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1 **Cenozoic deformation of the Tarim Basin and surrounding ranges (Xinjiang, China):**

2 **A regional overview**

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17 **Abstract**

18 With its central position between the Tibetan Plateau and the Tian Shan Range, the Tarim
19 Basin is a key element of the Cenozoic Asian orogenic system. However, a comprehensive
20 regional study, and more particularly the quantification of shortening through this basin and its
21 margins, are still needed to understand its role in the Cenozoic deformation of Asia. From a
22 compilation of previous works, together with an extensive dataset of satellite, field, seismic and
23 well data, we provide a tectonic map of the Cenozoic structures and four balanced geological
24 transects of the Tarim Basin and its surrounding ranges. Based on this map and these cross-
25 sections, we characterize the Cenozoic deformation of the original Proterozoic Tarim block.
26 From structural restorations and crustal budgets, we also quantify the compressive component of
27 this deformation. Most of the Cenozoic compressive deformation (from ~94% to 100%) is
28 concentrated in the ranges along the block margins. To the west, up to 78 ± 23 km and $54 +24/-$
29 18 km of crustal shortening are accommodated across the compressive Western Kunlun and
30 Southwestern Tian Shan ranges, while to the east, up to 38.6 ± 18 km and $15 +20/-15$ km are
31 accommodated across the transpressive Altyn Tagh and Southeastern Tian Shan ranges. A non-

32 negligible amount of compressive deformation (up to ~6%) is also accommodated within the
33 Tarim Basin by large basement-cored uplifts with a vergence synthetic to the deformation of the
34 Tibetan Plateau edge. To the west, the Bachu uplift absorbs ~5 km of the total crustal shortening
35 of the Western Kunlun thrust system, while to the south, the Tanan uplift accumulates ~0.6 km
36 of the Altyn Tagh strike-slip system. Structural inheritance has a major influence on the
37 Cenozoic deformation since ~33.3% to 100% of the total shortening is accommodated by
38 reactivated basement structures inherited from the Protero-Paleozoic history of the Tarim block.
39 Finally, we argue that the basement-cored uplifts in the centre of the basin imply a deformation
40 transfer from the Tibetan Plateau to the Tian Shan, above a deep crustal decollement decoupling
41 the deforming crust from an underlying rigid mantle.

42

43 **Keywords**

44 Asian Cenozoic deformation, Tarim Basin, Western Kunlun, Altyn Tagh, Tian Shan, shortening,
45 balanced geological cross-sections

46

47

48 **1. Introduction**

49

50 With the Himalaya-Tibetan Plateau orogenic system and active deformation as far as
51 3000 km north of the India-Asia collision zone (including the Tian Shan, Altai Range and Baikal
52 rift system) ([Fig. 1](#)), Asia is an outstanding natural laboratory to study continental deformation.
53 Following the pioneering studies of the late 1970s and early 1980s (e.g., [Molnar and Tapponnier,](#)
54 [1975, 1977; Tapponnier and Molnar, 1979; Molnar, 1984; England and Houseman, 1985](#)),
55 multiple investigations on the Cenozoic Asian mountain ranges, but also on the associated
56 sedimentary basins, which are key recorders of their tectonic and climatic evolution, have been
57 conducted to decipher the modalities of this deformation (e.g., [Avouac et al., 1993; Matte et al.,](#)
58 [1996; Métivier et al., 1998; Lavé and Avouac, 2001; Tapponnier et al., 2001; Yang and Liu,](#)
59 [2002; Yin et al., 2002; Guillot et al., 2003; Clark et al., 2006; DeCelles et al., 2007; Liu-Zeng et](#)
60 [al., 2008; Charreau et al., 2009; Lu et al., 2010; Cheng et al., 2015, 2016a; Coutand et al., 2016;](#)
61 [Gourbet et al., 2017; Jolivet et al., 2018](#)). Among these basins, the Tarim Basin occupies a
62 remarkable central position within the orogenic system, between the Tibetan Plateau and the

63 Tian Shan Range (Fig. 1). With a continuous Proterozoic to Present sedimentation, this wide
64 basin is thus a first-class repository for the geological history of the Asian continent (e.g., Li et
65 al., 1996; Métivier et al., 1999; Yin et al., 2002; Kent-Corson et al., 2009; Sun et al., 2009b; Lin
66 et al., 2012c; Bosboom et al., 2014a; Tang et al., 2014; Lin et al., 2015a; Liu et al., 2015a; He et
67 al., 2016; Zhu et al., 2017; Morin et al., 2018; Wu et al., 2018; Jiang et al., 2018). Furthermore,
68 the Tarim Basin seems to be a key element in the Cenozoic Asian deformation. Indeed, it is
69 usually considered as an undeformed rigid block, playing the role of an heterogeneity in the
70 strain field and/or of a secondary indenter that transmits the compressive stresses at the
71 continental scale far to the north of the India-Asia collision zone (e.g., England and Houseman,
72 1985; Avouac and Tapponnier 1993; Neil and Houseman, 1997; Mattauer et al., 1999; Wang et
73 al., 2007; Dayem et al., 2009; Craig et al., 2012; Calignano et al., 2015).

