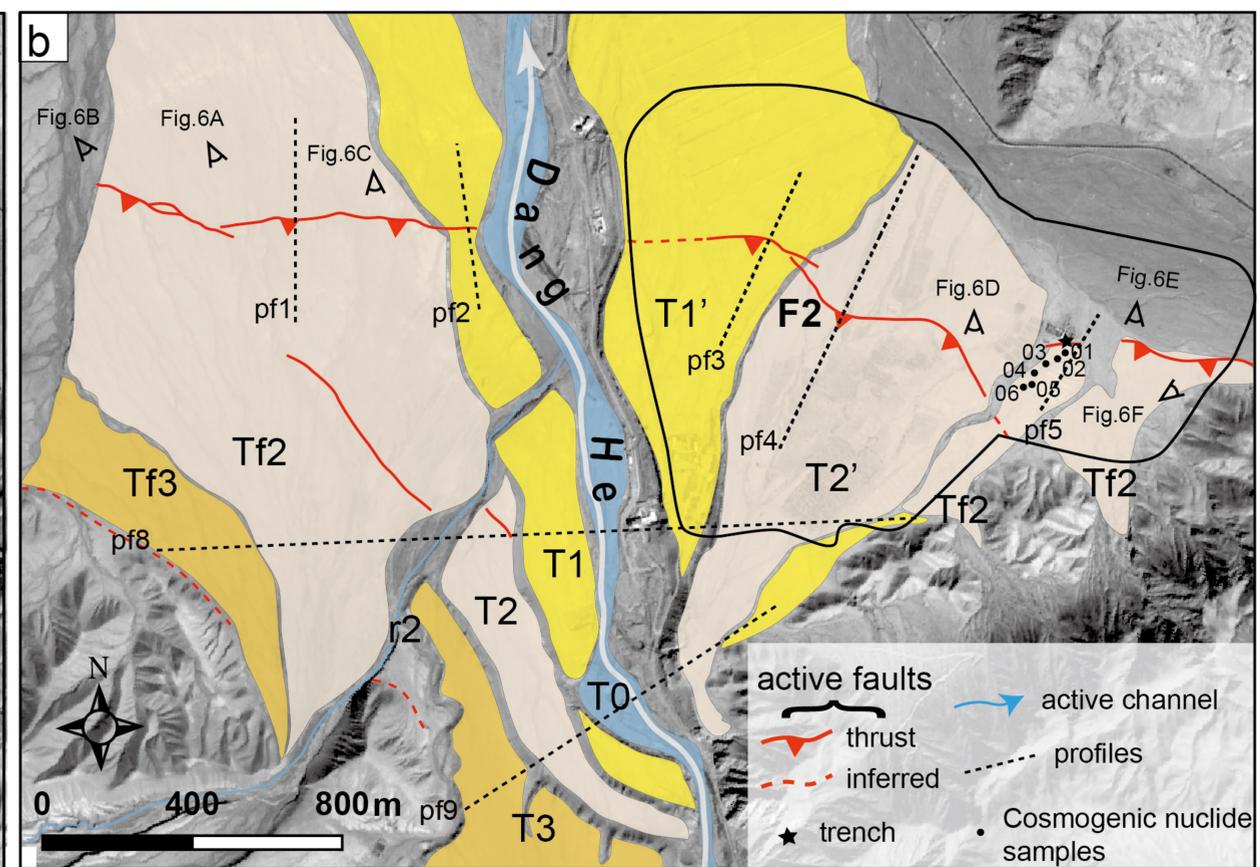
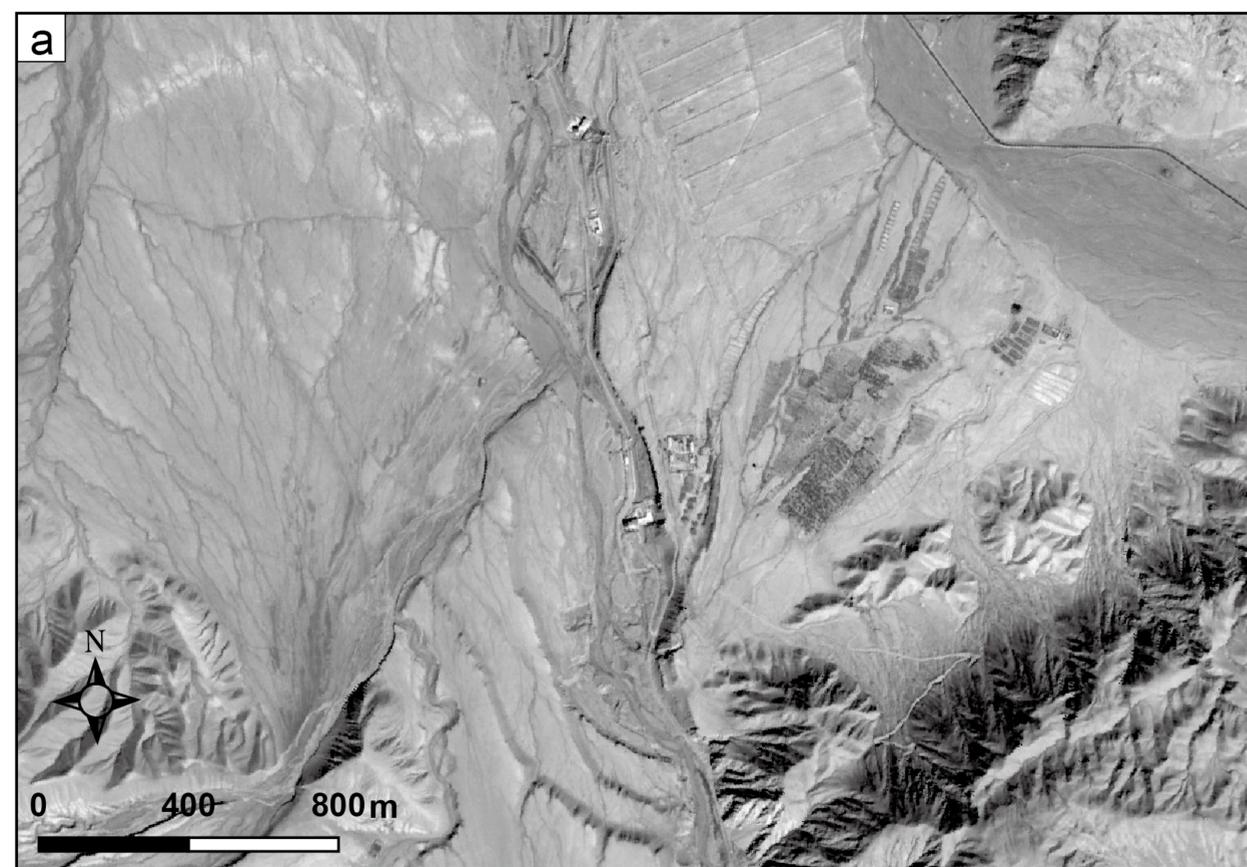


Figure 7.

