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1 Late Quaternary faulting on the Manas and Hutubi reverse faults in
2 the northern foreland basin of Tian Shan, China

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11
12 **Abstract:**

13 The Tian Shan Range lies in the actively deforming part of the India–Asia collision
14 zone. In the northern foreland basin of Tian Shan, the strata were intensively deformed by
15 Cenozoic folding and faulting. Slip rate studies along these faults are important for
16 understanding the dynamics of crustal deformation and evaluating the seismic hazards in
17 the region. Two reverse faults (the Manas and Hutubi faults) in the northern foreland
18 basin were investigated. Due to past faulting events along these faults, the terrace treads
19 along the Manas River were ruptured, forming fault scarps several meters in height.
20 Loess deposits were trapped and preserved at the surface ruptures along these scarps. The
21 thickness of the trapped loess is dependent on the size of the ruptures. The minimum and
22 maximum ages of these scarps are constrained by dating the loess preserved at the
23 surface ruptures and the terrace treads, respectively, using the quartz optically stimulated
24 luminescence (OSL) dating technique. Our dating results suggest that the loess trapped at
25 the ruptures was deposited from the early to mid-Holocene at the Hutubi Fault, and from
26 the mid- to late-Holocene at the Manas Fault. The vertical displacements of the faults
27 were evaluated by measuring the topographic profiles across the investigated fault scarps
28 using the differential global position system (DGPS). Our results suggest that, during the
29 late Quaternary in the studied region, the vertical slip rates of the Manas Fault were
30 between ~0.74 mm/yr and ~1.6 mm/yr, while the Hutubi Fault had a much lower vertical
31 slip rate between ~0.34 mm/yr and ~0.40 mm/yr. The tectonic implications of our results
32 are discussed.

33

34 Keywords: Tian Shan, Reverse fault, OSL dating, loess, DGPS

35

36 **1. Introduction**

37 In response to the Cenozoic collision of the Indian and Eurasian continental plates,
38 Tian Shan has been one of the most active intra-continental mountain building belts in
39 Central Asia (Fig. 1(a)) (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1979;
40 Avouac et al., 1993; Fu et al., 2003; Sun and Zhang, 2009). The folded Cenozoic deposits
41 in the foreland basins of Tian Shan indicate active tectonics and crustal thickening during
42 the Cenozoic era (Sun and Zhang, 2009). Previous studies suggested that crustal
43 shortening in the Tian Shan region can be up to 200 ± 50 km and that the average altitude
44 of Tian Shan has been uplifted to more than 4000 m (Avouac et al., 1993; Abdрахmatov
45 et al., 1996; Deng et al., 1996).

46 In recent years, cities such as Urumqi in the northern foreland basin of Tian Shan
47 have been developing rapidly. However, the cities are threatened by a number of major
48 active fault systems. Several significant surface-rupturing earthquakes have occurred in
49 the region during historical times, including the 1906 M=7.7 Manas earthquake at Manas
50 town. During the last twenty years, there are reports on paleo-seismicity along the major
51 active reverse faults in the northern foreland basin of Tian Shan (Fig. 1(b)) (e.g. Avouac
52 et al., 1993; Deng et al., 1996). However, many problems remain in deciphering the
53 timing and recurrence intervals of faulting events in the area, because there is a lack of
54 reliable age constrains for fault movement. One way to evaluate seismic hazards in the
55 region is to study the slip rate along the faults. Although different fault systems in the
56 northern foreland basin of Tian Shan have been identified and studied in previous studies,
57 such as the dip of the faults, the strike of the faults as well as the fault displacements (Xu
58 et al., 1992; Avouac et al., 1993; Deng et al., 1996; Zhao et al., 2001), the late Quaternary
59 slip rates along the major fault systems in the studied region are still poorly constrained,
60 mainly due to the lack of reliable chronological data.

61 Optically stimulated luminescence (OSL) dating determines the time since mineral

62 grains were last exposed to sunlight, and hence, it can provide a direct measure of the
63 depositional ages of fault-related sediments. OSL dating is well-suited to date sediments
64 such as loess and sand deposits that were deposited within the last several hundred
65 thousand years (e.g. Aitken, 1998; Wang et al., 2006; Rhodes, 2011; Li et al., 2014). It
66 has been applied to many aeolian-related sedimentary sequences, e.g. sandy deposits
67 from deserts in northeast China and loess deposits from Chinese Loess Plateau (e.g. Li et
68 al., 2002; Sun et al., 2006; Li et al., 2007; Li and Li, 2011; Fu et al., 2012; Li et al., 2012;
69 Gong et al., 2013). OSL dating has also been applied to study the faulting at different
70 fault zones (e.g. Chen et al., 1999; Cheong et al., 2003; Nogueira et al., 2010; Chen et al.,
71 2012, 2013).

72 The timing of past faulting events can be deciphered by dating fault-related colluvial
73 sediment (e.g. Forman et al., 1991; Deng et al., 1996). However, luminescence dating of
74 scarp-derived colluvial deposits is difficult and can be problematic because these
75 materials might be insufficiently bleached prior to deposition, due to short transportation
76 distances. Instead of dating the colluvial deposits, in this study, we propose that the ages
77 of scarps may be constrained by dating the loess trapped at the surface ruptures along the
78 fault scarps and the terrace treads, respectively. A single-aliquot regenerative-dose (SAR)
79 OSL dating technique (Murray and Wintle, 2000) is used. Together with high-resolution
80 differential global position system (DGPS) measurements on the fault scarps, the vertical
81 slip rates of the two reverse faults (the Manas and Hutubi faults) are evaluated and their
82 tectonic implications are discussed.

83

84 **2 .Geological setting**

85 In northwestern China, the Tian Shan Range separates the Tarim Basin to the south
86 from the Junggar Basin to the north (Fig. 1(a)). The Chinese Tian Shan extends east-west
87 more than 1700 km with a north-south width of 250-300 km (Fu et al., 2003). Cenozoic
88 folding and faulting in the Tian Shan Range resulted in strong deformation of the
89 Mesozoic to Cenozoic strata in the foreland basins (Fu et al., 2003; Zhang, 2004;
90 Charreau et al., 2005; 2009; Heermance et al., 2007; Sun and Zhang, 2009; Lu et al.,
91 2010). During the late Cenozoic, three approximately east-west striking, sub-parallel

92 zones of folding (I, II and III in Fig.1(b)) and reverse faulting were formed in the
93 northern foreland basin of Tian Shan (Deng et al., 1996; Burchfiel et al., 1999; Fu et al.,
94 2003).

95 In zone I, the Qingshuihe (QSH) and Qigu (QG) anticlines are composed of
96 Mesozoic to Cenozoic strata (Fig. 1(b)). In zone II, the Huoerguosi (HGS), Manas (MNS)
97 and Tugulu (TGL) anticlines consist of Oligocene to Pleistocene strata. In zone III, the
98 Dushanzi (DSZ) and Anjihai (AJH) anticlines consist of Miocene to Pleistocene strata
99 (Sun and Zhang, 2009). All the three anticlines are characterized by linear and
100 approximately west-east striking axes, indicating a north-south contraction of the range
101 (Xu et al., 1992; Deng et al., 1996). It was reported that zone I of such fold systems was
102 not active during the late Quaternary, while the reverse faults in zone II and III were
103 active during the late Quaternary (Deng et al., 1996). In this study, we focused on two
104 reverse faults (the Hutubi and Manas faults) within the Manas Anticline in zone II (Fig.
105 1(b)).

106 Field-based investigation and satellite observations reveal that the Hutubi Fault
107 extends roughly east to west for more than 60 km. Avouac et al. (1993) investigated the
108 outcrops of the Hutubi Fault at the Taxi and Hutubi rivers, respectively (Fig. 1(b)). It is
109 found that the Hutubi Fault dips 50~60° S at the Taxi River site and 50° S at the Hutubi
110 River site. Therefore, it is very likely that the Hutubi Fault dips at a similar angle near the
111 surface at the Manas River. The Manas Fault is approximately 2 km north of the west
112 side of the Hutubi Fault, striking in a similar direction for more than 30 km (Fig. 1(b)).
113 Zhao et al. (2001) investigated the shallow crustal structure characteristics of the Manas
114 Anticline in the Manas earthquake area using a high-precision shallow seismic
115 prospecting method. It is found that the Manas Fault dips ~45° S near the surface (Zhao
116 et al., 2001).

117 The Manas River originates from Tian Shan and it flows northward to the Junggar
118 Basin (Fig. 1(b)). Both the Manas and Hutubi faults extend across the Manas River,
119 deforming fluvial terraces of the Manas River (Fig. 2(a),(b)). Six fluvial terraces were
120 identified along the east bank of the Manas River within the Manas Anticline. These
121 terraces - from the lowest to the highest - are referred as T-1 to T-6, respectively (Fig.

122 2(b)). The luminescence chronology for the terrace sequence has been established by
123 dating the underlying fluvial sand and the overlying aeolian loess on the terraces (Gong
124 et al., 2014). The ages of the six terrace treads were dated at ~ 0.5 ka, ~1.4 ka, ~3.1 ka,
125 ~4.0 ka, ~12.4 ka and ~19.9 ka, respectively (Table 1) (Gong et al., 2014). For the six
126 terraces, fluvial gravels are overlying Tertiary bedrocks and gravels are mantled by
127 aeolian loess with different thicknesses ranging from around 10 cm to over 10 m. Field
128 investigation suggests that fluvial gravels of terraces T-5 and T-6 along the Manas River
129 were ruptured, forming fault scarps several meters in height (Fig. 2(a),(b)). Loess
130 deposits were trapped and preserved at the surface ruptures along the scarps and the
131 thickness of the loess trapped is variable and depends on the size of the ruptures.

132 The annual precipitation in the study area is less than 300 mm/yr and the precipitation
133 is concentrated during the spring and summer months (Poisson and Avouac, 2004). The
134 prevailing winds of the region are westerlies (Poisson and Avouac, 2004), with the area
135 being mainly controlled by strong near-surface north and northwest winds (Fang et al.,
136 2002; Sun, 2002). Under such winds, aeolian loess entrained from the Junggar Basin is
137 transported and accumulates along the northern piedmont of Tian Shan (Fig. 3).

138

139 **3. Surface ruptures along the reverse fault scarps**

140 To study the faulting along the Hutubi and Manas faults, six exposures for surface
141 ruptures along the reverse fault scarps were investigated. The locations of the exposures
142 and their locations in relation to the two reverse faults are shown in Fig. 4(a),(b),(c). The
143 exposures are termed as HTB0, HTB1 and MNS1-MNS4, respectively.

144 HTB0 (N 44° 10' 5.1", E 86° 08' 11.9") is located at the western part of the Hutubi
145 Fault (Fig. 1(b)). It belongs to the outcrop of the Hutubi Fault at T-6 on the east bank of
146 the Manas River (Fig. 5(a)). The gravel layer of T-6 was displaced vertically by the fault
147 for about 7.8 m (Fig. 5(a)). Due to the development of a gully along the fault, the original
148 scarp was destroyed so we could not find associated trapped loess on the fault scarp.

149 HTB1 (N 44° 10' 11.1", E 86° 07' 37.7") is the surface rupture of the terrace tread of
150 T-5 on the east bank of the Manas River (Fig. 4(b)). It belongs to the outcrop of the
151 Hutubi Fault at the west side. It is a part of the fault scarp, where the aeolian loess

152 deposits were trapped and preserved (Fig. 4(c)). The loess trapped at the rupture along the
153 fault scarp is about 1.8 m in thickness (Fig. 5(b)).