74 For these reasons, numerous studies have been performed on the Cenozoic deformation
75 and sedimentation along the margins or within the centre of this basin (e.g., Avouac et al., 1993;
76 Matte et al., 1996; Wittlinger et al., 1998, 2004; Allen et al., 1999; Burchfield et al., 1999; Yin et
77 al., 2002; Scharer et al., 2004; Sun et al., 2005, 2009b; Heermance et al., 2007; Hubert-Ferrari et
78 al., 2007; Fu et al., 2010; Turner et al., 2010; Li et al., 2012a; Jiang et al., 2013; Wei et al., 2013;
79 Chang et al., 2014a; Cheng et al., 2016b; Li et al., 2016b; Li et al., 2016e; Lu et al., 2016; Wang
80 and Wang, 2016; Guilbaud et al., 2017; Thompson et al., 2017; Izquierdo-Llavall et al., 2018;
81 Jiang et al., 2018). However, they were mainly undertaken to characterize tectonic structures and
82 their activity at a local scale. Moreover, the few studies conducted at a regional scale remain
83 schematic or imprecise in terms of hierarchy and relationships between the structures, with
84 sometimes no detail about which ones were activated or not during the Cenozoic (e.g., Wei et al.,
85 2013; Tang et al., 2014; Lin et al., 2015a; Li et al., 2016e; Jiang et al., 2018). Nevertheless, it is
86 clear that the original Proterozoic Tarim block is not totally rigid, with deformation along the
87 margins and even in the center of the Tarim Basin (e.g., Avouac et al., 1993; Matte et al., 1996;
88 Allen et al., 1999; Wittlinger et al., 2004; Guo et al., 2005; Heermance et al., 2007; Li et al.,
89 2012a; Tong et al., 2012; Chang et al., 2014a; Tang et al., 2014; Fig. 15 of Wang et al., 2014a; Li
90 et al., 2016e; Daout et al., 2018; Izquierdo-Llavall et al., 2018; Jiang et al., 2018). Yet, the
91 amount and distribution of this deformation, as well as the type of structures accommodating it,
92 still need to be precisely documented. A comprehensive regional study, and more particularly the
93 quantification of shortening through this area, is therefore necessary to further discuss these

94 issues, and to understand how the Tarim participates to the propagation of the Cenozoic
95 compressive deformation within Asia.

96 In this article, we present such a comprehensive study of the Cenozoic deformation at the
97 scale of the whole Tarim Basin and its surrounding ranges. Thanks to a compilation of existing
98 geological maps and of published cross-sections, together with an extensive dataset of surface
99 (satellite images and field constraints) and subsurface (seismic profiles and well data) data, we
100 hereafter propose a tectonic map and four balanced geological transects highlighting the
101 structures activated during the Cenozoic in the Tarim area. On the basis of this map and these
102 transects, we characterize and quantify the Cenozoic deformation across the whole basin and its
103 margins, as well as its relationships with the surrounding ranges.

104

105 **2. Geological setting**

106

107 *2.1. Tarim Basin*

108

109 The Tarim Basin is located in Central Asia, in the Xinjiang Province in China (Fig. 1). It
110 corresponds to a ~1500-km-long and ~700-km-wide active sedimentary basin with an average
111 elevation of ~1200 m (Fig. 1). Presently, this basin is an endorheic compressive basin
112 surrounded by mountain ranges: the Western Kunlun Range to the southwest, the Altyn Tagh
113 Range to the southeast, and the Tian Shan to the north.

114 This region of Central Asia was initially structured during the Proterozoic and Paleozoic
115 by the accretion of several blocks: the Kunlun and Qaidam blocks to the south, the Tarim block
116 in the centre, and the Central/South Tian Shan block to the north (e.g., Windley et al., 1990;
117 Matte et al., 1996; Sobel and Arnaud, 1999; Zhou et al., 2001; Xiao et al., 2004, 2013; Charvet et
118 al., 2011; Lin et al., 2013; Zhang et al., 2013a; Li et al., 2018a). Prior to this accretion, the
119 tectonic evolution of the Tarim block was dominated by extension from Neoproterozoic to
120 Ordovician (e.g., Li et al., 1996; Zhang et al., 2013a; Tang et al., 2014; Liu et al., 2015a; Wang et
121 al., 2015; Dong et al., 2016; Zhu et al., 2017; Ren et al., 2018; Wu et al., 2018). Then, from
122 Ordovician to Permian, compression generalized in the area due to the closures of oceanic
123 domains, followed by collisions between (1) the Tarim and Kunlun blocks to the southwest
124 (from Ordovician to Devonian), (2) the Tarim and Qaidam blocks to the southeast (from Silurian