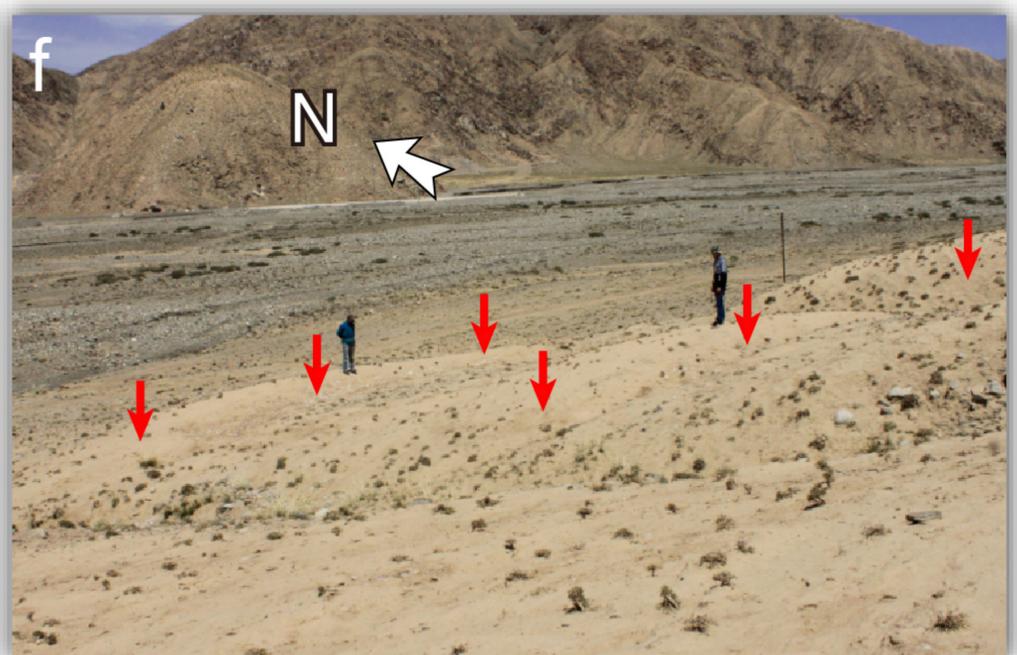
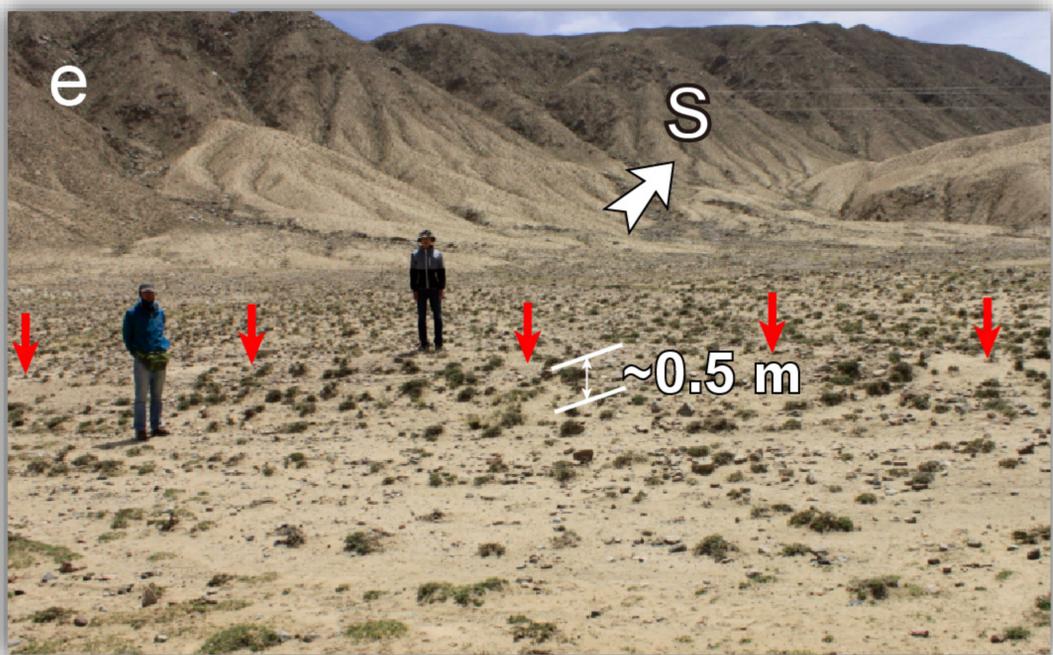
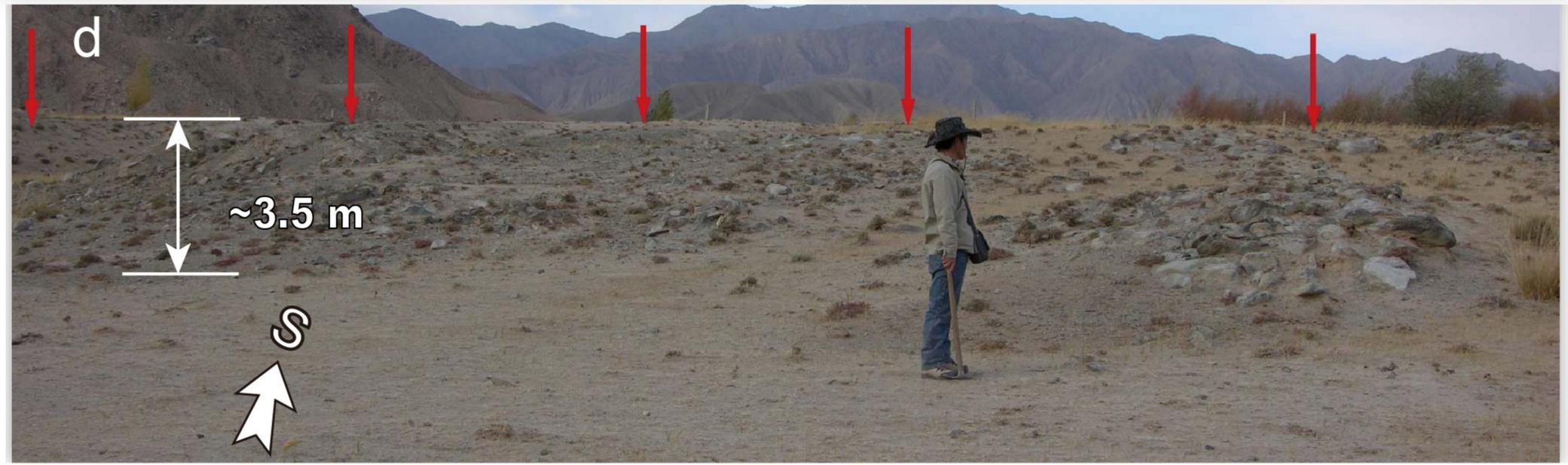
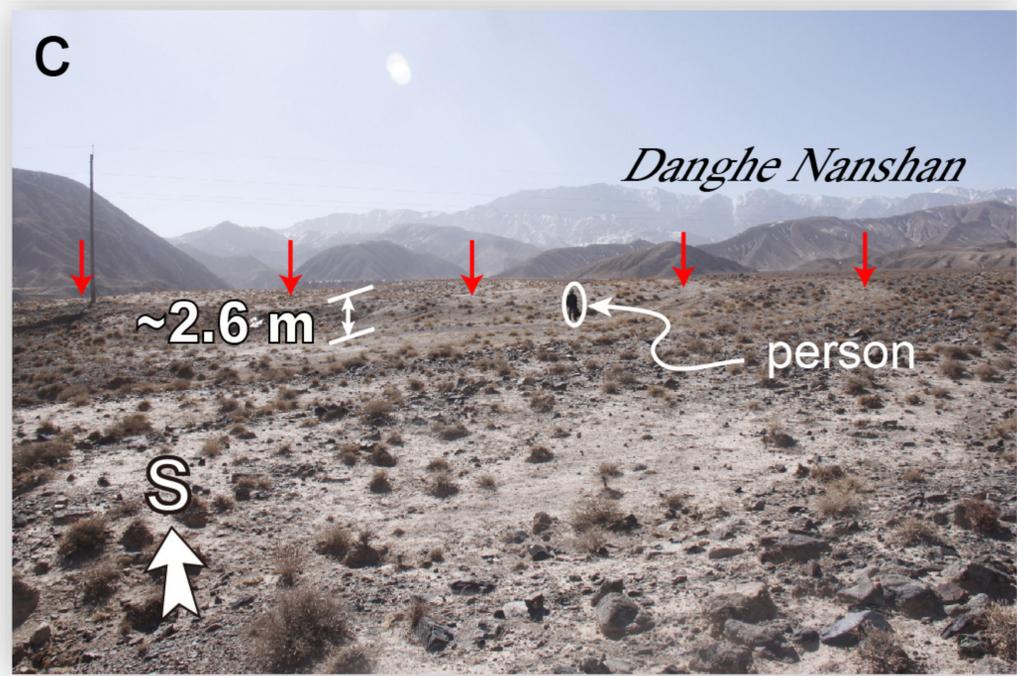
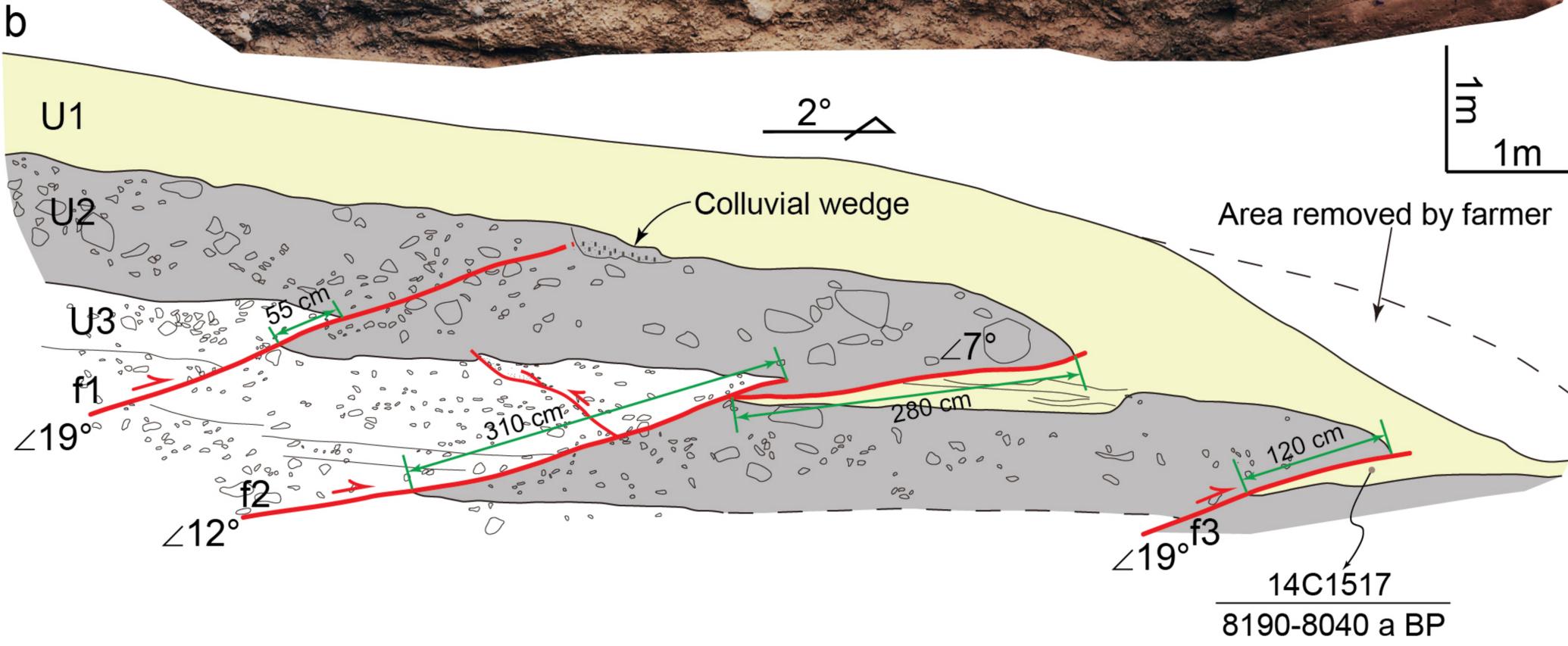


Figure 8.