154 MNS1 (N 44° 11' 17.6", E 86° 08' 16.8") is the surface rupture of the terrace tread of
155 T-5 on the east bank of the Manas River (Fig. 4(b)) and it belongs to the outcrop of the
156 Manas Fault (Fig. 5(c)). MNS1 is at approximately 2 km north of HTB1. The exposure
157 is only part of the scarp of the Manas Fault, which was previously studied by Avouac et al.
158 (1993) and Deng et al. (1996). The strata in the hanging wall of the scarp consist of
159 Tertiary gray, gray-green and reddish deformed mudstones and the late Pleistocene
160 gravels (Xu et al, 1992). At the foot wall, only the late Pleistocene gravels were exposed.
161 MNS1 shows complex slope morphology and the trapped loess at the surface rupture
162 along the scarp is about 1 m in thickness (Fig. 5(c)).

163 MNS2 (N 44° 11' 18.7", E 86° 07' 58.3") (Fig. 5(d)) is at about ~300 m west of
164 MNS1 (Fig. 4(b)). The exposure also shows the surface rupture of the terrace tread of T-5
165 on the east bank of the Manas River, and it belongs to the outcrop of the Manas Fault
166 (Fig. 5(d)). The trapped loess at the rupture along the fault scarp is about 1 m in
167 thickness.

168 MNS3 (N 44° 11' 27.5", E 86° 08' 53.18") is shown in Fig. 5(e). It is parallel to
169 exposures MNS1 and MNS2. This exposure is the surface rupture of the terrace tread of
170 T-5 on the west bank of the Manas River, and it belongs to the outcrop of the Manas Fault.
171 The trapped loess at the rupture along the fault scarp is about 1 m in thickness.

172 MNS4 (N 44° 11' 27.7", E 86° 06' 51.8") is shown in Fig. 5(f). It is the surface
173 rupture of the terrace tread of T-5 on the west bank of the Manas River. MNS4 is about
174 100 m west of MNS3. MNS4 is a part of the scarp of the Manas Fault, where loess
175 deposits were trapped and preserved. The loess trapped at the rupture along the scarp is
176 about 1 m in thickness.

177 The scarps of the Manas and Hutubi faults extend more than tens of meters so it is very
178 difficult to dig a trench to fully reveal the strata along these scarps. From the above
179 exposures (HTB1, MNS1-MNS4), it is found that the hanging and foot walls along the
180 fault scarp are composed of fluvial gravels, which are rather weakly consolidated. The
181 fault-scarp-derived colluvial gravels have been mixed well with the hanging and foot walls

182 at the ruptures and their boundary are obscured. Thus, it is very difficult to distinguish
183 individual faulting events by lithofacies analysis of colluvial sediments at these ruptures.
184 Alternatively, trapped loess was commonly preserved at the ruptures along the fault scarps.
185 We think the trapped loess at the surface ruptures along the fault scarps may provide
186 important chronological information linked to past faulting events.

187 We follow Avouac et al.'s. (1993) method to estimate the fault displacements for both
188 reverse faults, i.e. the vertical displacements of the faults were evaluated by measuring the
189 topographic profiles across the scarps using the differential global position system (Real
190 Time Kinematic) (Fig. 4(b)). The terrace treads of T-5 and T-6 were used as the offset
191 markers for the faulting along the Hutubi and Manas faults and the elevation data were
192 recorded by DGPS measurements along the cross-sections B-B', C-C' and D-D'(Fig. 4(b)).
193 For our field measurements, the topographic profiles B-B', C-C' and D-D' were measured
194 long enough to cross the faulted terrace treads at either side of the scarps until the surface
195 was nearly planar and there was no significant degradation. In addition, it was found that
196 the regional slope of the fluvial terraces was very small (typically at 1-2°) at either side of
197 scarps studied (estimated from Google Earth). Thus, the measured vertical offset is nearly
198 equal to the vertical throw on the fault, independent of fault dip, as suggested by Avouac et
199 al. (1993). The vertical displacement by the fault can be estimated by the difference in the
200 elevations between the faulted terrace treads at either side of the scarps. We assigned a 10 %
201 error for the fault displacements for both reverse faults. The vertical displacements along
202 the Manas and Hutubi faults were shown in Fig. 4(b) and Fig. 5(a). It was found that the
203 Hutubi Fault has vertically displaced terrace T-5 by 4.2 ± 0.4 m (cross-section B-B', Fig.
204 4(b)) and terrace T-6 by 7.8 ± 0.8 m (cross-section C-C', Fig. 5(a)). The Manas Fault has
205 vertically displaced T-5 by 9.2 ± 0.9 m (cross-section D-D', Fig. 4(b)). The cross-section
206 (D-D') is at the middle site between MNS1 and MNS2. Our displacement measurement
207 results are consistent with those from Avouac et al. (1993).

208

209 **4. The relationship between loess at the ruptures and faulting along the faults**

210 To explain why the ages of the loess deposits preserved at the surface ruptures can be
211 used to constrain the minimum ages of scarps, a schematic model is provided in Fig. 6 to

212 illustrate the evolution of ruptures along the fault scarp on the terrace and the infilling of
213 loess deposits along the ruptures.

214 The river terrace was formed at time (t_0) when the fluvial gravels were deposited on
215 the bedrock (Fig. 6, stage 1). The terrace tread can be used as the offset marker for the
216 faulting along the reverse fault. When a faulting event occurred (referred to as event 1) at
217 time t_1 , the terrace tread was deformed and a fresh rupture was created. As the gravels
218 from the hanging wall are weakly consolidated, they soon collapse and fill in the ruptures,
219 forming a colluvial unit (C) (Fig. 6, stage 2). The rupture then started to trap aeolian loess
220 deposits (L1) until a balance between deposition and erosion (Fig. 6, stage 3) was
221 achieved. Assuming that the loess deposited soon after the surface rupture was formed,
222 the age (t_1') of the loess layer 1 (L1) is close to the time of faulting event 1 (t_1). A similar
223 process has been documented by the reverse fault induced San Fernando earthquake of
224 1971 (Kahle, 1975). In that case, the hanging-wall was weakly consolidated Tertiary
225 sandstone and conglomerate. More than 50% of the 0.3 m to 1 m high overhanging free
226 face collapsed after three months.

227 When a second faulting event (event 2) occurred along the reverse fault at time t_2 , the
228 hanging wall was further displaced above the foot wall, leading to the rupture growth
229 (Fig. 6, stage 4). At this time, some surface soil (S) or loess at the rupture might be
230 reworked and re-deposited on the foot wall block. Then, the reworked soil or loess might
231 be eroded away, or mixed with new aeolian loess deposits (L2) to be trapped in the
232 rupture (Fig. 6, stage 5) until a new balance reached (Fig. 6, stage 6). Such faulting and
233 loess infilling in the rupture along the scarp was repeated, which created different loess
234 layers. Thus, the oldest loess deposits trapped in the rupture can be used to constrain the
235 minimum ages of scarps.

236 It should be noted that, an important assumption must be made for using the trapped
237 loess deposits at the surface ruptures along the fault scarps to infer the minimum ages —
238 that is there was a continuous loess deposition during the fault movement period. This
239 assumption is expected to be valid for this study area. The Junggar Desert is located at the
240 north of Tian Shan, occupying an area of 48,800 km² (Sun, 2002). Such a desert serves as
241 a huge and constant holding area for aeolian dust. Besides, the study area is mainly

242 controlled by strong near-surface north-west and north winds (Sun, 2002). Under such
243 winds, aeolian loess entrained from the Junggar Basin is transported and accumulates in
244 the northern piedmont of Tian Shan - the upper limit of loess depositions can be at 2400
245 m above sea level (Fig. 3). Gong et al. (2014) demonstrated that there were continuous
246 loess depositions on the terraces from ~20 ka to ~0.5 ka in the same study area. In other
247 mountain belts of Asia, such as the Qilian Shan, it is also demonstrated that there was
248 continuous loess deposition along the mountain front during the Holocene (e.g. Stokes et
249 al., 2003; Küster et al., 2006). Thus, the aeolian dusts transported from the Junggar
250 Desert to the northern piedmont of Tianshan (Sun, 2002) were very likely to be trapped at
251 the surface ruptures and the loess deposits offer the possibility to constrain minimum
252 ages of the scarps along the Hutubi and Manas faults in the study area.

253 Based on the model described above, multiple samples were collected from each of
254 the scarps. The detailed sampling name, material and positions for the loess deposits at
255 the ruptures along the fault scarps are shown in Table 2 and Fig. 5.

256

257 **5. Luminescence dating**

258 **5.1 Experimental procedure**

259 The OSL samples were collected by hammering stainless steel tube horizontally into
260 freshly cut vertical sections. These tubes were then covered with a lid soon after taking
261 them from the section, then sealed in a black plastic bag. The OSL samples were analysed
262 in the Luminescence Dating Laboratory at the Department of Earth Science of the
263 University of Hong Kong. Two centimeters of material at each end of the tube was
264 scrapped away and used for dose rate measurements. Raw samples in the centre of the
265 tubes were first treated with 10% hydrochloric acid (HCl) and 10% hydrogen peroxide
266 (H₂O₂) to remove carbonates and organic materials, respectively. Coarse grains between
267 90 and 125µm were selected by mechanical dry sieving. Quartz grains were separated
268 using sodium polytungstate heavy liquid with density between 2.62-2.75 g/cm³. The
269 quartz grains were etched with 40% HF for ~80 minutes to remove the outer alpha dosed
270 layer as well as any remaining feldspar. HCl (10%) was used again to dissolve any
271 residual fluorides after etching before final rinsing and drying. The etched grains were

272 mounted as a monolayer on 9.8 mm wide aluminum discs using silicone oil as an adhesive.
273 Grains covered the central ~3 mm diameter portion of each disc, corresponding to several
274 hundreds of grains per aliquot. The purity of the quartz grains was tested by monitoring
275 the presence of feldspar by measuring the infrared stimulated luminescence (IRSL) and
276 110 °C thermoluminescence (TL) peak (Li et al., 2002). At least twenty aliquots were
277 measured for each sample.

278 OS� signals were measured using automated Risø TL/OSL readers equipped with
279 excitation units of blue light emitting diodes (LEDs) ($\Delta=470\pm30$ nm) and infrared LEDs
280 ($\Delta=870 \pm 40$ nm). The blue LEDs deliver ~50 mW/cm² of light to the sample
281 (Bøtter-Jensen et al., 2003). Ninety percent of their full powers were used for stimulation
282 in this study. The quartz OS� signals were detected through 7.5 mm Hoya U-340 filters,
283 which allow a transmission from 290 nm to 370 nm with a peak at ~340 nm (Aitken,
284 1998). Irradiation was carried out using ⁹⁰Sr/⁹⁰Y beta sources built into the readers.

285 The equivalent doses (D_e) of quartz were determined using a SAR OS� dating
286 protocol (Wintle and Murray, 2000). In the protocol, preheating condition was
287 determined from a preheat plateau test on the sample MNS2-4, using preheat
288 temperatures from 200 °C to 300 °C in increments of 20 °C (Fig. 7). A D_e plateau was
289 observed between 220 °C and 260 °C. Based on the preheat plateau results, a preheating
290 at 220 °C for 10 s and a cut-heat at 180 °C were selected before the regenerative dose and
291 test dose OS� measurements, respectively. Additionally, a zero dose was used for
292 monitoring recuperation effects and a repeated dose the same as the first regeneration
293 dose was used for checking the reproducibility of the sensitivity correction (i.e. recycling
294 ratio). Aliquots with recuperation higher than 5% or recycling ratio falling out of the
295 range of 1.0 ± 0.1 were discarded for age calculation.

296

297 **5.2 Dose rate measurements**

298 The environmental dose rate was determined using a variety of techniques. The
299 Thick-source alpha counting technique (Aitken, 1998) was used to measure the dose
300 contributions from the uranium (U) and thorium (Th) decay chains. A Littlemore 7286

301 TSAC system with 42-mm-diameter ZnS screens was used. Potassium content was
302 measured by X-ray fluorescence (XRF). Water content was calculated as the ratio of water
303 weight to dried sample weight, obtained from the sample weights before and after drying
304 at 105 °C in an oven. As the water content may have changed through time, a 20% relative
305 standard error was assigned for the measured water content to take account of the
306 long-term variation in water content (refer to Table 2). The cosmic ray contribution to the
307 dose rate was calculated from the burial depth, altitude, latitude and longitude of the
308 samples (Prescott and Hutton, 1994). The dosimetry data were summarized in Table 2.