125 to Devonian), and (3) the Tarim and Central/South Tian Shan blocks to the north (from Devonian
126 to Permian) (e.g., Windley et al., 1990; Li et al., 1996; Matte et al., 1996; Sobel and Arnaud,
127 1999; Gehrels et al., 2003; Xiao et al., 2004, 2013; Charvet et al., 2011; Lin et al., 2012c; Lin et
128 al., 2013; Zhang et al., 2013a; Tang et al., 2014; Alexeiev et al., 2015; Liu et al., 2015; Han et al.,
129 2016; He et al., 2016; Li et al., 2018a). Driven by all these oceanic closures and subsequent
130 collisions, Paleozoic ranges comprising ophiolitic sutures (Fig. 1), as well as plutonic and
131 metamorphic rocks, formed around the Tarim block (e.g., Windley et al., 1990; Matte et al.,
132 1996; Gao et al., 1998; Sobel and Arnaud, 1999; Cowgill et al., 2003; Gehrels et al., 2003; Xiao
133 et al., 2005, 2013; Cowgill, 2010; Charvet et al., 2011; Glorie et al., 2011; Cao et al., 2013;
134 Zhang et al., 2014a; Li et al., 2018a). In the contemporaneous Tarim Basin settled in the centre of
135 the block, this Protero-Paleozoic history also induced the development of large structures known
136 as the Bachu, Tanan, Tazhong, Tadong, Tabei, Kalpin, and Kuruk Tagh basement-cored uplifts,
137 which were created in extension at the end of the Proterozoic-beginning of the Paleozoic, before
138 being reactivated in compression (e.g., Li et al., 1996; Allen et al., 1999; Carroll et al., 2001;
139 Turner et al., 2010; Lin et al., 2012b, 2012c; Zhang et al., 2013a; Gao and Fan, 2014; Tang et al.,
140 2014; Lin et al., 2015a; Liu et al., 2015a; He et al., 2016; Zhu et al., 2017; Wu et al., 2018).

141 At the end of the Paleozoic, the emplacement of a Permian Large Igneous Province (LIP)
142 affected the Tarim block (e.g., Zhang et al., 2008; Zhang et al., 2010; Zhang et al., 2012b; Yang
143 et al., 2013; Li et al., 2014a; Xu et al., 2014; Yang et al., 2016; Loury et al., 2018). Afterwards,
144 during the Mesozoic, other blocks (Qiangtang and Lhasa blocks) were accreted further south, in
145 the palaeo-Tibetan Plateau region (e.g., Matte et al., 1996; Yin and Harrison, 2000; Pan et al.,
146 2012; Jolivet, 2017; and references therein). Within the Tarim region, these events produced a
147 complex zone of deformation, with minor strike-slip, compressive or extensive reactivations of
148 the inherited Protero-Paleozoic structures (e.g., Hendrix et al., 1992; Li et al., 1996; Sobel, 1999,
149 2001; Dumitru et al., 2001; Arnaud et al., 2003; Chen et al., 2003; De Grave et al., 2007; Jolivet
150 et al., 2010, 2013; Wang et al., 2012; Chang et al., 2014b; Tang et al., 2014; Li et al., 2015a;
151 Morin et al., 2018; Yang et al., 2018b). Finally, during the Cenozoic, India collided into the
152 Asian continent around 55-45 Ma (e.g., Wang et al., 2014a and references therein). This last
153 collision led to a major compressive reactivation of the mountain ranges surrounding the Tarim
154 block, as well as of inherited structures within the Tarim Basin (e.g., Molnar and Tapponnier,
155 1975; Tapponnier and Molnar, 1977, 1979; Windley et al., 1990; Avouac et al., 1993; Allen et

156 al., 1994; Hendrix et al., 1994; Matte et al., 1996; Jolivet et al., 1999, 2001, 2010; Dumitru et al.,
157 2001; Yin et al., 2002; De Grave et al., 2007; Turner et al., 2010; Yin, 2010 and references
158 therein; Tong et al., 2012; Tang et al., 2014; Li et al., 2015b and references therein; Li et al.,
159 2016e; Jiang et al., 2018; Yang et al., 2018b). In this study, because these ranges involve parts of
160 the block itself, we refer to the Tarim block as the area comprised in between the various
161 Paleozoic ophiolitic sutures located in the mountain belts (Fig. 1), rather than as the present-day
162 Tarim Basin only.

163 Resulting from this long geological history, the Tarim Basin stratigraphy is characterized
164 by a Proterozoic basement composed by igneous rocks, overlain by a thick, more or less
165 continuous series of sediments from Neoproterozoic (Sinian) to Quaternary (Fig. 2) (e.g., Li et
166 al., 1996; Carroll et al., 2001; Yin et al., 2002; Guo et al., 2005; Wang et al., 2009; Shu et al.,
167 2011; Lin et al., 2012a, 2012c; Zhang et al., 2013a; Tang et al., 2014; Guo et al., 2015; Lin et al.,
168 2015a; Liu et al., 2015a; He et al., 2016; Li et al., 2016d; Zheng et al., 2015; Tian et al., 2016;
169 Zhu et al., 2017; Wu et al., 2018). During the Sinian and Paleozoic, marine calcareous and
170 terrigenous deposits were dominant until the Carboniferous (Fig. 2) (e.g., Carroll et al., 2001;
171 Chen and Shi, 2003; Shu et al., 2011; Lin et al., 2012c; Zhang et al., 2013a; Gao and Fan, 2014;
172 Li et al., 2015a; Liu et al., 2015a; He et al., 2016; Zhu et al., 2017; Wu et al., 2018). In the
173 western part of the basin, the Paleozoic sediments also contain an upper Cambrian interval of
174 gypsiferous shales (Fig. 2), which is a major decollement level in this area (e.g., Allen et al.,
175 1999; Lin et al., 2012a, 2012c; Li et al., 2016c, 2016e; Lu et al., 2016; Yang et al., 2018b).
176 Within the Neoproterozoic-Paleozoic series, field observations, seismic profiles and well data
177 document unconformities, together with sedimentary thickness and facies changes at the top of
178 the basement-cored structures in the Tarim. Combined with thermochronological data, this
179 shows that these structures were already active during the early (Sinian to Devonian) stages of
180 the basin evolution (e.g., Allen et al., 1999; Carroll et al., 2001; Dumitru et al., 2001; Chen and
181 Shi, 2003; Tang et al., 2008; Turner et al., 2010, 2011; Chang et al., 2012b; Lin et al., 2012b,
182 2012c; Gao and Fan, 2014; Lin et al., 2015a; Liu et al., 2015a; He et al., 2016; Zhang et al., 2016;
183 Zhu et al., 2017; Wu et al., 2018).