West wall



East wall

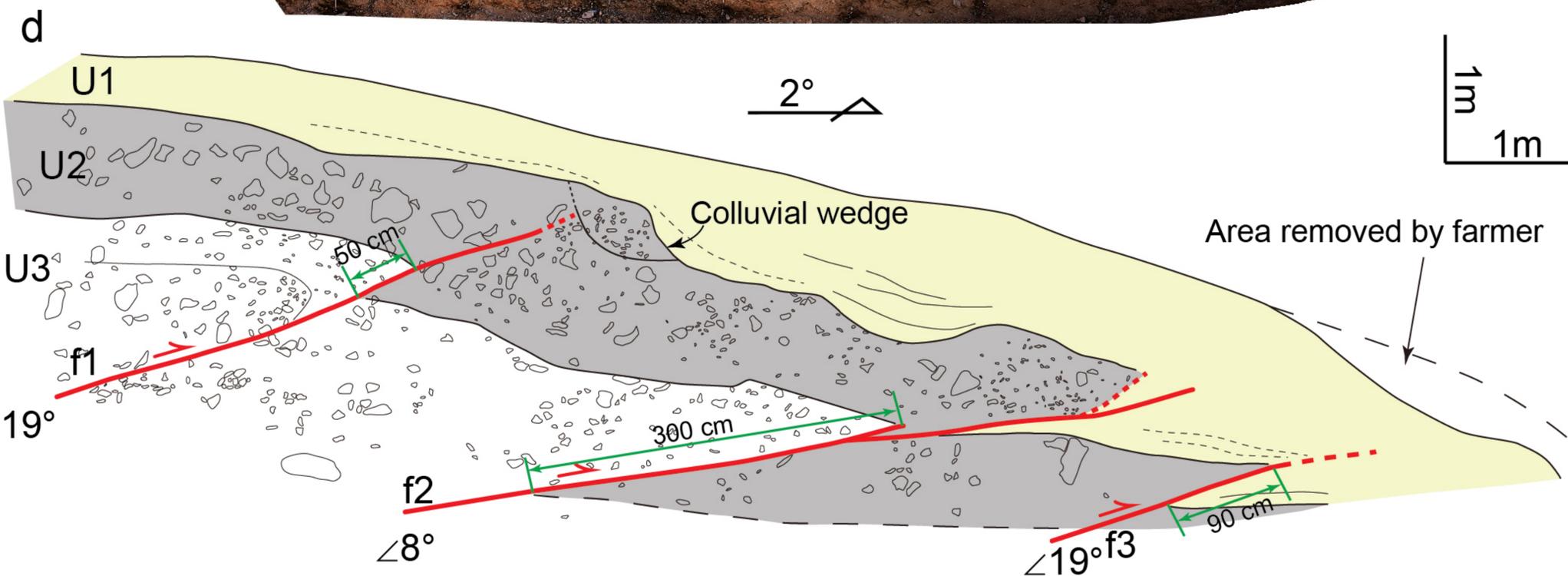


Figure 9.