309

310 **5.3 OSL dating results**

311 Typical OSL shine down curves and the regenerative growth curves of quartz grains
312 from the loess samples are shown in Figure 8. It shows that the OSL contains a strong
313 fast component, with the OSL emission decreasing by ~80-90% during the first 4 s of
314 stimulation. The D_e distribution from each sample was analyzed using radial plot, which
315 has been widely used to show the distributions of single-grain or single-aliquot D_e
316 estimations (Galbraith et al., 1999; Olley et al., 1999; Jacobs et al., 2003; Zhang et al.,
317 2009). Fig. 9 displays examples of radial plots for six loess samples from the five
318 exposures (HTB1 and MNS1-MNS4). In the plots, the central lines of the shaded regions
319 with 2σ width of the D_e distribution represent the mean D_e values of the measured
320 aliquots. Based on the radial plots, there is no evidence of variable bleaching for these
321 loess samples, i.e. most of the D_e values are symmetrically distributed around the central
322 values. We conclude that our OSL ages are reliable based on the following aspects: 1) the
323 loess deposits are aeolian in origin and are therefore unlikely to suffer from insufficient
324 bleaching problems in OSL dating - this is supported by the D_e distribution of our
325 samples (Fig. 9). 2) our samples yield good results in the SAR performance tests, in
326 terms of pre-heat plateau, recycling ratio and recuperation. 3) most of the samples yield
327 over-dispersion (OD) values between 10% and 35% (Table 3). Although the effect of
328 inhomogeneous gamma radiation might make some contributions to the scatter of D_e ,
329 Gong et al. (2014) demonstrated that the uncertainty from spatially heterogeneous dose
330 rates to the final age estimation is less than 5% for the OSL samples in the same study

331 area.

332 Table 3 summarizes the D_e estimates and the OSL ages of all the samples based on
333 mean age and the central age models, respectively. It is found that the two age models
334 produce consistent ages within 2σ for all the samples. The mean age model was chosen for
335 final determination of the OSL ages. Fig. 10 shows the five exposures of the fault scarps
336 together with the OSL dating results of the loess samples. The trapped loess preserved at
337 HTB1 produced OSL ages from ~ 10.5 ka to ~ 4.2 ka. It is found that the inner trapped
338 loess produced older ages than the outer trapped loess. In comparison, the trapped loess
339 preserved along the Manas Fault (MNS1-MNS4) produced younger OSL ages, ranging
340 from ~ 5.7 ka to ~ 1.7 ka (Fig. 10). A possible explanation for the multiple OSL ages of the
341 loess deposits at the ruptures along the scarps is episodic faulting-induced rupture growth
342 and subsequent loess sedimentation, as suggested by our model (Fig. 6). Therefore, the
343 chronology data of the trapped loess along HTB1 suggests that the Hutubi Fault was active
344 from the early to mid-Holocene, while the Manas Fault was active from the mid- to late
345 Holocene. However, the different OSL ages of the loess at ruptures along the scarps can be
346 also explained in other ways, e.g. colluvial re-deposition, disturbance of the loess after
347 multiple rupture events or continuous (discontinuous) loess deposition after a single
348 rupture event. Nevertheless, in any case, the ages of the terrace treads and the oldest loess
349 preserved at ruptures along the fault scarps should constrain the maximum and minimum
350 ages of the scarps studied, respectively.

351

352 **6. Discussion**

353 **Slip rates of the Hutubi and Manas faults**

354 Based on the topographic profiles measured across the investigated fault scarps and the
355 chronology data, we calculated vertical slip rates for both reverse faults (Table 4). The
356 Hutubi Fault has vertically displaced terrace T-5 by 4.2 ± 0.4 m (cross-section B-B'). As
357 the oldest loess sample preserved in exposure HTB1 and age of the terrace tread of T-5
358 were constrained at 10.5 ± 1.0 and 12.4 ± 0.8 ka (Gong et al., 2014), respectively, the
359 maximum and minimum vertical slip rates of the Hutubi Fault within the Manas Anticline
360 were calculated at 0.40 ± 0.05 and 0.34 ± 0.04 mm/yr, respectively, during the late

361 Quaternary. Similarly, the Hutubi Fault has vertically displaced terrace T-6 by 7.8 ± 0.8 m
362 (cross-section C-C') (Fig. 4(b)). The age of the tread of T-6 is 19.9 ± 1.5 ka (Gong et al.,
363 2014), which constrains the maximum age of HTB0. The offset and age of the T-6 tread
364 yields a minimum vertical slip rate of 0.39 ± 0.05 mm/yr (Table 3). The two sets of data
365 from T-5 and T-6 suggest that the vertical slip rate of the Hutubi Fault within the Manas
366 Anticline was at about 0.4 mm/yr during last ~ 12 ka.

367 T-5 was vertically displaced by 9.2 ± 0.9 m by the Manas Fault, as shown in
368 cross-section D-D' in Fig. 4(b). As the oldest loess sample preserved along exposures of
369 MNS1-MNS4 and the age of terrace tread T-5 were constrained at 5.7 ± 0.6 and $12.4 \pm$
370 0.8 ka, respectively (Gong et al., 2014), the maximum and minimum vertical slip rates of
371 the Manas Fault are 1.6 ± 0.2 mm/yr and 0.74 ± 0.09 mm/yr, respectively. If it is assumed
372 that the loess was deposited and preserved at the surface ruptures soon after the faulting
373 events, the vertical slip rate of the Manas Fault is close to 1.6 ± 0.2 mm/yr during the late
374 Holocene. It is to be noted that the evaluation of the vertical slip rates of the two reverse
375 faults also depends on the observation timescale. Due to the intermittency of faulting
376 events along the Manas and Hutubi faults, measurements of fault slip rates over
377 timescales shorter than the longest gap will overestimate long-term average fault slip
378 rates. For instance, an earthquake in the near future with an offset of 1-2 m would
379 increase the cumulative offset of 4.2 m (fault displacement of HTB1) to 5.2-6.2 m, with a
380 similar rise in the slip rate. To better compare the faulting along the Manas and Hutubi
381 faults, their vertical slip rates were compared over the same timescale: i.e. from the
382 formation the tread of T-5 (~ 12.4 ka) to the present. It is found that the correspondingly
383 vertical slip rates of the Manas and Hutubi faults are calculated at 0.74 ± 0.09 mm/yr and
384 0.34 ± 0.04 mm/yr, respectively. In addition, it is also interesting to calculate the
385 horizontal shortening rate and the fault slip rate of the two reverse faults. Although the
386 strata along the scarps were not fully revealed by our trenches, the dip angles of the two
387 reverse faults at the surface can be estimated based on the results from previous studies
388 (Avouac et al., 1993; Zhao et al., 2001). In this study, a dip angle of $45 \pm 5^\circ$ of the Manas
389 Fault (Zhao et al., 2001) is used to calculate the slip rates of the Manas Fault. The
390 horizontal shortening rate and the total slip rate of the Manas Fault are calculated at 0.74

391 ± 0.17 mm/yr and 1.05 ± 0.16 mm/yr, respectively. For the Hutubi Fault, a dip angle of
392 $55 \pm 5^\circ$ (Avouac et al., 1993) is used to calculate the slip rates, and the horizontal
393 shortening rate and the total slip rate of the Hutubi Fault are calculated at 0.24 ± 0.06
394 mm/yr and 0.42 ± 0.06 mm/yr, respectively. The results show that the Manas Fault
395 exhibits a higher slip rate than the Hutubi Fault at the site during the Holocene.

396 It is also interesting to note that the oldest OSL age of the loess trapped at the
397 surface ruptures along the Manas Fault (MNS1-MNS4) is ~ 5.7 ka, while significantly
398 older loess deposits were found at HTB1. One of the possible explanations is that the
399 Manas Fault was relatively inactive during the early to mid-Holocene and the faulting at
400 the studied area mainly occurred at the Hutubi Fault during that period. From the
401 mid-Holocene to the present, tectonic deformation probably accelerated in the northern
402 foreland basin, not only making the Manas Fault active, but also leading to more
403 significantly surface rupturing events along the Manas Fault. In addition, the loess of ~ 4
404 ka was identified at the surface ruptures along both the Hutubi and Manas faults,
405 indicating that both the Hutubi Fault and the Manas Fault were ruptured in a large,
406 surface-rupturing earthquakes at ~ 4 ka. An alternative explanation for our results is that
407 the Manas Fault was also active during the early to mid-Holocene, but the faulting events
408 along the Manas Fault older than 6 ka were not recorded due to the lack of terrace
409 preservation. Thus, to further examine whether there is accelerated tectonic deformation
410 in the northern foreland basin of Tian Shan, more studies on different fault segments in
411 the area should be carried out in the future, in order to obtain more data on the faulting at
412 different timescale during the Quaternary.

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415 **7. Conclusions**

416 In the northern foreland basin of Tian Shan, the fluvial terraces along the Manas
417 River were ruptured by the Manas and Hutubi faults, forming fault scarps several meters
418 in height. OSL dating results for the trapped loess preserved at the surface ruptures along
419 the fault scarps show that the Hutubi Fault was active from the early to mid-Holocene,
420 while the Manas Fault was active from the mid- to late Holocene. Together with

421 high-resolution differential global position system (GDPS) measurements on the fault
422 scarps, the corresponding vertical slip rates of the Manas and Hutubi faults were
423 calculated and compared. During the late Quaternary, the vertical slip rates of the Manas
424 Fault are constrained to between ~0.74 mm/yr and ~1.6 mm/yr, whilst those of the
425 Hutubi Fault are constrained to between ~0.34 mm/yr and ~0.40 mm/yr in the studied
426 region. The tectonic uplift activity in the northern foreland basin of Tian Shan likely
427 accelerated during the late Quaternary.

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592 Table 1: the abandonment ages of the six terraces along the Manas River in the study area.

593 For the detail of the ages of the terraces, please refer to Gong et al. (2014).

Terraces	T-1	T-2	T-3	T-4	T-5	T-6
Optical ages, ka	0.5±0.1	1.4±0.3	3.1±0.3	4.0±0.4	12.4±0.8	19.9±1.5

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618 Table 2: the dose rate results for loess samples from the five exposures of the fault scarps

619 (HTB1, MNS1-MNS4) in the study area.