184 During the Permian, a marine to continental transition occurred, with a change from
185 calcareous to terrigenous deposits characterizing most of the Mesozoic sediments (Fig. 2) (e.g.,
186 Hendrix et al., 1992; Gao et al., 1998; Sobel, 1999; Carroll et al., 2001; Chen and Shi, 2003; Guo

187 et al., 2015; Li et al., 2015a; Li et al., 2016d; Morin et al., 2018; Peng et al., 2018). A thick (up to
188 ~10 km) and continuous Cenozoic sedimentary series, mainly supplied by the erosion of the
189 surrounding mountain ranges, was then trapped in the contemporaneous compressive Tarim
190 Basin (e.g., Yin et al., 1998, 2002; Métivier and Gaudemer, 1997; Heermance et al., 2007; Wang
191 et al., 2009; Zheng et al., 2015; Tian et al., 2016). From the late Cretaceous to Eocene, the Tarim
192 deposits were mostly continental and terrigenous, except in the western part of the basin where
193 an alternation of marine transgressions/regressions occurred (e.g., Yin et al., 1998; Wei et al.,
194 2013; Bosboom et al., 2014a, 2014b, 2017; Tian et al., 2016; Chen et al., 2018; Zhang et al.,
195 2018; Kaya et al., 2019). In that area, the marine Paleocene-Eocene Kashgar Group is thus
196 composed of gypsum, limestones, siltstones, and sandstones related to these marine incursions
197 (Fig. 2) (e.g., Jin et al., 2003; Bosboom et al., 2011, 2014a, 2014b, 2017; Wei et al., 2013; Wang
198 et al., 2014; Zheng et al., 2015; Zhang et al., 2018; Blayney et al., 2019). This Paleocene-Eocene
199 evaporitic assemblage corresponds to another major decollement level in the southwest of the
200 Tarim (Southwest depression) (e.g., Li et al., 2016e; Lu et al., 2016). At the end of the Eocene,
201 the western marine deposits ended while a continental sedimentation finally developed
202 throughout the basin (e.g., Yin et al., 1998, 2002; Jin et al., 2003; Bosboom et al., 2014a; Zheng
203 et al., 2015; Tian et al., 2016; Zhang et al., 2018). During the Oligocene and early Miocene, the
204 sediments mostly consist in claystones, siltstones, and sandstones (Fig. 2) (e.g., Yin et al., 1998,
205 2002; Jin et al., 2003; Heermance et al., 2007; Wang et al., 2009; Zheng et al., 2015; Tian et al.,
206 2016; Blayney et al., 2019). In these series however, the lacustrine Eocene-Oligocene
207 Kumugeliemu Group and Suweiyi Formation are made up of gypsum, siltstones, sandstones, and
208 a few conglomerates (Fig. 2) (e.g., Yin et al., 1998; Huang et al., 2006; Wang et al., 2009; Tian
209 et al., 2016). This Eocene-Oligocene evaporitic assemblage is also a major decollement level in
210 the north of the basin (Kuqa depression) (e.g., Chen et al., 2004; Li et al., 2012a; Izquierdo-
211 Llavall et al., 2018; Neng et al., 2018). From Miocene to Quaternary, the gravel content in the
212 sediments progressively increased along the margins of the Tarim until the deposition of the
213 conglomeratic Xiyu Formation, while sandstones and siltstones still deposited in the centre of the
214 basin (Fig. 2) (e.g., Yin et al., 1998, 2002; Jin et al., 2003; Sun et al., 2005, 2009b, 2017;
215 Heermance et al., 2007; Jin et al., 2008; Wang et al., 2009; Chang et al., 2012a; Thompson et al.,
216 2015; Zheng et al., 2015; Tian et al., 2016). In compressive contexts, this kind of long-term
217 coarsening-upward sequence is typically controlled by the growth and erosion of the mountain

218 ranges and their progressive thrusting over the adjacent basins (e.g., Puigdefabregas et al., 1986;
219 Sinclair and Allen, 1992; Simoes and Avouac, 2006; Heermance et al., 2007; Charreau et al.,
220 2009; Dubille and Lavé, 2015).

221

222 2.2. *Western Kunlun Range*

223

224 The Western Kunlun Range bounds the Tarim Basin to the southwest, between the towns
225 of Kashgar and Hotan (Fig.1). It is a ~700-km-long, ~100-130-km-wide and up to ~7600-m-high
226 mountain belt at the northwestern edge of the Tibetan Plateau. Nowadays, this mountain belt
227 corresponds to a northeastward-vergent thrust wedge that developed between the Kongur Shan
228 normal fault and the Karakax left-lateral fault to the west and south, and the Tarim Basin to the
229 northeast (Fig.1) (e.g., Brunel et al., 1994; Matte et al., 1996; Wittlinger et al., 2004; Lu et al.,
230 2016). From its overall orientation, this range can be subdivided into two parts: (1) a NNW-SSE-
231 oriented part from Kashgar to Yecheng to the west, showing a structural and stratigraphic
232 continuity with the northeastern Pamir, and (2) a WNW-ESE-oriented part from Yecheng to
233 Hotan to the east.