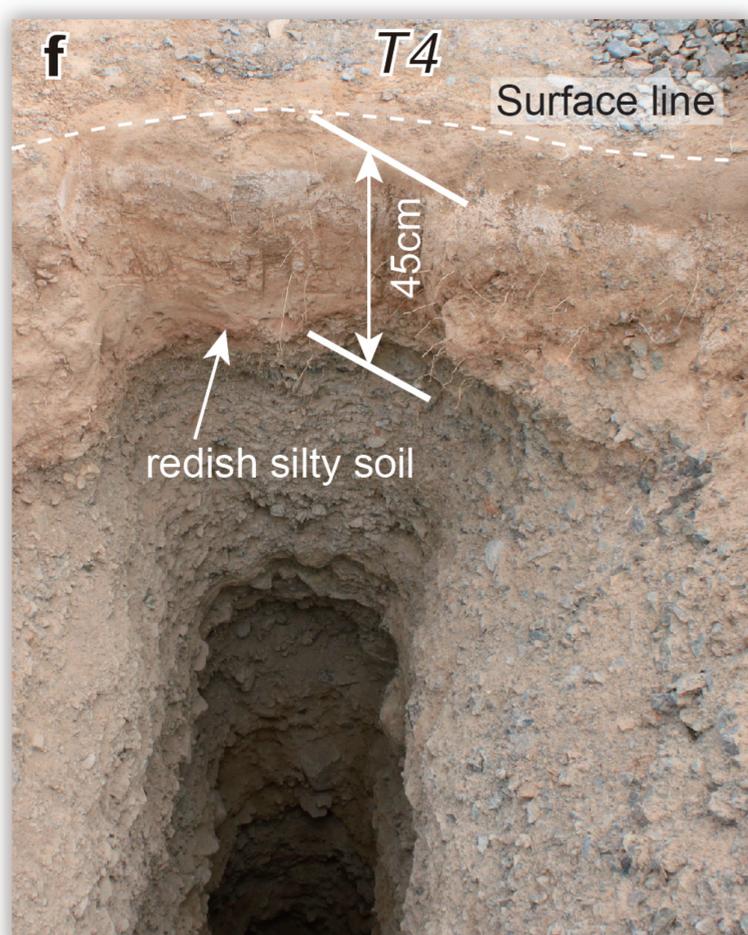
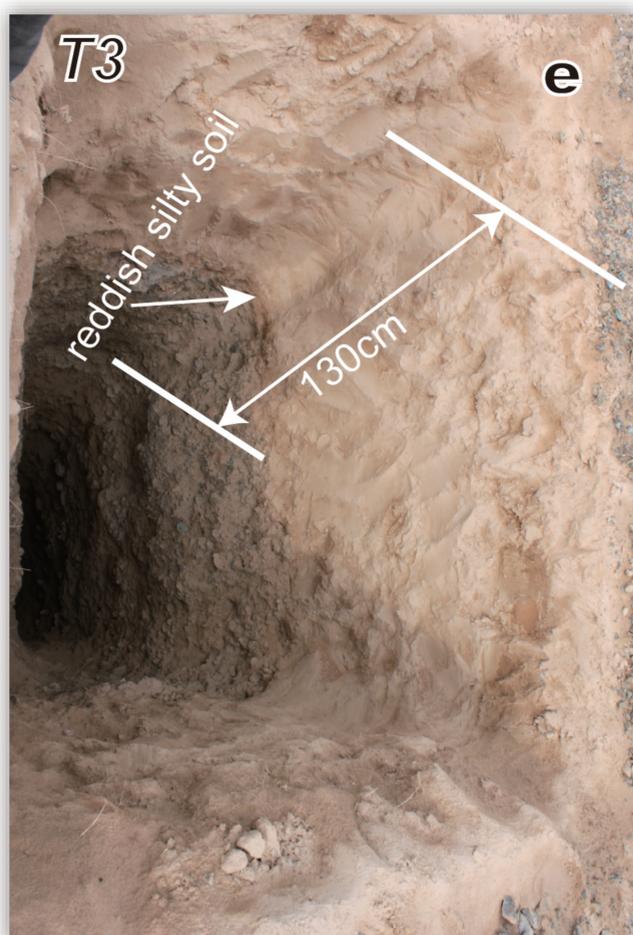
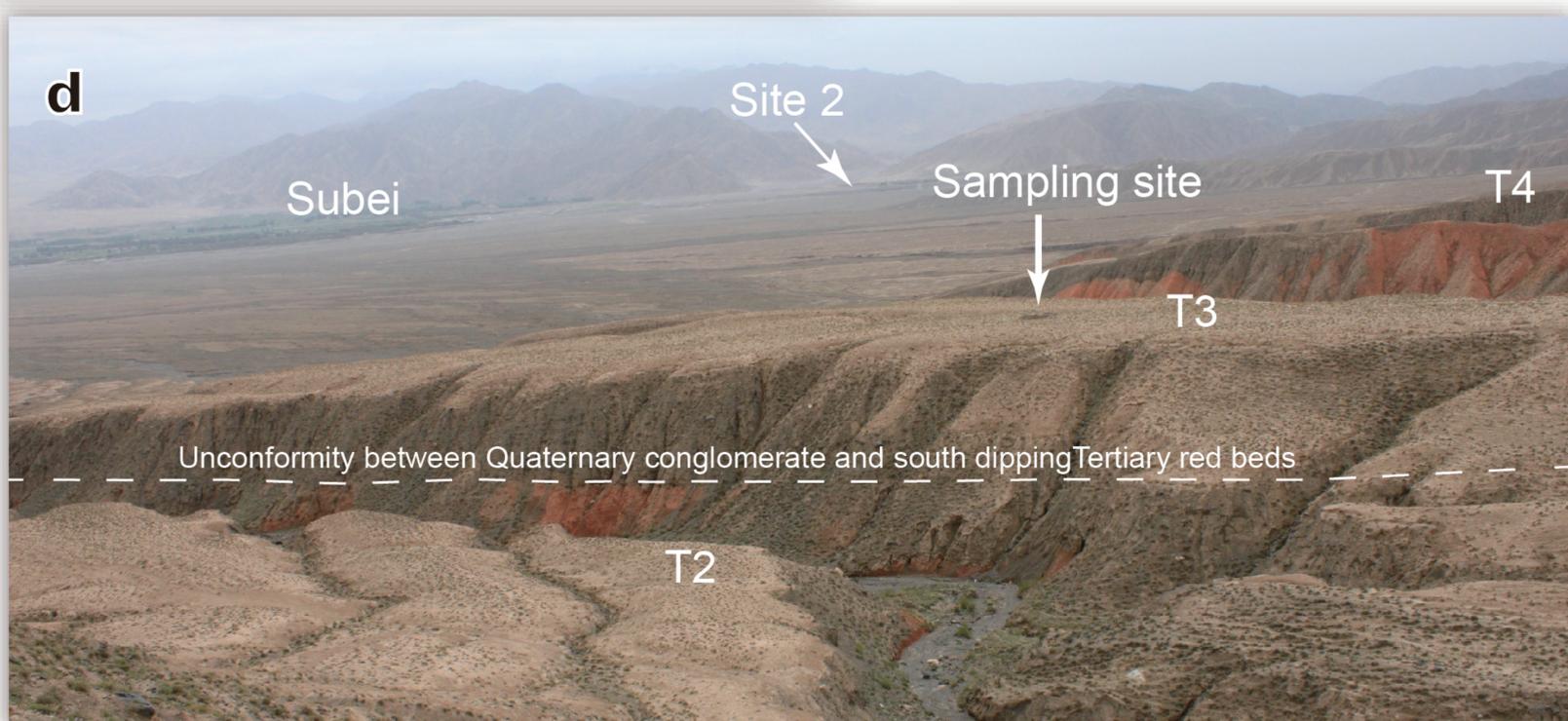
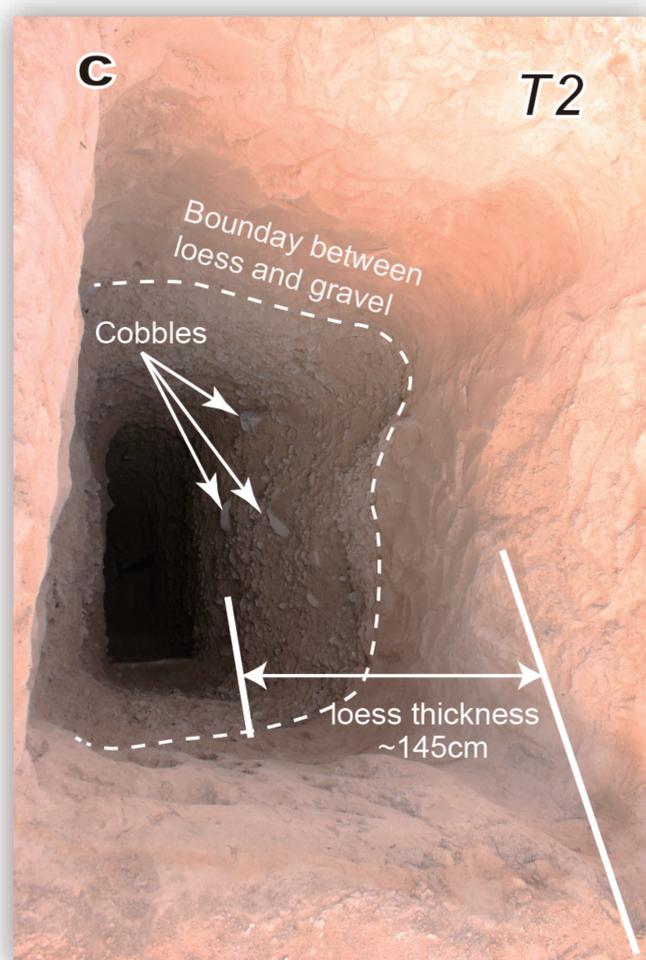
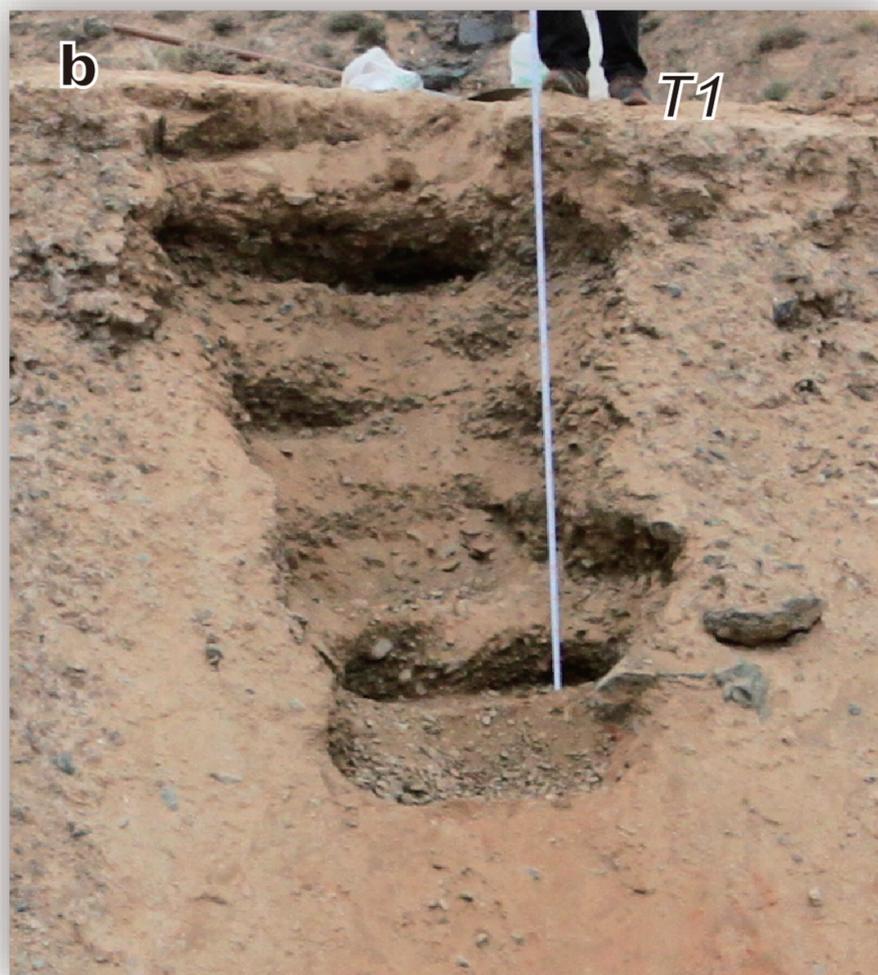
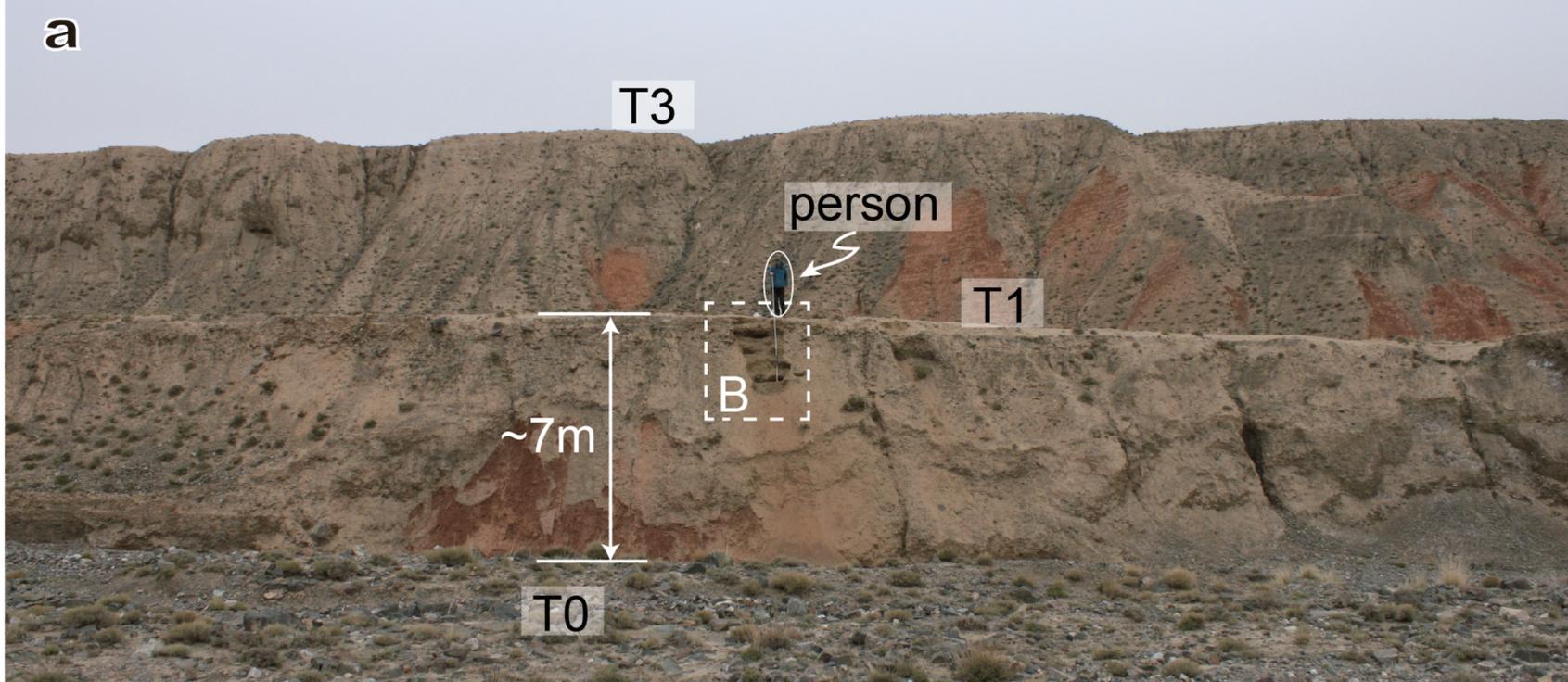