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Sample	depth (m)	Alpha counting rate ^a	K content(%)	Water content ^b (%)	Cosmic ray ^c (Gy/ka)	Dose rate (Gy/ka)
HububiFault						
HTB1-1	0.3	10.9±0.2	2.27±0.23	6.0	0.23	3.85±0.16
HTB1-2	0.4	9.53±0.19	1.86±0.19	4.6	0.22	3.32±0.16
HTB1-3	0.5	10.7±0.21	2.10±0.21	3.2	0.22	3.77±0.16
HTB1-4	1.2	11.5±0.21	2.09±0.21	3.3	0.20	3.85±0.16
HTB1-5	0.8	10.9±0.21	2.30±0.23	2.0	0.21	3.86±0.16
HTB1-6	1.6	10.8±0.21	2.10±0.21	6.7	0.19	3.63±0.16
HTB1-7	1.0	10.7±0.20	2.20±0.22	8.7	0.21	3.60±0.16
ManasFault						
MNS1						
MNS1-1	0.3	10.5±0.21	1.86±0.19	5.6	0.22	3.39±0.14
MNS1-2	0.7	10.4±0.20	2.01±0.20	7.6	0.21	3.42±0.15
MNS1-3	0.5	10.1±0.20	2.00±0.20	7.1	0.22	3.40±0.15
MNS1-4	0.5	10.6±0.21	2.00±0.20	8.8	0.22	3.40±0.16
MNS1-5	1.0	10.4±0.21	2.12±0.21	9.6	0.20	3.48±0.16
MNS1-6	1.0	9.55±0.19	2.20±0.22	8.7	0.20	3.47±0.16
MNS1-7	0.8	9.34±0.19	2.05±0.21	8.5	0.21	3.32±0.16
MNS2						
MNS2-1	0.5	10.2±0.2	2.30±0.23	3.4	0.22	3.82±0.17
MNS2-2	1.0	10.1±0.2	2.18±0.22	3.0	0.20	3.70±0.16
MNS2-3	1.6	11.0±0.2	2.14±0.21	6.3	0.19	3.64±0.16
MNS2-4	2.0	11.1±0.2	2.21±0.22	9.2	0.18	3.59±0.16
MNS3						
MNS3-1	0.5	10.7±0.2	2.18±0.22	3.7	0.22	3.83±0.17
MNS3-2	0.5	10.3±0.2	2.17±0.22	2.0	0.22	3.82±0.17
MNS3-3	1.0	11.0±0.2	2.39±0.24	2.1	0.20	4.11±0.19
MNS3-4	1.0	9.77±0.22	1.93±0.19	1.0	0.20	3.54±0.15
MNS3-5	1.0	10.5±0.2	2.31±0.23	1.0	0.20	4.01±0.18
MNS4						
MNS4-1	0.5	9.57±0.20	2.13±0.21	10.2	0.22	3.38±0.16
MNS4-2	0.8	9.56±0.20	2.11±0.21	11.0	0.21	3.32±0.16
MNS4-3	1.0	10.1±0.2	2.24±0.22	7.7	0.20	3.62±0.17
MNS4-4	0.5	10.6±0.2	2.20±0.22	10.1	0.22	3.58±0.17
MNS4-5	1.0	9.71±0.20	2.30±0.23	8.4	0.20	3.59±0.17

621 ^aThe alpha counting rate is for a 42-mm-diameter ZnS screen and is given in units of counts per kilo second.622 ^bThe error for the water content is estimated at ±20 %.

623^c The error for the cosmic rays dose rate is estimated at ± 0.02 Gy/ka.

624 Table 3: quartz OSL ages for the loess samples from the five exposures of the fault scarps

625 (HTB1, MNS1-MNS4) in the study area.

626

Sample	aliquots	Over-dispersion values (%)	Mean Age model De (Gy)	Mean Age model Age (ka)	Central Age model D _e (Gy)	Central Age model Age (ka)
HububiFault						
HTB1-1	20	26.5±2.2	16.2±1.4	4.2±0.4	15.7±1.4	4.1±0.4
HTB1-2	20	28.5±1.7	29.1±2.7	8.8±0.9	27.5±2.7	8.3±0.9
HTB1-3	26	39.3±1.2	32.5±3.4	8.6±1.0	30.0±2.4	7.9±0.7
HTB1-4	20	31.5±1.4	40.3±3.6	10.5±1.0	38.4±3.0	10.0±0.9
HTB1-5	20	29.8±1.1	16.1±1.2	4.2±0.3	15.4±1.1	4.0±0.3
HTB1-6	20	24.2±0.9	23.3±1.4	6.4±0.5	22.6±1.2	6.2±0.4
HTB1-7	20	20.5±0.9	15.3±1.2	4.3±0.4	15.5±0.8	4.3±0.3
ManasFault						
MNS1						
MNS1-1	20	33.5±2.3	8.1±1.7	2.4±0.5	6.8±1.1	2.0±0.3
MNS1-2	20	30.5±1.4	19.4±1.8	5.7±0.6	18.9±1.5	5.5±0.5
MNS1-3	20	27.6±1.6	8.1±0.7	2.4±0.2	7.9±0.6	2.3±0.2
MNS1-4	20	30.0±1.4	10.4±0.8	3.1±0.3	10.3±0.8	3.0±0.3
MNS1-5	20	24.2±0.9	14.5±0.8	4.2±0.3	14.1±0.8	4.0±0.3
MNS1-6	20	34.8±1.6	17.8±1.7	5.1±0.5	16.6±1.5	4.8±0.5
MNS1-7	20	24.2±1.9	13.7±1.1	4.1±0.4	13.3±1.1	4.0±0.4
MNS2						
MNS2-1	20	22.1±1.6	6.5±0.5	1.7±0.1	7.1±0.5	1.9±0.2
MNS2-2	20	16.8±1.0	9.5±0.6	2.6±0.2	9.6±0.4	2.6±0.2
MNS2-3	20	32.9±1.6	11.3±0.9	3.1±0.3	10.5±1.0	2.9±0.3
MNS2-4	20	27.6±1.3	11.9±0.8	3.3±0.3	11.4±0.8	3.2±0.3
MNS3						
MNS3-1	20	34.1±1.7	15.4±1.5	4.0±0.4	14.1±1.5	3.7±0.4
MNS3-2	20	39.8±1.6	15.4±1.9	4.0±0.5	14.4±1.4	3.8±0.4
MNS3-3	20	13.3±0.7	14.0±0.4	3.4±0.2	14.1±0.5	3.4±0.2
MNS3-4	20	33.6±1.6	19.2±1.7	5.4±0.5	17.9±1.6	5.1±0.5
MNS3-5	21	25.8±1.0	19.1±1.3	4.7±0.4	18.4±1.1	4.6±0.3
MNS4						
MNS4-1	20	35.0±1.5	7.7±0.8	2.3±0.3	7.3±0.6	2.2±0.2
MNS4-2	20	39.5±1.5	14.0±1.4	4.2±0.5	13.0±1.2	3.9±0.4
MNS4-3	20	26.4±1.1	15.7±1.1	4.3±0.4	15.1±1.0	4.2±0.3
MNS4-4	20	27.6±1.1	7.5±0.6	2.1±0.2	7.4±0.5	2.1±0.2
MNS4-5	20	34.0±1.4	19.1±1.9	5.3±0.6	18.1±1.5	5.0±0.5

627

628

629

630 Table 4: Estimations of the late Quaternary vertical slip rates of the Hutubi and Manas
631 faults in the northern piedmont of Tian Shan.

632

633

	Deformed terrace	Cumulative vertical offset (m)	The onset time of deformation (ka)		Vertical slip rate (mm/yr)	
			minimum	maximum	minimum	maximum
Hutubi Fault	T-5	4.2±0.4	10.5 ± 1.0	12.4 ± 0.8	0.34 ± 0.04	0.40 ± 0.05
				20 ± 1.5	0.39 ± 0.05	
Manas Fault	T-5	9.2 ± 0.9	5.7 ± 0.6	12.4 ± 0.8	0.74 ± 0.09	1.6 ± 0.2

634

635

636

637 **Figure captions**

638

639 Figure 1(a): A color-shaded relief map with a simplified active tectonic map of Asia and
640 surrounding regions (modified from Fu et al. (2011)). (b): Interpretative geologic map of
641 three roughly east-west stretching fold and fault zones along the northern piedmont of
642 Tian Shan (modified from Sun et al. (2009)). The dip of the Hutubi and Manas faults at
643 the surface were from Avouac et al. (1993) and Zhao et al. (2001), respectively.

644

645 Figure 2(a): Landsat TM image showing the deformational features of the Manas and
646 Hutubi faults in the northern piedmont of Tian Shan. (b): Schematic geomorphology of
647 Manas River terraces within the Manas Anticline area. The terraces (T-5 and T-6) were
648 deformed intensively by the Hutubi and Manas faults, forming fresh-looking fault
649 scarps..

650

651 Figure 3: Conceptual graph showing the transportation and deposition of aeolian dust
652 from the Junggar Desert to the northern piedmont of Tian Shan (modified from Sun
653 (2002)).

654

655 Figure 4 (a): Google Earth map showing the studied sites and the locations of the fault
656 scarps. (b): A schematic graph showing the geomorphology of the T-5 and T-6 of the
657 Manas River. Six exposures (HTB0, HTB1, MNS1-MNS4) were studied to decipher late
658 Quaternary faulting along the Hutubi and Manas faults. The cumulative vertical
659 displacements of the reverse faults were measured from B-B', C-C' and D-D'. (c): A
660 schematic graph showing the relation among the surface ruptures, fault scarp and terrace
661 tread.

662

663 Figure 5: The exposures of the ruptures along the fault scarps and detailed OSL dating
664 sampling positions and (a) (HTB0) (we could not find any appropriate loess samples for
665 OSL dating, as the original fault scarp was destroyed as a result of the development of a
666 gully along the fault); (b) (HTB1); (c) (MNS1); (d) (MNS2); (e) (MNS3); (f) (MNS4).

667

668 Figure 6: Schematic diagrams showing the loess infilling at the surface ruptures along the
669 reverse fault scarps, in response to past faulting events. The process is illustrated with six
670 representative stages.

671

672 Figure 7: Single-aliquot regeneration (SAR) equivalent dose versus pre-heat temperature
673 for MNS2-4. The mean values with the standard errors are obtained from results of six
674 aliquots. The horizontal line denotes the pre-heat plateau was reached from 220 °C to
675 260 °C.

676

677 Figure 8: Typical OSL shine-down curves and OSL growth curves of quartz grains from
678 four loess samples (HTB1-4, MNS1-5, MNS2-1 and MNS3-4).

679

680 Figure 9: Radial plot results for D_e distributions of six loess samples from the five
681 exposures (HTB1, MNS1-MNS4). In the plot, the D_e and the OSL ages of the six
682 samples were calculated with the mean age model.

683

684 Figure 10: OSL dating results of the loess samples from the five exposures (HTB1,
685 MNS1-MNS4) along the scarps of the Hutubi and Manas faults.

686

Figure 1(a)

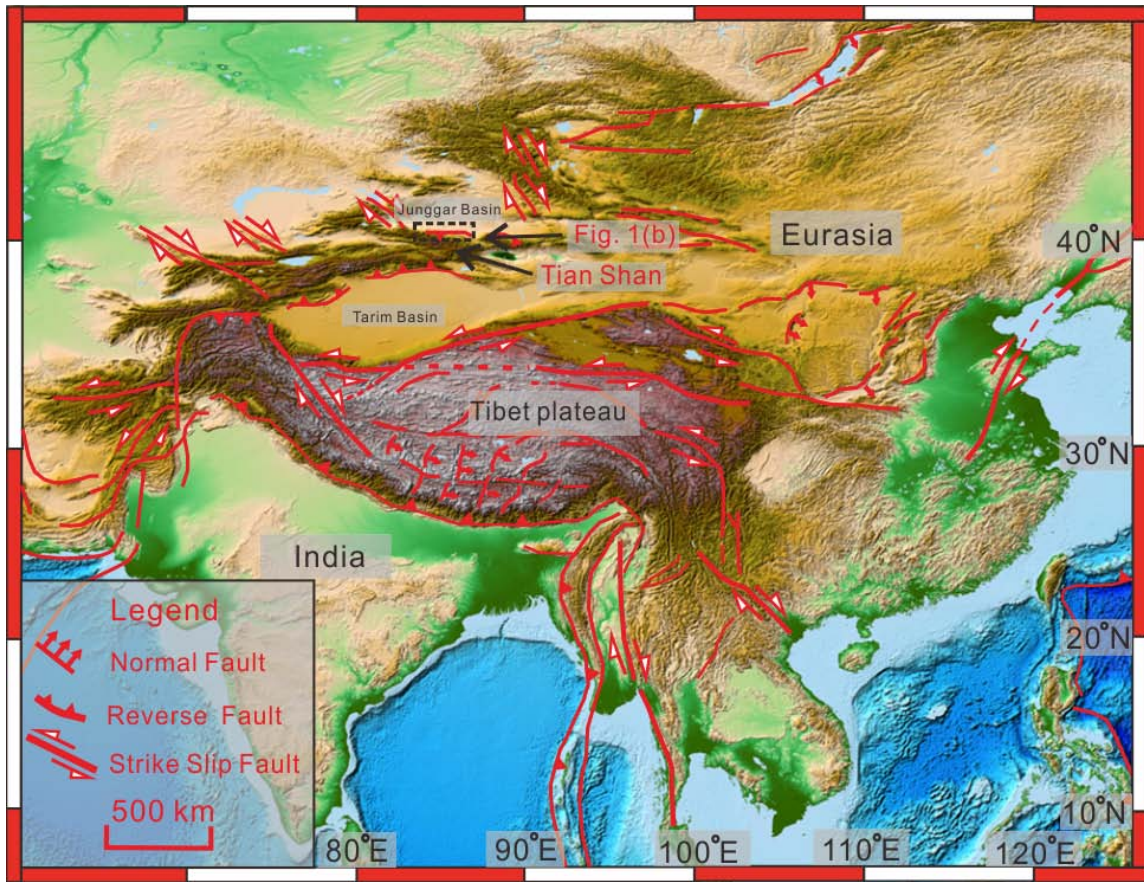


Figure 1(b)