234 Within this mountain belt, the southwestern boundary of the Tarim block is marked by
235 the Oytag-Kudi suture (in red in Fig.1) related to the closure of the Proto-Tethys Ocean and to
236 the associated Paleozoic orogeny spanning from Ordovician to Devonian (e.g., Matte et al.,
237 1996; Mattern and Schneider, 2000; Xiao et al., 2005; Li et al., 2018a). During the Mesozoic, a
238 complex strike-slip to compressive history then followed this orogeny (e.g., Sobel, 1999; Arnaud
239 et al., 2003; Cao et al., 2015). Finally, a major reactivation of the Western Kunlun Range
240 occurred during the Cenozoic in response to the India–Asia collision (e.g., Tapponnier and
241 Molnar, 1977; Matte et al., 1996; Sobel and Dumitru, 1997; Yin et al., 2002; Cao et al., 2015).
242 Based on thermochronological data (e.g., Sobel et al., 1997; Cao et al., 2013, 2015; Cheng et al.,
243 2017; Blayney et al., 2016; Yang et al., 2018a), sediment provenance and growth strata (e.g., Yin
244 et al., 2002; Bershaw et al., 2012; Jiang et al., 2013; Cao et al., 2014; Thompson et al., 2015;
245 Cheng et al., 2016b, 2017; Blayney et al., 2016; Sun et al., 2016b), as well as increasing
246 sediment accommodation and accumulation rates in the southwestern Tarim Basin (Métivier and
247 Gaudemer, 1997; Metivier et al., 1999; Yang and Liu, 2002; Jiang and Li, 2014; Blayney et al.,
248 2019), this reactivation seems to initiate by Eocene with subsequent Oligocene to Miocene

249 accelerations. The present thrust wedge then formed above an intra-crustal decollement level that
250 progressively shallows from the south to the north. Indeed, the interior of the mountain range
251 results from the imbrication and exhumation of Protero-Paleozoic basement and Paleozoic cover,
252 whereas the foothills are characterized by a fold-and-thrust belt of Paleozoic, Mesozoic and
253 Cenozoic sediments (e.g., XBGMR, 1976, 1985; Brunel et al., 1994; Matte et al., 1996;
254 Wittlinger et al., 2004; Li et al., 2012b, 2016b; Jiang et al., 2013; Cheng et al., 2016b; Wang and
255 Wang, 2016; Lu et al., 2016; Guilbaud et al., 2017; Thompson et al., 2017).

256 The basement of the Western Kunlun is thus composed of Proterozoic to Paleozoic
257 (Ordovician to Silurian) plutonic and metamorphic rocks that recorded the lower Paleozoic
258 orogeny of the range (e.g., Matte et al., 1996; Mattern and Schneider, 2000; Cowgill et al., 2003;
259 Cowgill, 2010; Cao et al., 2013), while within the Tarim Basin, the contemporaneous
260 sedimentary series have not endured such plutonic and metamorphic events (see section 2.1)
261 (Fig. 2). Over this basement, the Western Kunlun Paleozoic cover consists in Devonian
262 continental conglomerates coming from the erosion of the Paleozoic mountain belt, followed by
263 Carboniferous marine limestones (Fig. 2) (e.g., Matte et al., 1996; Mattern and Schneider, 2000;
264 Cowgill, 2010; Cao et al., 2013). Within the foothills of the range, the stratigraphy above these
265 units then corresponds to a Permian to Cenozoic, mostly continental and terrigenous sedimentary
266 series, containing an upper Cretaceous-Paleogene marine interval of terrigenous, calcareous, and
267 evaporitic deposits (Fig. 2) (e.g., Matte et al., 1996; Sobel, 1999; Yin et al., 2002; Jin et al.,
268 2003; Sun and Jiang, 2013; Bosboom et al., 2014a, 2014b, 2017; Guo et al., 2015; Zheng et al.,
269 2015; Chen et al., 2018). Although coarser-grained, this series is similar to the contemporaneous
270 sediments deposited at the front of the Western Kunlun Range, along the southwestern Tarim
271 Basin margin.

272

273 2.3. *Altyn Tagh Range*

274

275 The Altyn Tagh Range bounds the Tarim Basin to the southeast, from the west of
276 Minfeng to the east of Dunhuang (Fig. 1). It is a ~1500-km-long, ~40-80-km-wide and up to
277 ~6200-m-high, WSW-ENE oriented mountain belt at the centre of the Tibetan Plateau northern
278 edge. Presently, this mountain range corresponds to a northwestward-vergent strike-slip wedge
279 that developed between the Altyn Tagh left-lateral fault to the southeast and the Tarim Basin to

280 the northwest (Fig. 1) (e.g., Wittlinger et al., 1998; Cowgill et al., 2000; Li et al., 2002a; Yin et
281 al., 2002; Tang et al., 2014; Lin et al., 2015b; Zhao et al., 2016).