Figure 10.

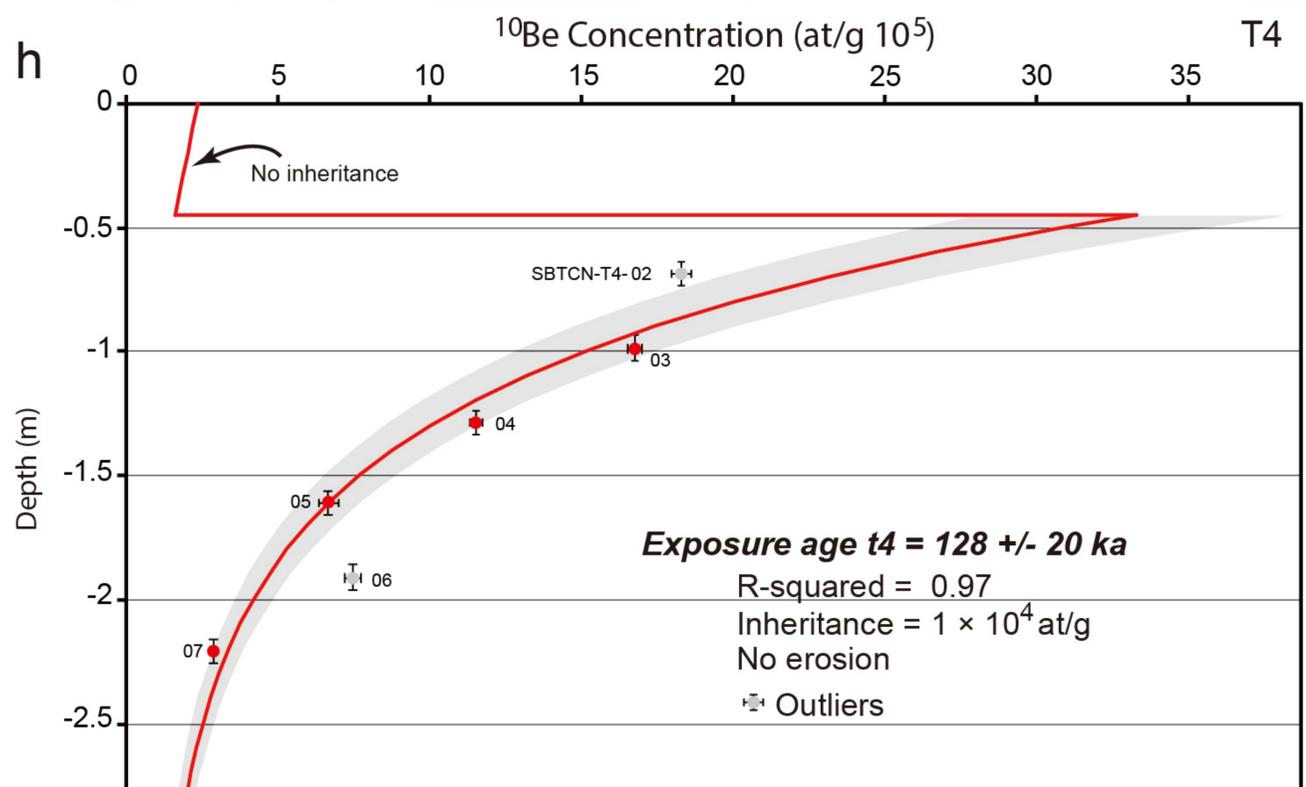
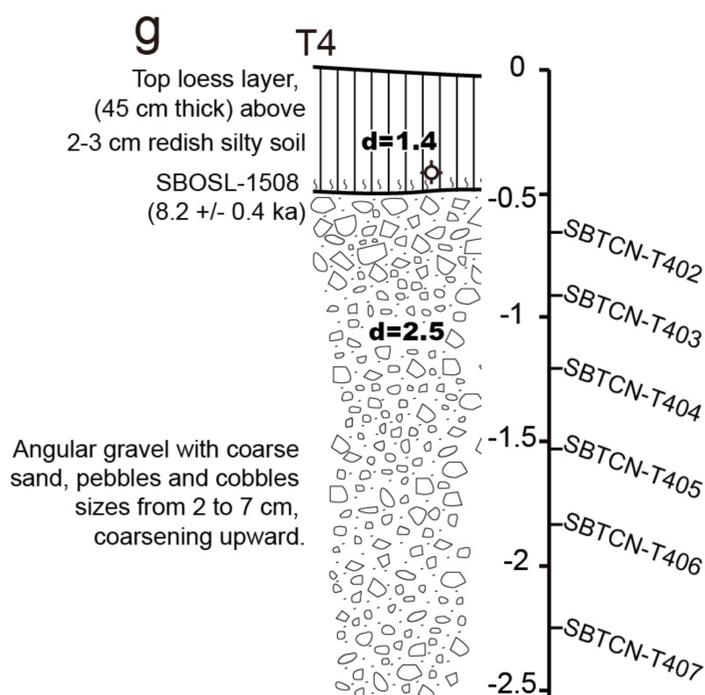
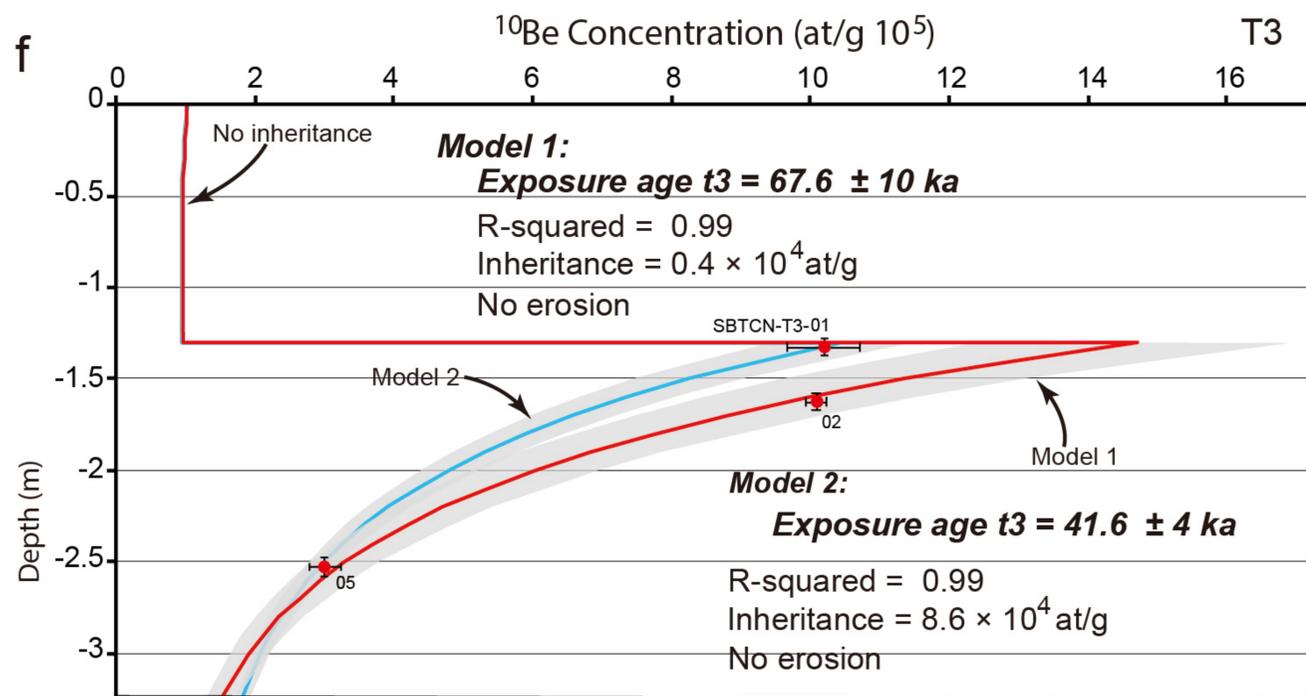
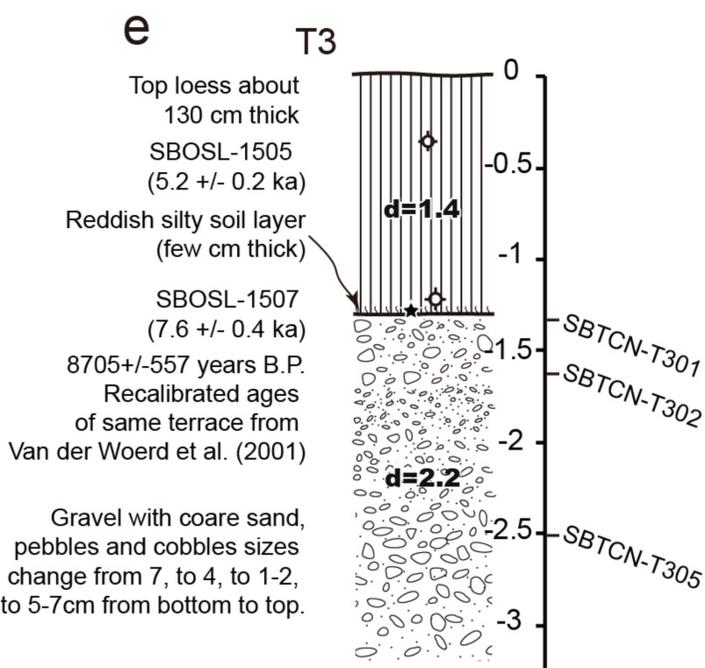