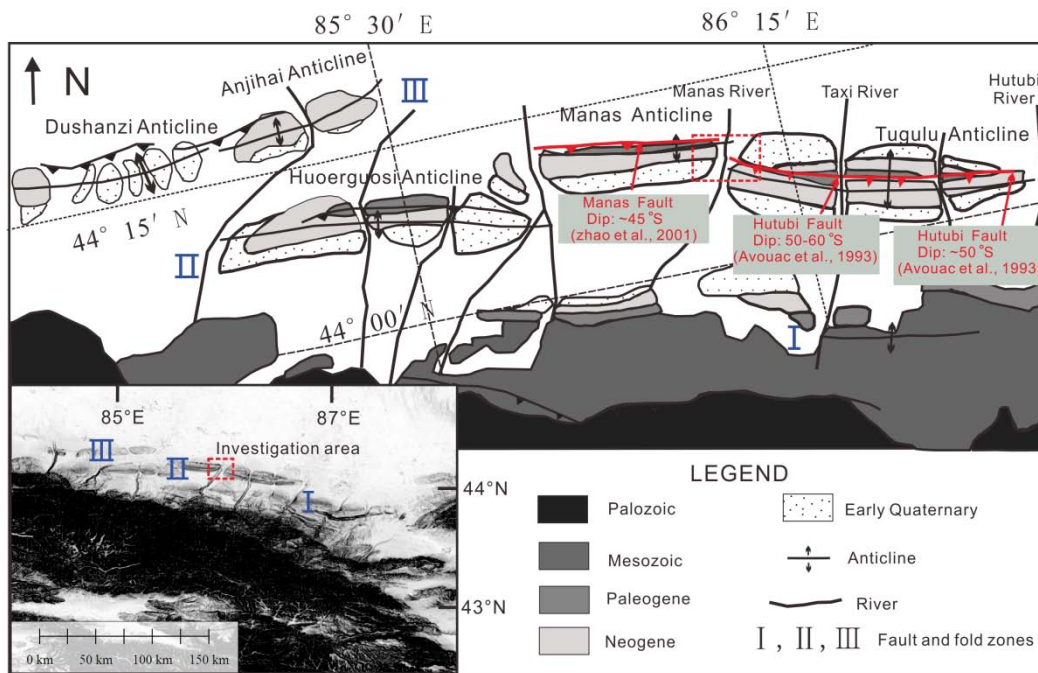


Figure 2(a)

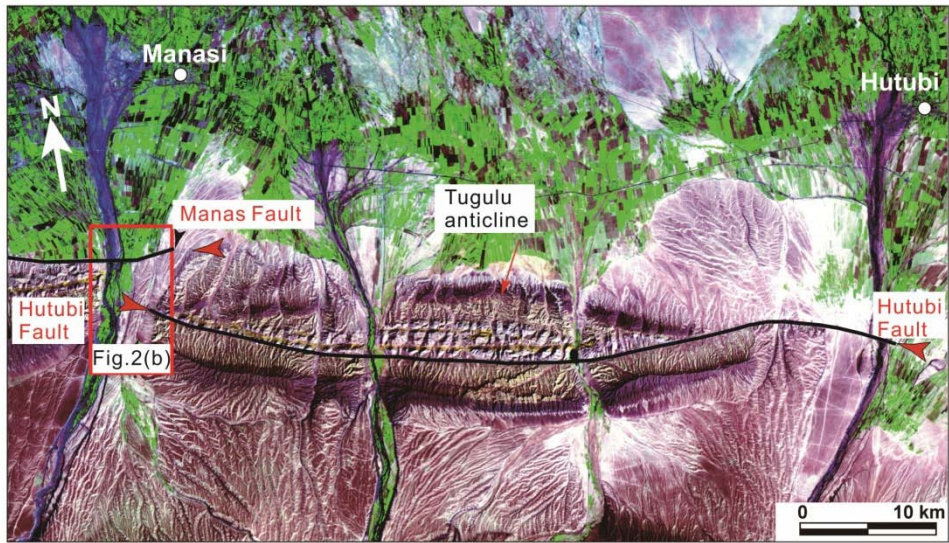


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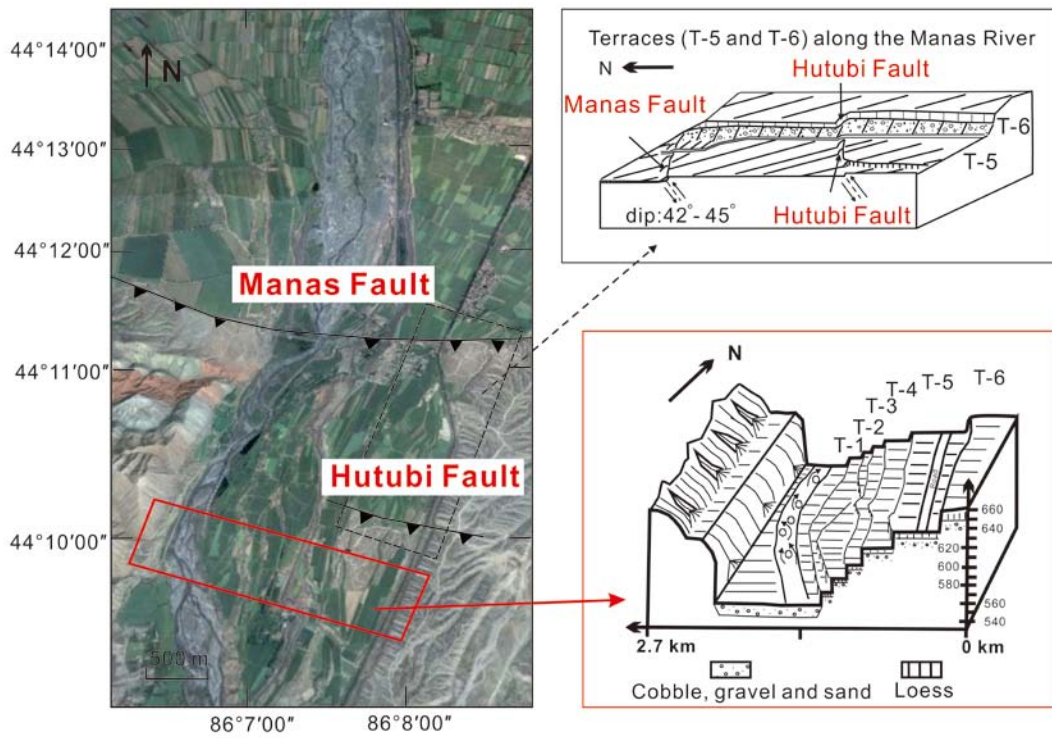


Figure 3

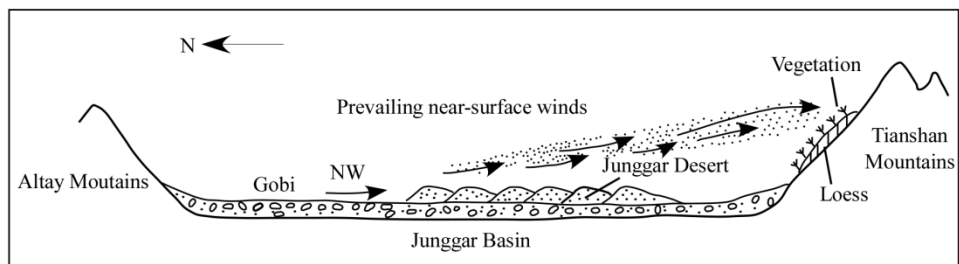
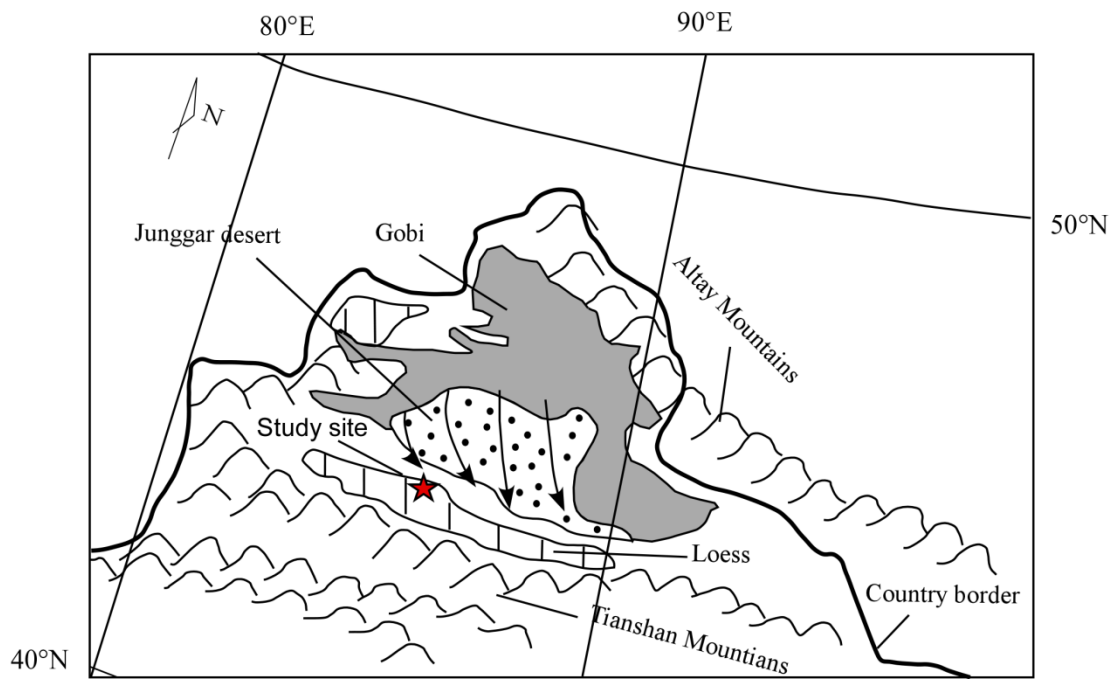


Figure 4(a)