282 In this mountain belt, the southeastern boundary of the Tarim block is marked by the
283 Lapeiquan-Hongliugou suture (in red in Fig. 1) related to the closure of the Proto-Tethys Ocean
284 and to the associated Paleozoic orogeny spanning from Silurian to Devonian (e.g., Sobel and
285 Arnaud, 1999; Gehrels et al., 2003; Li et al., 2018a). After a complex deformation history during
286 the Mesozoic (e.g., Sobel, 2001; Chen et al., 2003; Robinson et al., 2003; Liu et al., 2007), the
287 Altyn Tagh Range was then reactivated during the Cenozoic due to the India–Asia collision (e.g.,
288 Tapponnier and Molnar, 1977; Yin et al., 2002; Jolivet et al., 2001; Yin, 2010). Based on
289 thermochronological data (e.g., Jolivet et al., 1999, 2001; Sobel, 2001; Yin et al., 2002; Liu et
290 al., 2003, 2007; Ritts et al., 2008), sediment provenance (e.g., Yue et al., 2001; Yin et al., 2002;
291 Sun et al., 2005; Wu et al., 2012; Li et al., 2014b; Lin et al., 2015b; Lu et al., 2014; Cheng et al.,
292 2015, 2016a), growth strata (e.g., Cheng et al., 2015, 2016a; Zhao et al., 2016), and accumulation
293 rates (e.g., Métivier et al., 1998; Sun et al., 2005; Lu et al., 2014), this reactivation seems to
294 initiate by Eocene with subsequent Oligocene to Miocene accelerations. The present strike-slip
295 wedge then formed, presumably, above a deep intra-crustal decollement level. Indeed, the range
296 interior results from the imbrication and exhumation of Protero-Paleozoic basement and Paleo-
297 Mesozoic cover, while in the foothills, compressive structures involve the same basement and
298 cover units overlain by Cenozoic sediments (e.g., XBGMR, 1976, 1985; Jolivet et al., 1999;
299 Chen et al., 2003; Liu et al., 2007; Tang et al., 2014; Lin et al., 2015b).

300 The basement of the Altyn Tagh is thus composed of Proterozoic to Paleozoic (Silurian to
301 Devonian) plutonic and metamorphic rocks that recorded the middle Paleozoic orogeny of this
302 mountain belt (e.g., Sobel and Arnaud, 1999; Cowgill et al., 2003; Gehrels et al., 2003; Zhang et
303 al., 2014a), whereas the contemporaneous Tarim sedimentary series have not endured such a
304 metamorphic event (see section 2.1), similar to what is observed at the front of the Western
305 Kunlun Range. Over this basement, the Altyn Tagh Paleo-Mesozoic cover consists in
306 Carboniferous marine limestones and Jurassic continental and terrigenous deposits (e.g., Sobel
307 and Arnaud, 1999; Chen et al., 2003). Within the foothills of the range, the stratigraphy above
308 these units corresponds to a continental and terrigenous Cenozoic series in continuity with the
309 contemporaneous sediments deposited in the Tarim Basin (e.g., Yin et al., 2002; Sun et al., 2005;
310 Lu et al., 2014).

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2.4. Southern Tian Shan

The Southern Tian Shan bounds the Tarim Basin to the north, between the towns of Kashgar and Dunhuang (Fig. 1). It is a ~2200-km-long, ~100-250-km-wide and up to ~7400-m-high mountain belt corresponding to the southern edge of the Tian Shan Range. From its overall orientation, this belt can be subdivided into two parts: (1) the WSW-ENE-oriented Southwestern Tian Shan from Kashgar to the east of Kuqa, and (2) the WNW-ESE-oriented Southeastern Tian Shan from the east of Kuqa to Dunhuang. Presently, the Southwestern Tian Shan consists in a southeastward-vergent thrust wedge developed between the Nikolaev Line and Narat Fault to the northwest and the Tarim Basin to the southeast (e.g., Poupinet et al., 2002; Heermance et al., 2008; Makarov et al., 2010; Xiao et al., 2013; Alexeiev et al., 2015; Izquierdo-Llavall et al., 2018), whereas the Southeastern Tian Shan corresponds to a southwestward-vergent strike-slip wedge developed between the right-lateral Baluntay Fault to the northeast and the Tarim Basin to the southwest (e.g., Allen et al., 1994; Charvet et al., 2011; Li et al., 2016a) (Fig. 1).

In these mountain belts, the northern boundary of the Tarim block is marked by the South Tian Shan suture (in red in Fig. 1) related to the closure of the South Tian Shan Ocean or back-arc basin and to the associated Paleozoic orogeny spanning from Devonian to Permian (e.g., Windley et al., 1990; Gao et al., 1998; Charvet et al., 2007, 2011; Glorie et al., 2011; Xiao et al., 2013; Alexeiev et al., 2015; Han et al., 2016). During the Mesozoic, a complex compressive to strike-slip history then followed this orogeny (e.g., Hendrix et al., 1992; Sobel, 1999; Dumitru et al., 2001; Wang et al., 2009; Jolivet et al., 2010, 2013; Glorie et al., 2011; Chang et al., 2012b; Li et al., 2015a; Morin et al., 2018; Yang et al., 2018b). Finally, a major reactivation of the Tian Shan occurred during the Cenozoic in response to the India–Asia collision (e.g., Tapponnier and Molnar, 1979; Windley et al., 1990; Avouac et al., 1993; Allen et al., 1994; Hendrix et al., 1994; De Grave et al., 2007; Dumitru et al., 2001; Jolivet et al., 2010). Based on thermochronological data (e.g., Hendrix et al., 1994; Sobel and Dumitru, 1997; Dumitru et al., 2001; Sobel et al., 2006; Jolivet et al., 2010; Glorie et al., 2011; Chang et al., 2012b; Yu et al., 2014; Jia et al., 2015), sediment provenance and growth strata (e.g., Yin et al., 1998; Heermance et al., 2007; Li et al., 2012a; Bande et al., 2017a; Izquierdo-Llavall et al., 2018; Thompson et al., 2018; Li et al., 2019), as well as increasing sediment accommodation and accumulation rates in the northern