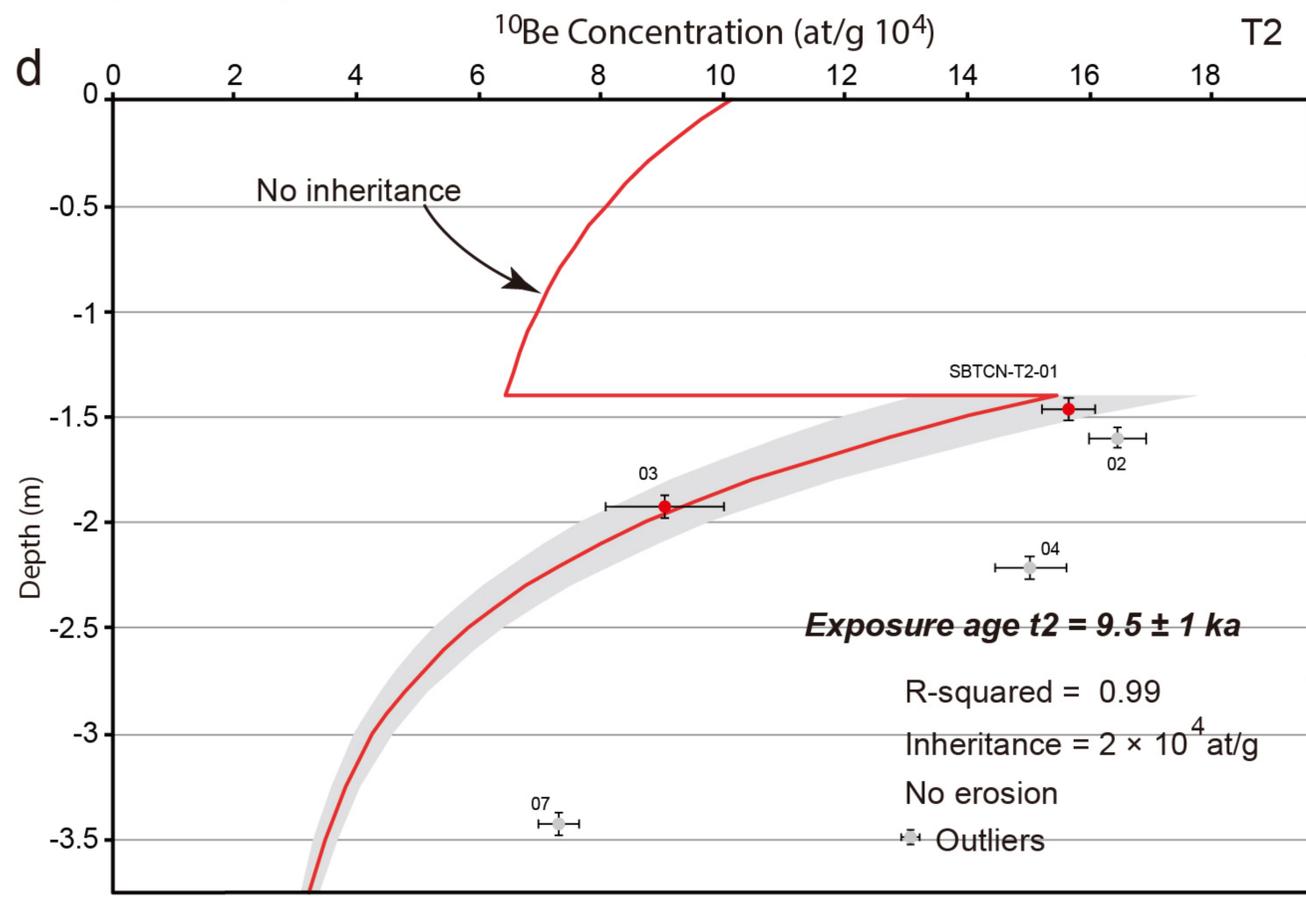
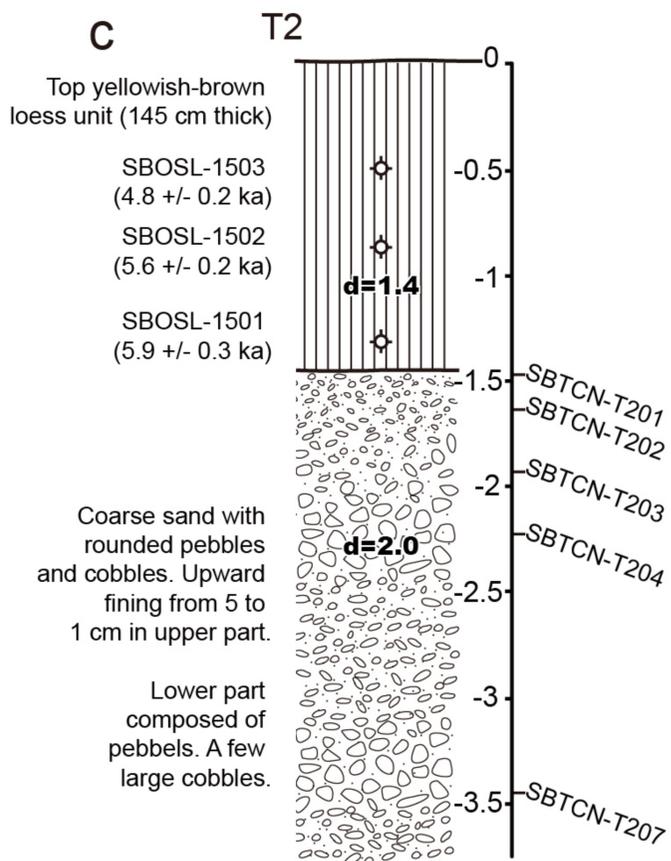
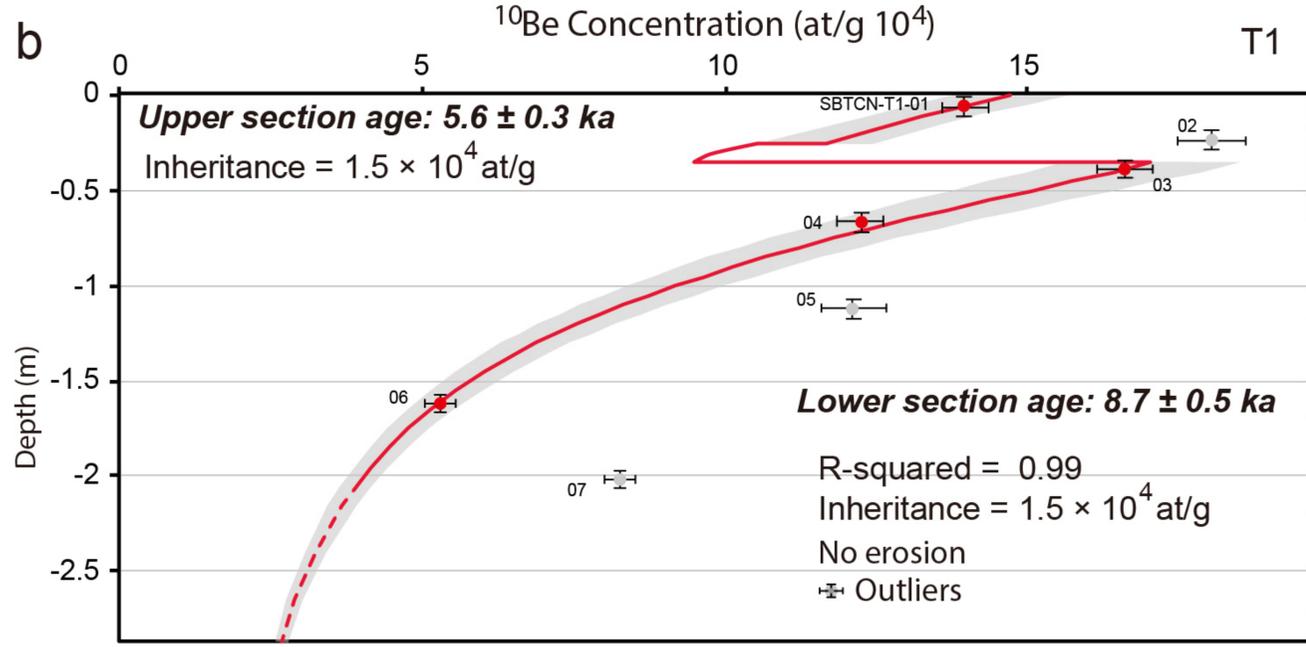
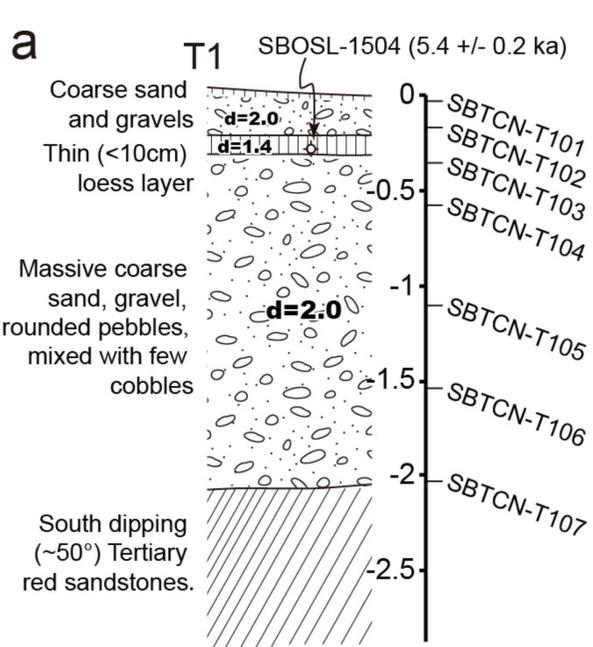