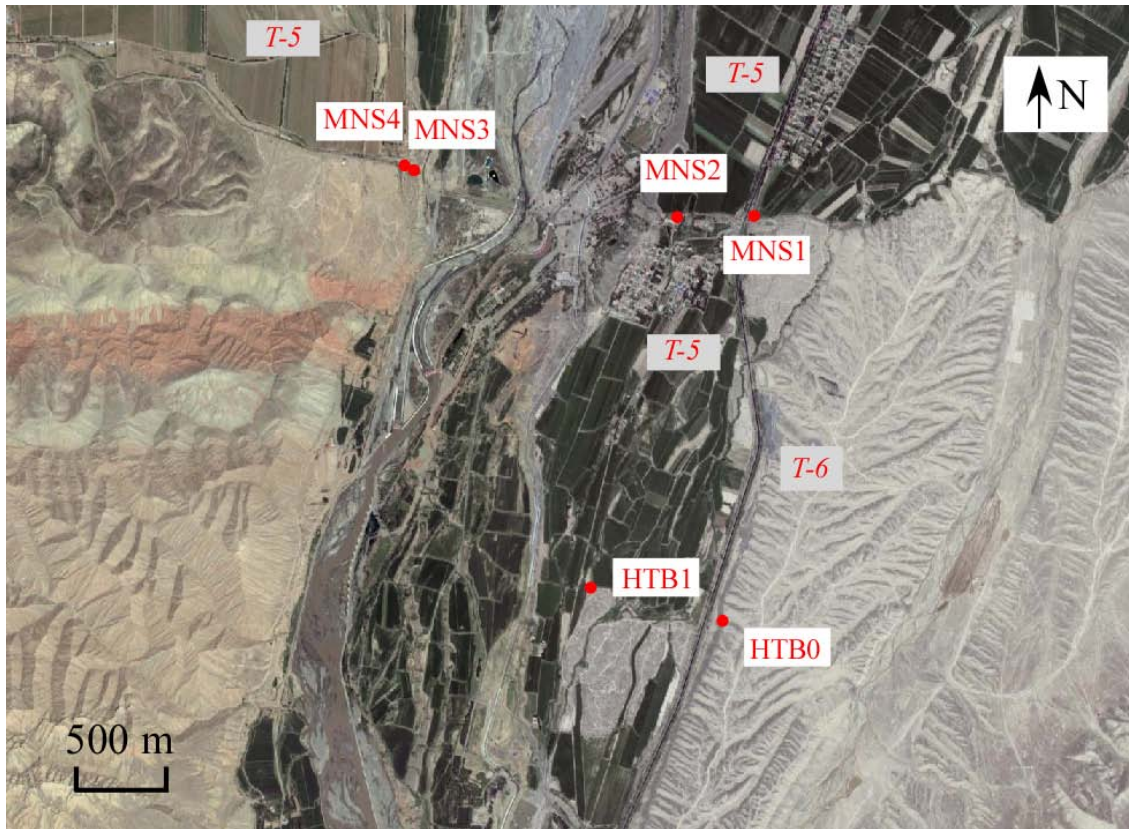


Figure 4(b)

Fault scarps and trenches studied at T-5 and T-6 along the Manas River

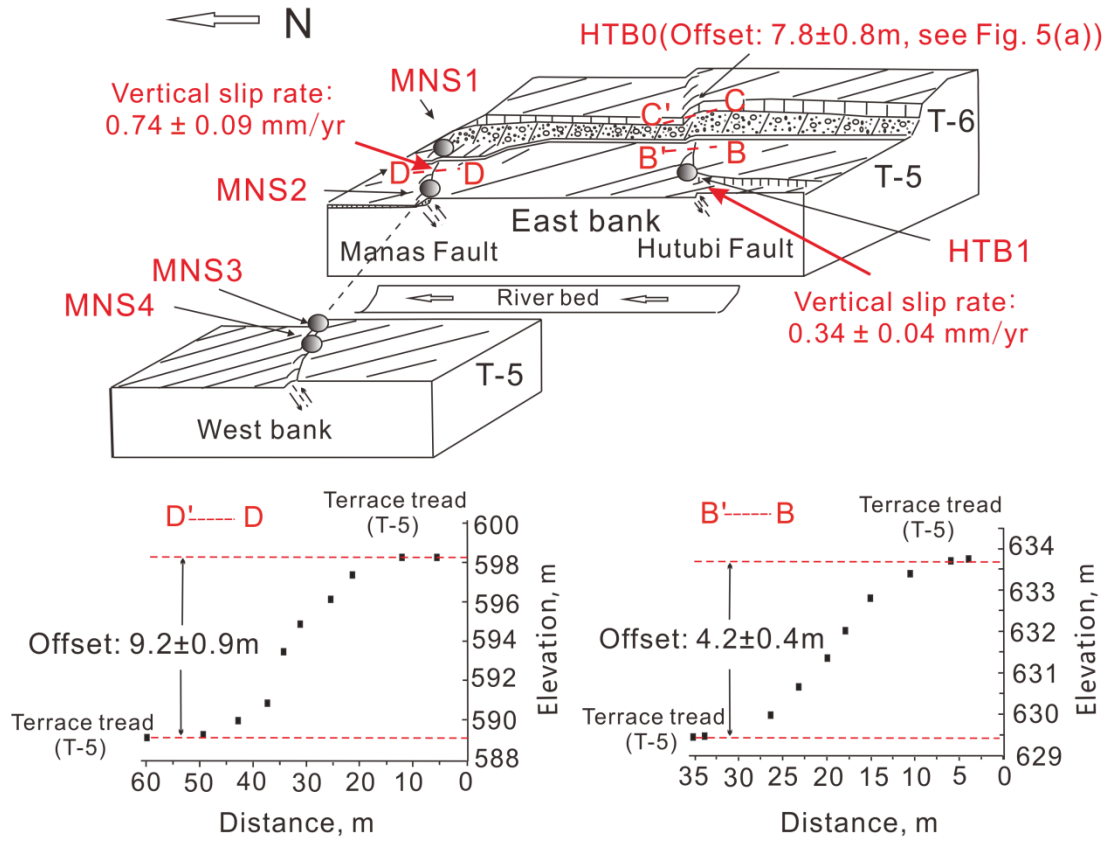


Fig. 4(c)

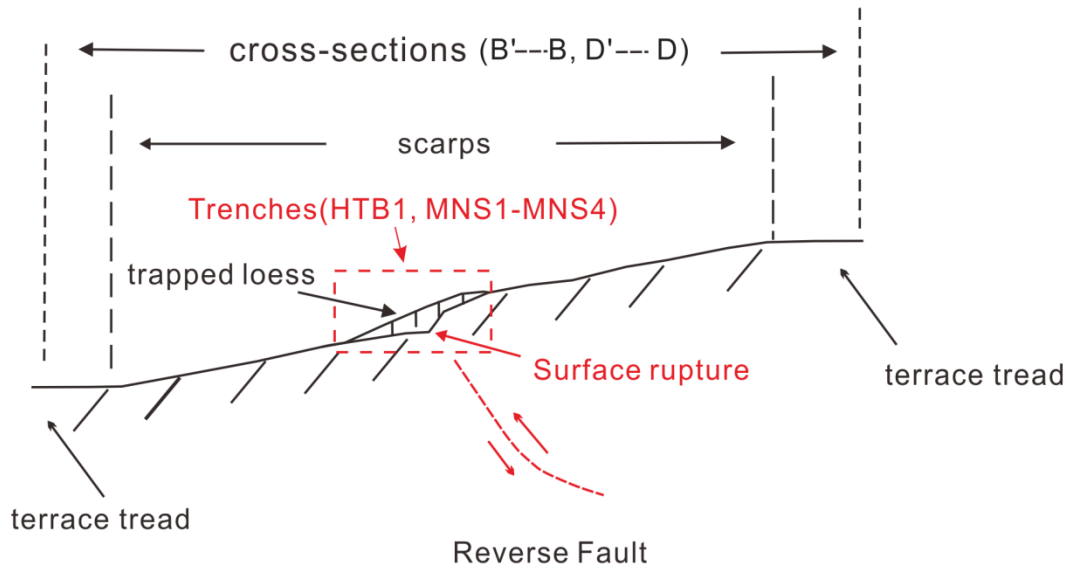


Figure 5(a)

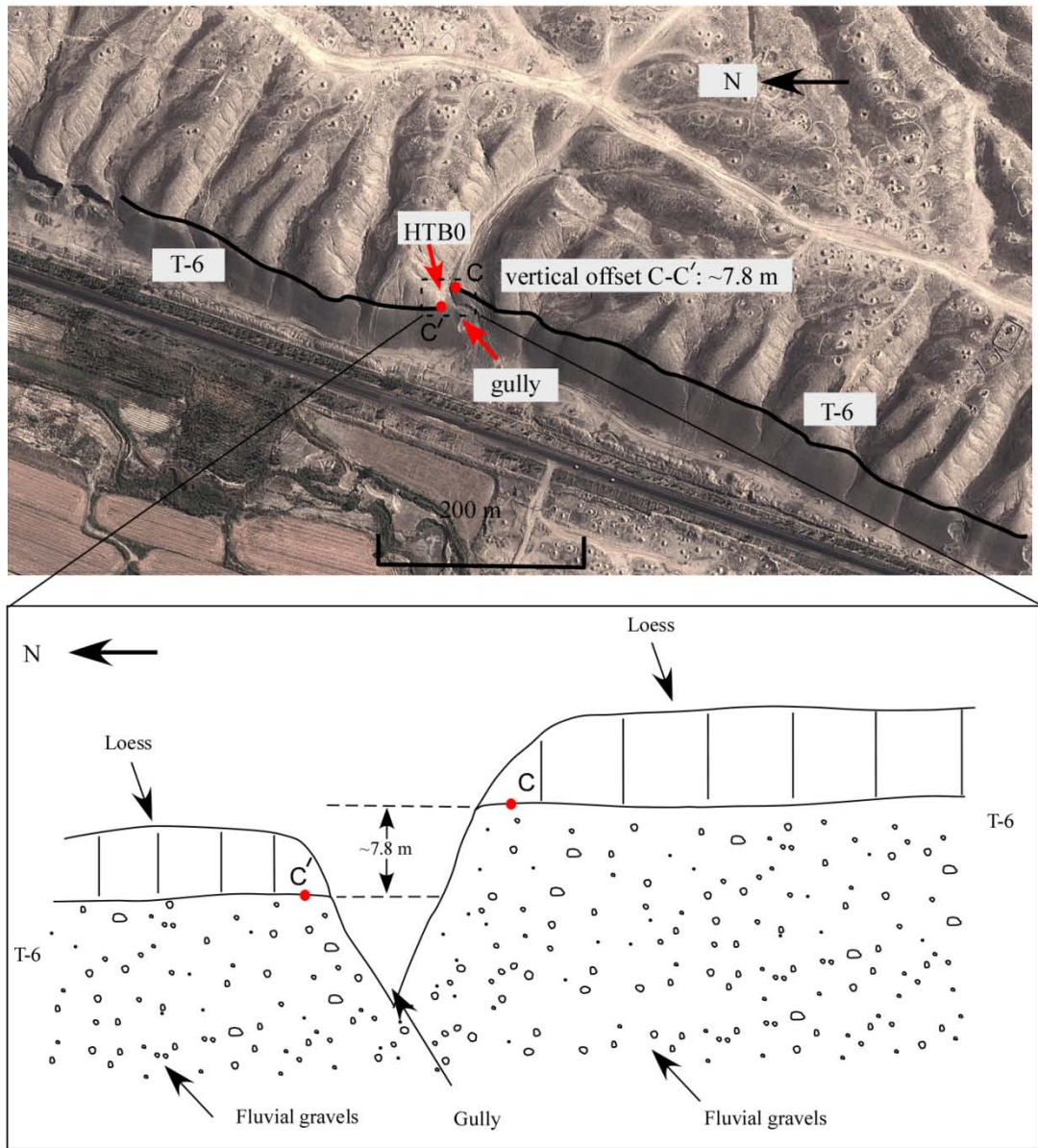


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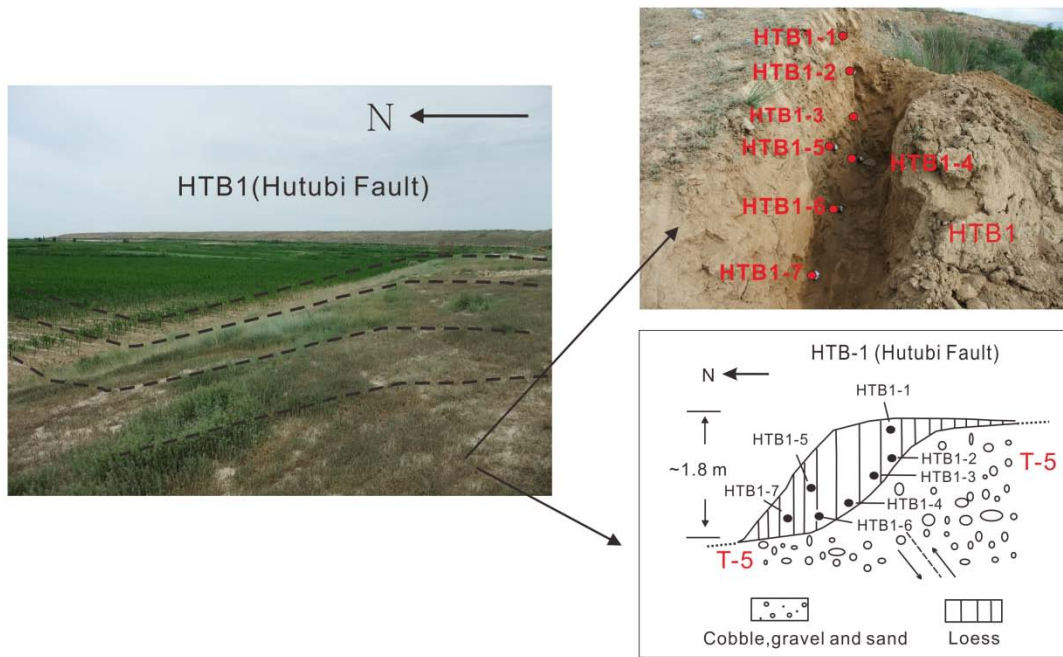