342 Tarim Basin (e.g., [Métivier and Gaudemer, 1997](#); [Métivier et al., 1999](#); [Yang and Liu, 2002](#);
343 [Huang et al., 2006](#); [Heermance et al., 2007](#); [Zhang et al., 2014b](#); [Li et al., 2019](#)), this reactivation
344 seems to initiate by Oligocene with subsequent Miocene accelerations.

345 Following this reactivation, the Southwestern Tian Shan thrust wedge formed above an
346 intra-crustal decollement level that progressively shallows from the north to the south. The
347 interior of this range thus results from the imbrication and exhumation of Protero-Paleozoic
348 basement and Paleo-Mesozoic cover (e.g., [XBGMR, 1976, 1985](#); [Yin et al., 1998](#); [Xiao et al.,](#)
349 [2013](#); [Alexeiev et al., 2015](#); [Jourdon et al., 2017](#); [Loury et al., 2017](#)). It also comprises a few
350 compressive intra-mountainous basins filled by Cenozoic sediments (e.g., [XBGMR, 1976, 1985](#);
351 [Xiao et al., 2013](#); [Morin et al., 2019](#)). The foothills of the Southwestern Tian Shan then present a
352 deformation style that strongly varies laterally. In the centre, near the town of Aksu, compressive
353 structures involving Proterozoic basement and Paleo-Cenozoic cover developed from a large
354 basement uplift (the Kalpin uplift) inherited from the Protero-Paleozoic history of the Tarim
355 block (e.g., [XBGMR, 1976, 1985](#); [Carroll et al., 2001](#); [Turner et al., 2010](#)). In contrast, to the
356 west and to the east, the foothills are characterized by fold-and-thrust belts (the Kashgar-Artux,
357 Kepingtage and Kuqa belts) involving Paleozoic, Mesozoic and Cenozoic sediments (e.g.,
358 [XBGMR, 1976, 1985](#); [Allen et al., 1999](#); [Yin et al., 1998](#); [Burchfield et al., 1999](#); [He et al., 2005](#);
359 [Heermance et al., 2008](#); [Turner et al., 2010](#); [Li et al., 2012a](#); [Alexeiev et al., 2015](#); [Li et al.,](#)
360 [2016d](#); [Izquierdo-Llavall et al., 2018](#)).

361 Further east, the Southeastern Tian Shan strike-slip wedge could have formed above an
362 intra-crustal or deeper decollement level. As in the Southwestern Tian Shan, the range interior
363 indeed results from the imbrication and exhumation of Protero-Paleozoic basement and Paleo-
364 Mesozoic cover, and it comprises a few compressive intra-mountainous basins filled by
365 Cenozoic sediments (e.g., [XBGMR, 1976, 1985](#); [Allen et al., 1994](#); [Charvet et al., 2011](#); [Li et al.,](#)
366 [2016a](#); [Morin et al., 2019](#)). The foothills then entirely correspond to compressive and strike-slip
367 structures involving Proterozoic basement and Paleo-Cenozoic cover, which derive from a large
368 inherited Protero-Paleozoic basement uplift of the Tarim (the Kuruk Tagh uplift) (e.g., [XBGMR,](#)
369 [1976, 1985](#); [Yin et al., 1998](#); [Tang et al., 2008](#); [Lin et al., 2012b](#)).

370 The basement of the Southern Tian Shan interior is thus composed of Proterozoic and
371 Paleozoic (Silurian to Devonian) metamorphic and plutonic rocks that recorded the middle-upper
372 Paleozoic orogeny of the range ([Fig. 2](#)) (e.g., [Gao et al., 1998](#); [Charvet et al., 2007, 2011](#); [Gao et](#)