Figure 11.

a



b

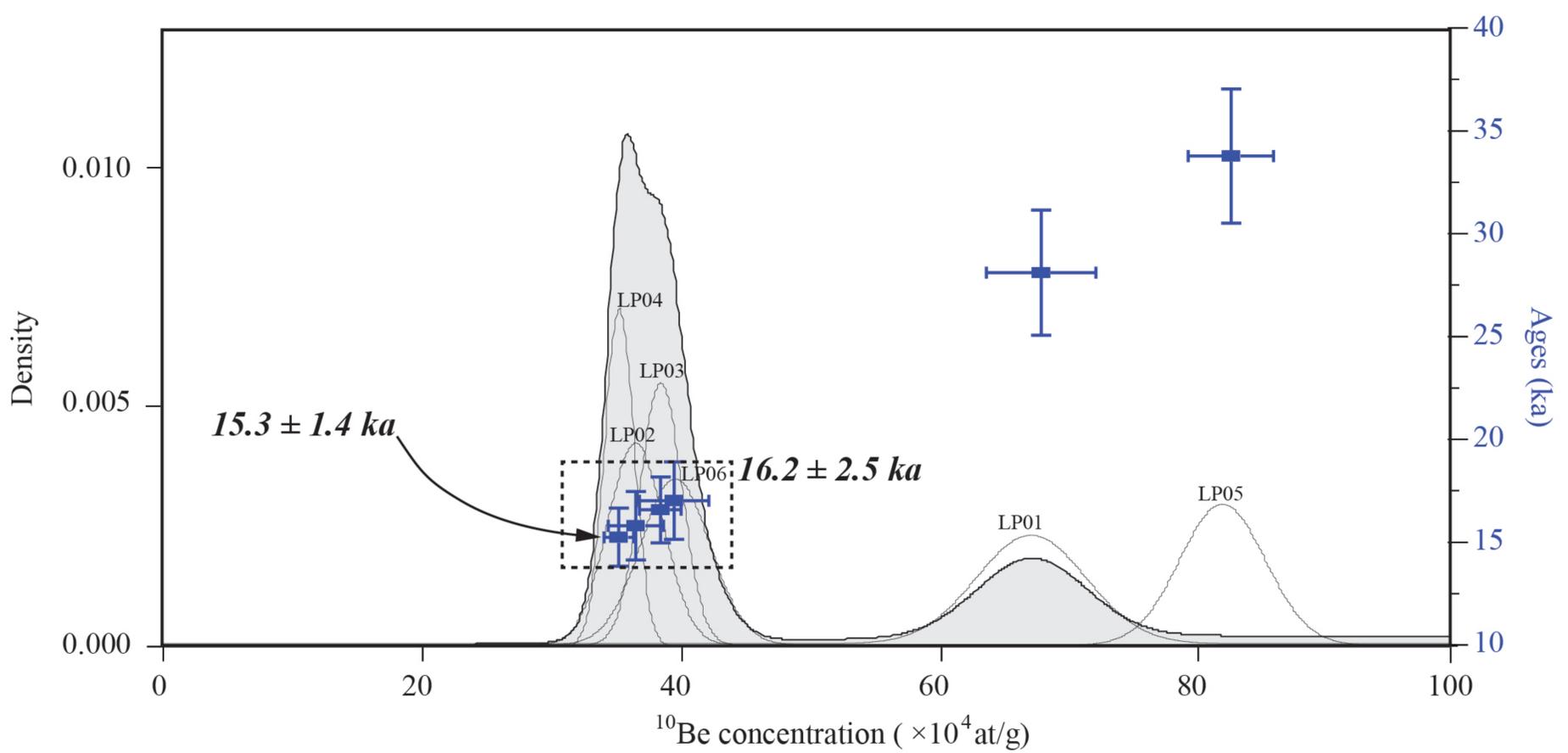
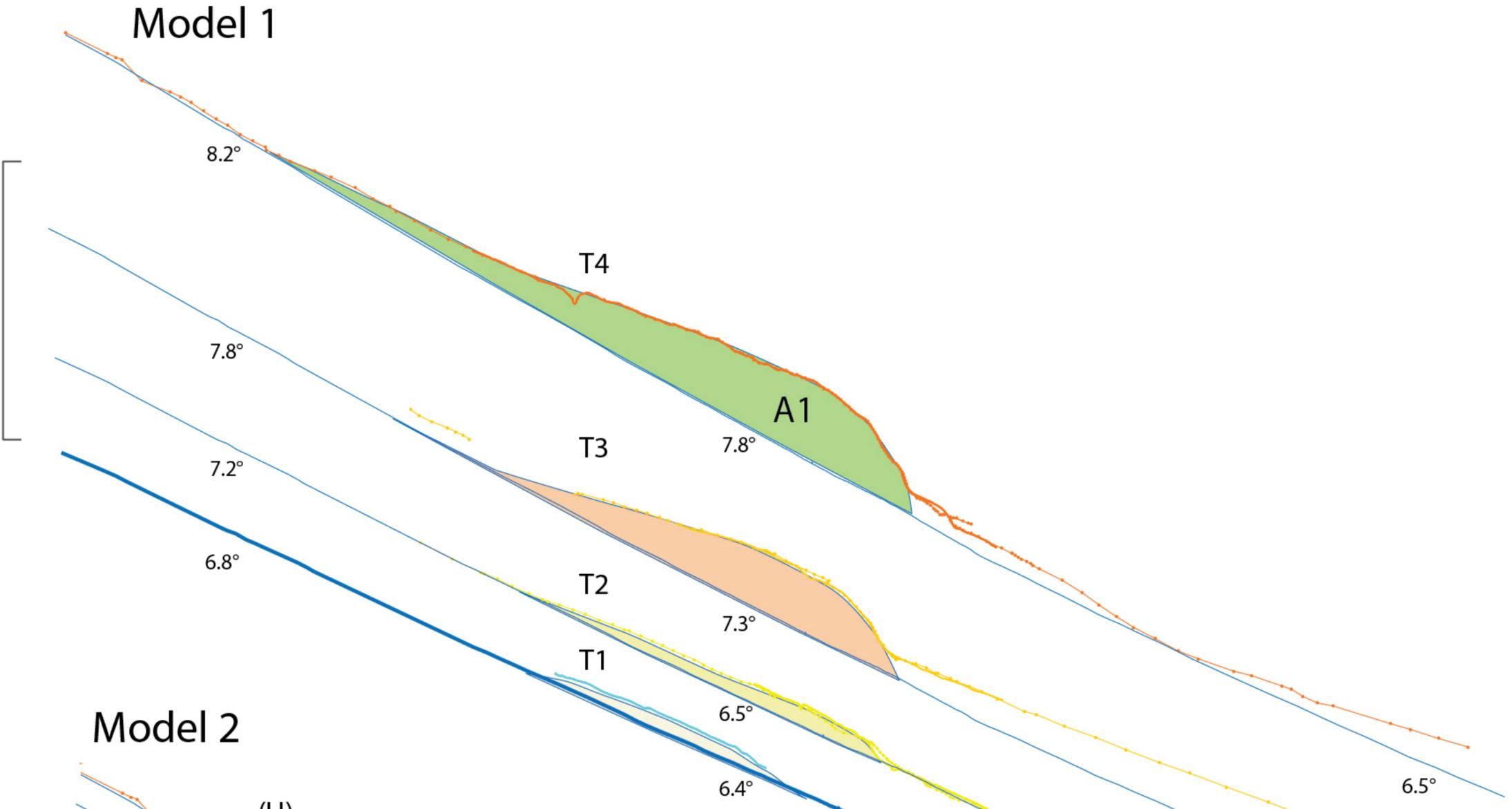


Figure 12.