Figure 5(c)

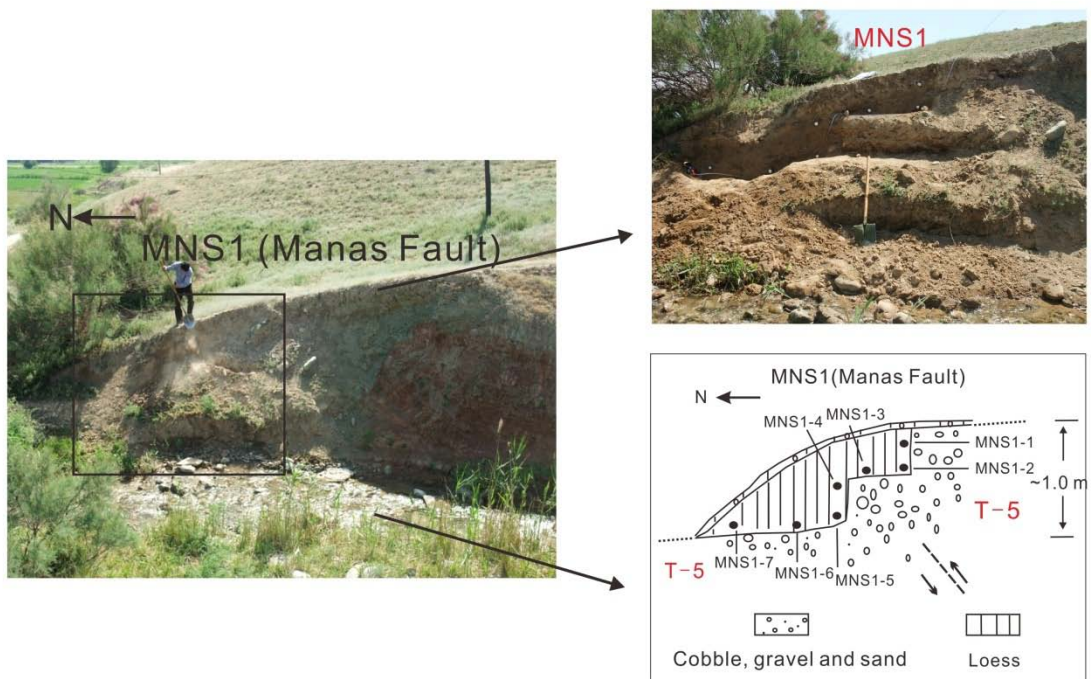


Figure 5(d)

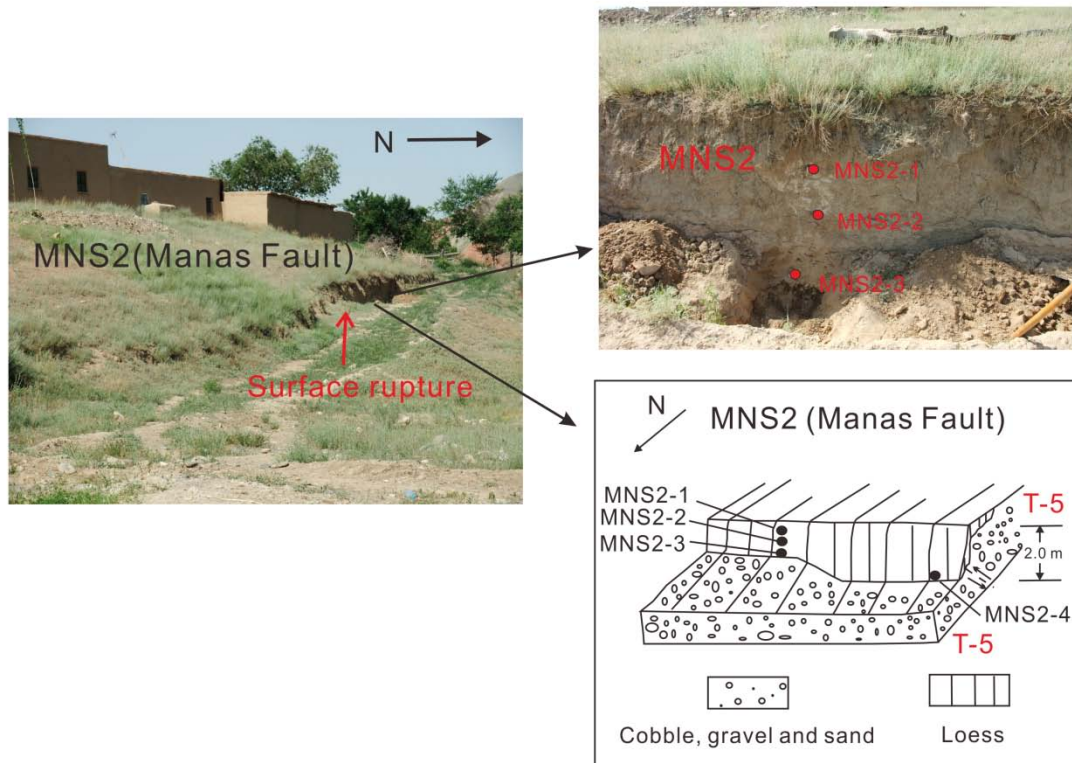


Figure 5(e)

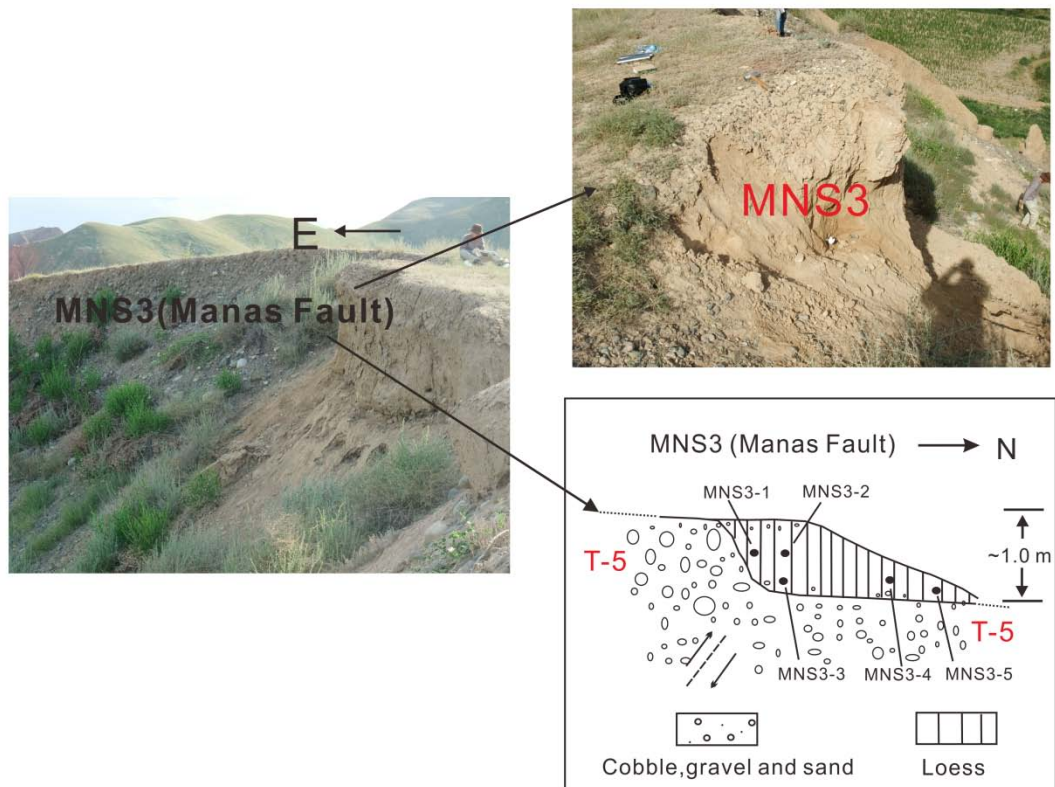


Figure 5(f)

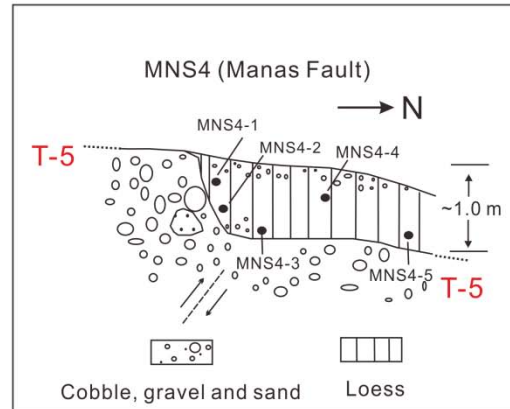
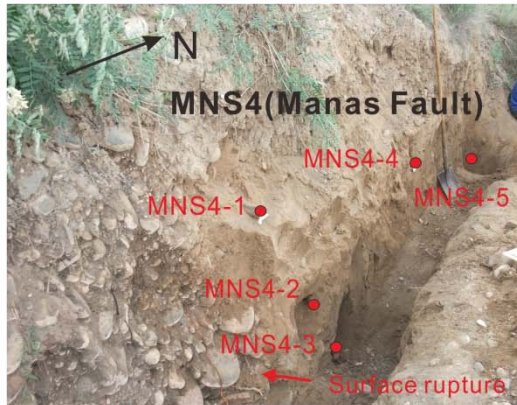


Figure 6

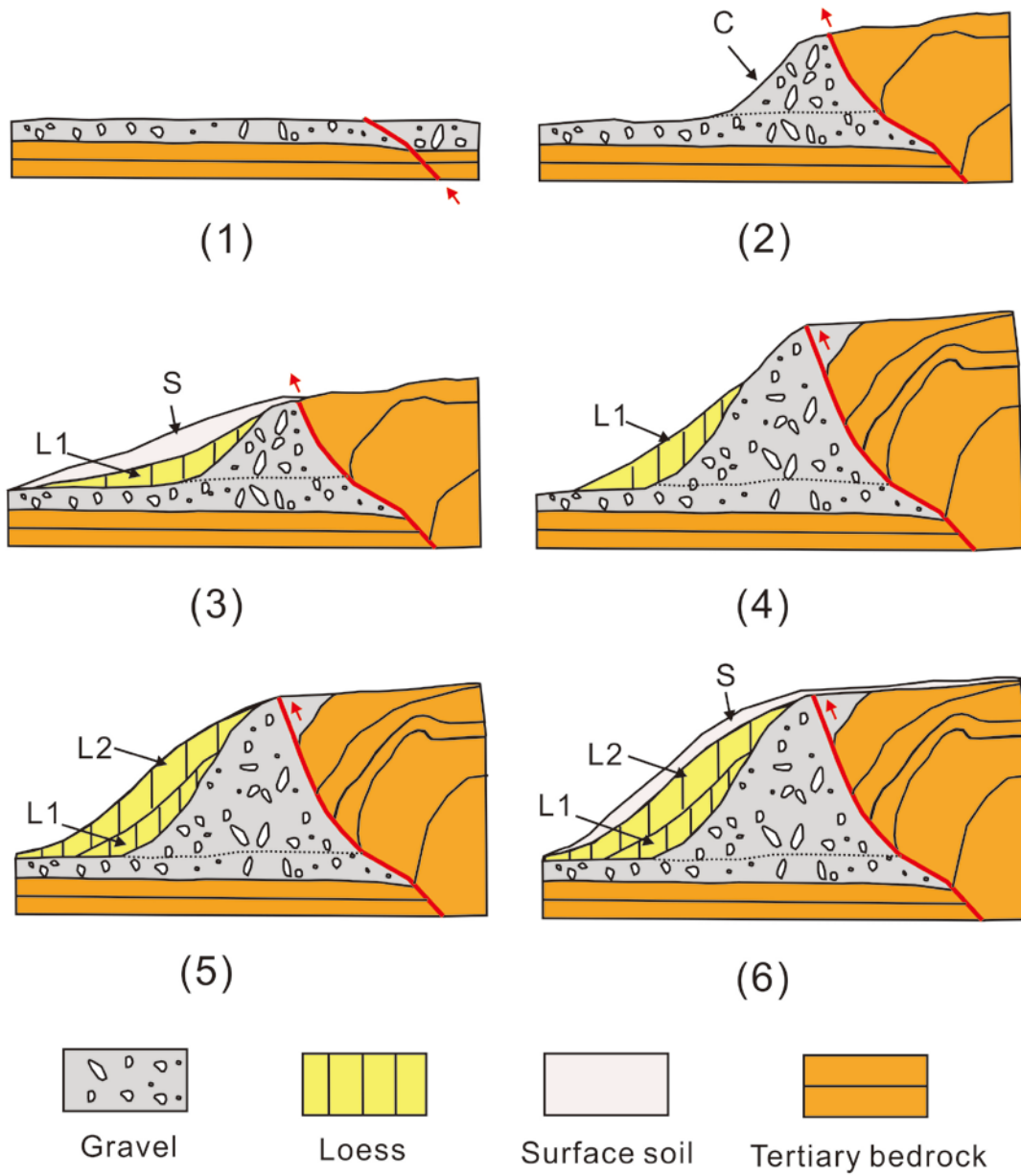


Figure 7

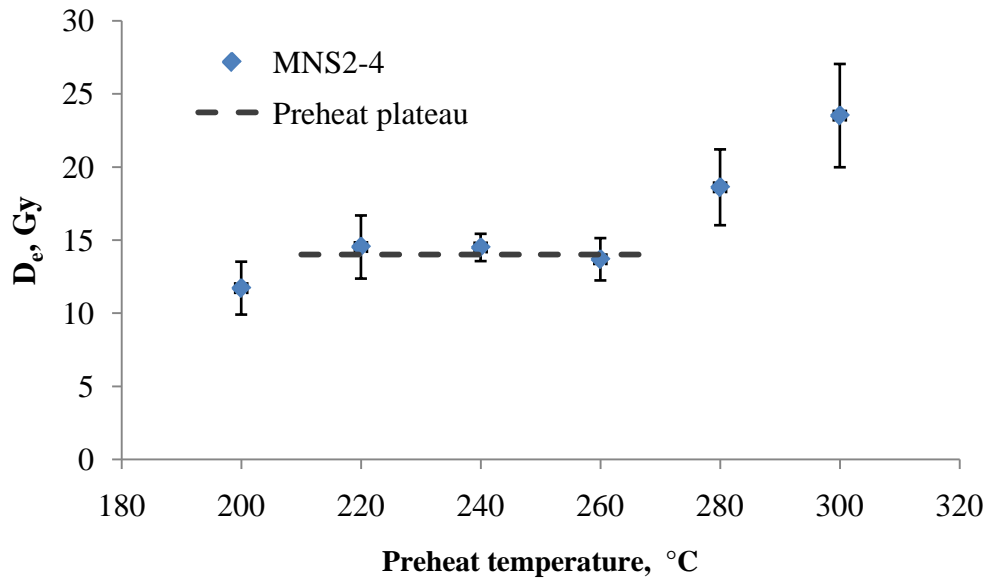


Figure 8

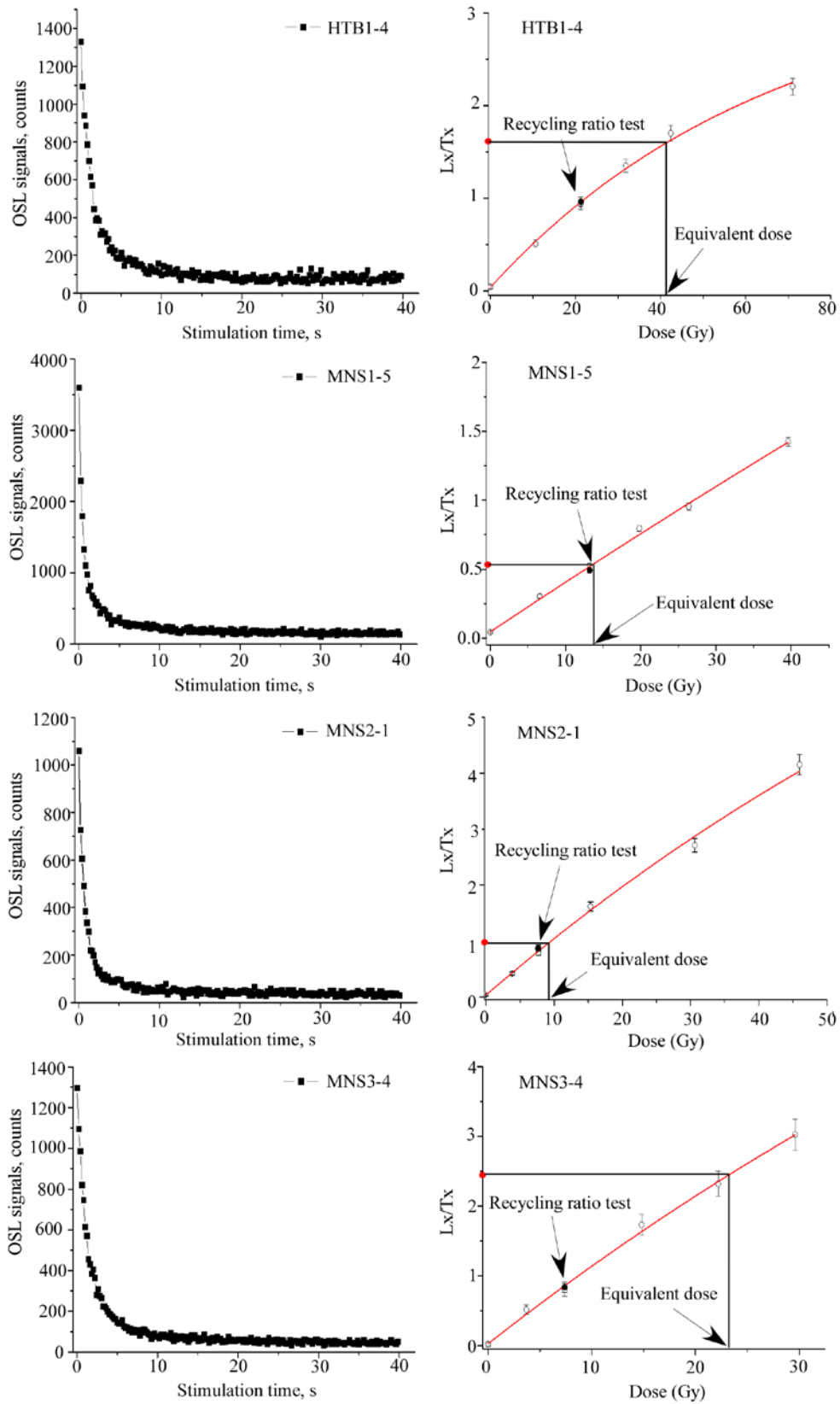


Figure 9

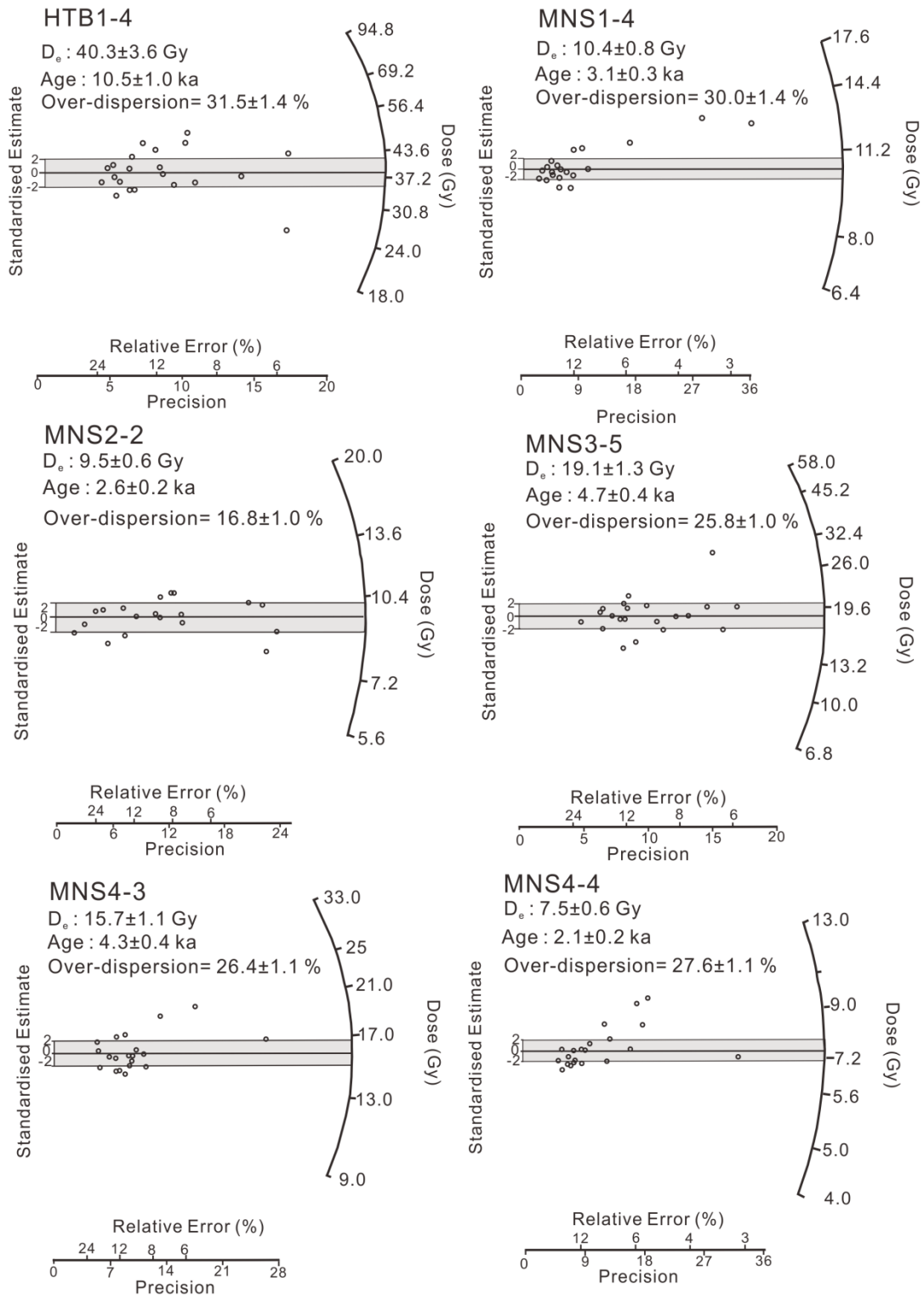


Figure 10