Model 1

200 m
(x4)



Model 2

200 m
(x4)

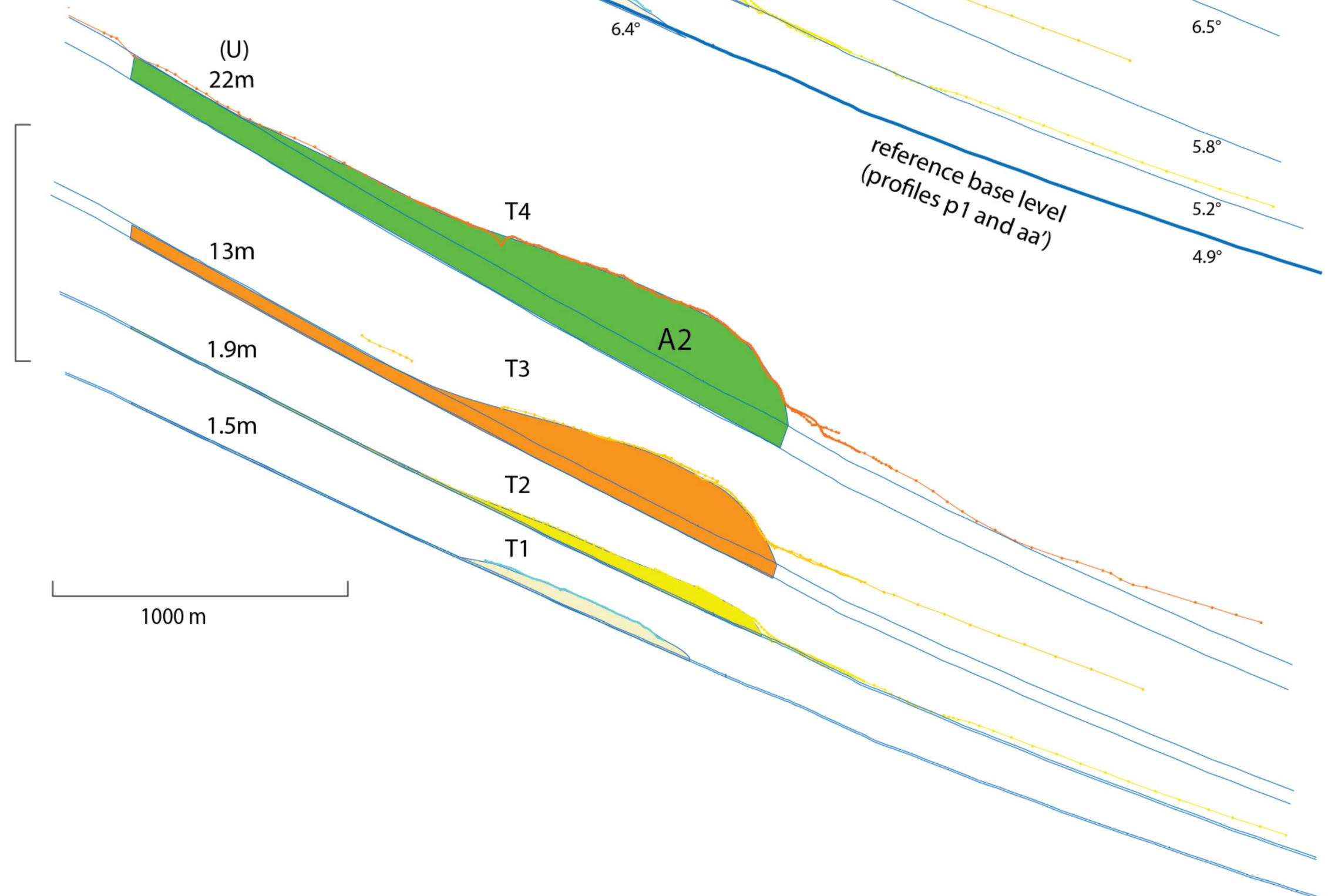
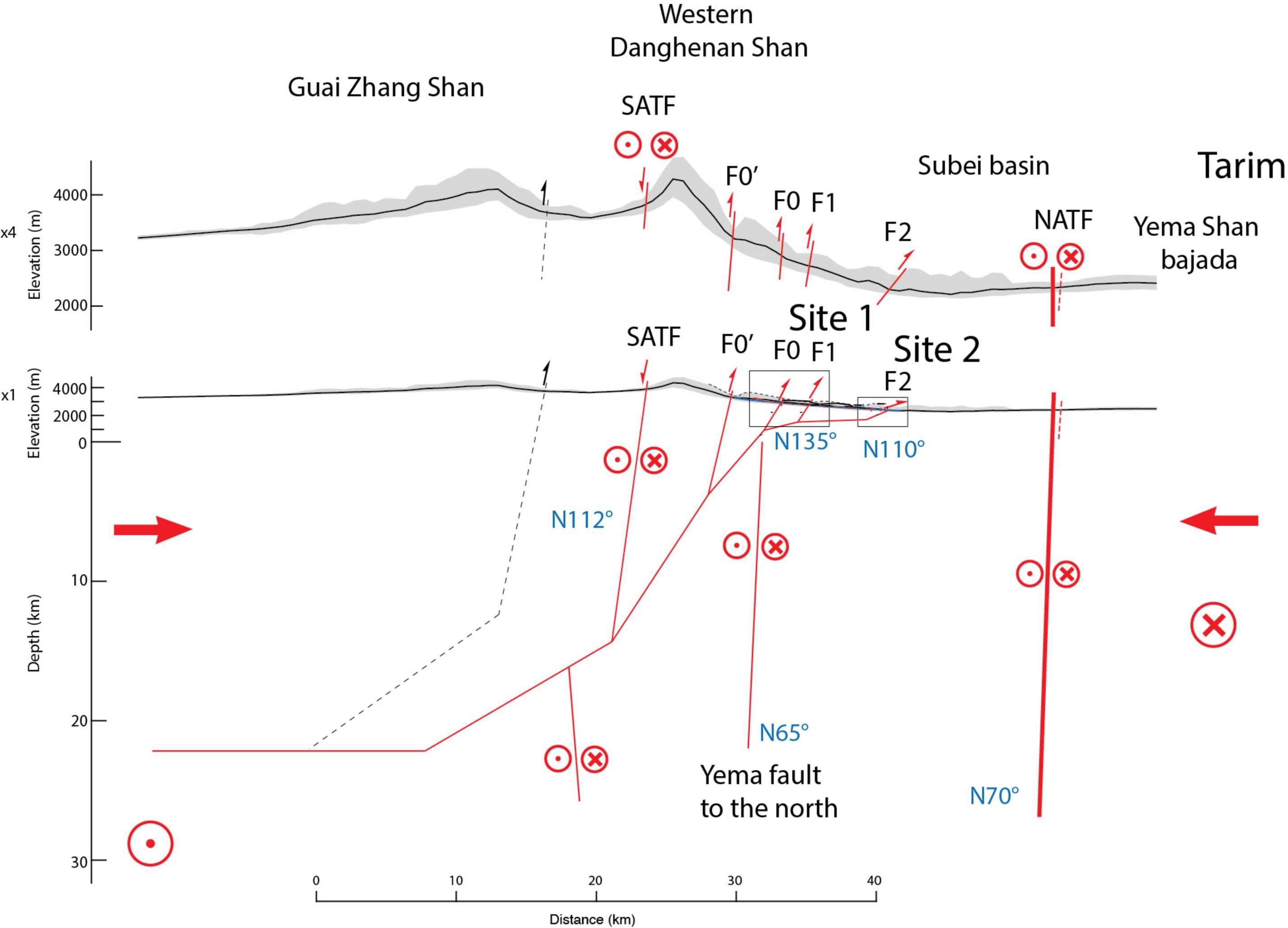


Figure 13.



projection N15°

Figure 14.