373 al., 2009; Ge et al., 2012; Xiao et al., 2013; Alexeiev et al., 2015). Over this basement, the
374 Paleozoic cover consists in Silurian-Devonian to Permian marine limestones and continental to
375 marine, volcano-sedimentary or terrigenous deposits (Fig. 2) (e.g., Gao et al., 1998; Charvet et
376 al., 2011; Liu et al., 2013; Xiao et al., 2013; Alexeiev et al., 2015; Li et al., 2015a; Jourdon et al.,
377 2017). Both the basement and cover units are intruded by Paleozoic (Devonian to Permian)
378 plutons (Fig. 2) (e.g., Gao et al., 1998; Charvet et al., 2007, 2011; Gao et al., 2009; Glorie et al.,
379 2011; Xiao et al., 2013; Alexeiev et al., 2015). However, in the foothills of the range as in the
380 Tarim Basin, the coeval stratigraphy is characterized by a Proterozoic igneous basement below a
381 thick, more or less continuous series of Sinian to Permian sediments that have not endured the
382 Paleozoic orogeny (e.g., Carroll et al., 1995, 2001; Zhou et al., 2001; Tang et al., 2008; Turner et
383 al., 2010; Shu et al., 2011; Li et al., 2015a; Cai et al., 2018; Wu et al., 2018). Above all these
384 units, the Southern Tian Shan cover is composed of Triassic to Cenozoic, mostly continental and
385 terrigenous sediments containing a Paleogene lacustrine interval of terrigenous and evaporitic
386 deposits (Fig. 2) (e.g., Hendrix et al., 1992; Yin et al., 1998; Heermance et al., 2007; Tang et al.,
387 2008; Wang et al., 2009; Li et al., 2015a; Li et al., 2016d; Tian et al., 2016; Izquierdo-Llavall et
388 al., 2018; Morin et al., 2018). Although coarser-grained, this series is similar to the
389 contemporaneous sediments deposited in the basin at the front of the range.

390

391 **3. Data and method**

392

393 *3.1. Data*

394

395 To characterize the Cenozoic deformation of the Tarim Basin, we have constructed a
396 structural map and four balanced regional geological transects. To construct this map and these
397 transects, we first synthesized the studies that were previously published in the area (Fig. 1).
398 Indeed, numerous data and interpretations were already available such as (1) regional and local
399 geological maps (e.g., XBGMR, 1976, 1985; Matte et al., 1996; Burchfield et al., 1999; He et al.,
400 2005; Cowgill, 2010; Fu et al., 2010; Turner et al., 2010; Cao et al., 2013; Cheng et al., 2015;
401 Wang and Wang, 2016; Guilbaud et al., 2017; Thompson et al., 2017), (2) geological cross-
402 sections from field observations and seismic profiles (e.g., Fig. 1 and references therein), and (3)
403 isopach and isobath maps (e.g., Lee, 1985; Li et al., 1996; Yang and Liu, 2002; Chen and Shi,

404 2003; Li et al., 2016e). Eventually, we also used the previous biostratigraphical and
405 magnetostratigraphical studies in the region to constrain the ages of the sedimentary series (e.g.,
406 Yin et al., 1998, 2002; Sobel, 1999; Sun et al., 1999; Chen and Shi, 2003; Sun et al., 2005,
407 2009a, 2009b, 2016a, 2017; Charreau et al., 2006, 2009; Huang et al., 2006, 2010; Heermance et
408 al., 2007; Wang et al., 2009; Chang et al., 2012a, 2014a; Bosboom et al., 2014a, 2014b, 2017; Lu
409 et al., 2014; Zhang et al., 2014b; Zheng et al., 2015; Peng et al., 2018; Blayney et al., 2019).

410 In addition to these previously-published data, we used surface observations performed in
411 the field and on satellite images (Landsat, Spot and DigitalGlobe images available through the
412 Google Earth database), as well as on digital elevation models (SRTM30, SRTM3, SRTM1,
413 ASTER GDEM v2, and AW3D30, with a horizontal resolution from ~900 to ~30 m) (e.g., Figs.
414 4 and 5). Thanks to the Sinopec Group of the China Petrochemical Corporation, we also had
415 access to an extensive subsurface dataset comprising ~31120 km of regional and local seismic
416 profiles covering the entire basin (Figs. 1 and S2), together with the data of 29 wells. In
417 particular, Vertical Seismic Profiles (VSP) of P-waves were available for four of these boreholes
418 (red dots in Fig. 1). From these VSP, seismic velocities can be deduced and a velocity model can
419 be proposed (Fig. 3), in order to convert the travel time of seismic interpretations into depth (see
420 section 3.4).

421

422 3.2. Seismic interpretation

423

424 To interpret our seismic dataset, we followed a standard procedure (Nester and Padgett,
425 1992). First, we identified the faults based on offsets of the seismic reflectors. Then, we picked
426 seven horizons along the seismic profiles through the whole seismic grid covering the basin: the
427 base of the (1) Sinian (i.e. the basement-sedimentary cover boundary in the Tarim Basin), (2)
428 upper Cambrian (i.e. the base of the main Paleozoic decollement in the region), (3) Permian, (4)
429 Paleocene, (5) Oligocene, (6) middle Miocene, and (7) Pliocene deposits (Fig. 2). The ages of the
430 pre-Cenozoic horizons were mainly constrained by biostratigraphical dating on outcrops and
431 cores (e.g., Sobel, 1999; Chen and Shi, 2003; Zhang et al., 2014b; Peng et al., 2018), as well as
432 on samples collected in the boreholes of the Sinopec Group dataset. The ages of the Cenozoic
433 horizons were determined on the basis of previous biostratigraphical and magnetostratigraphical
434 studies mainly performed on outcrops. For this, we took advantage of sections sampled along

435 outcrops that can be correlated with the seismic data (Yin et al., 2002; Charreau et al., 2006,
436 2009; Huang et al., 2006, 2010; Peng et al., 2006; Sun et al., 2009a, 2009b, 2017; Wang et al.,
437 2009; Bosboom et al., 2014a; Zheng et al., 2015; Blayney et al., 2019). These sections are
438 located to the southwest, in the centre, and to the north of the basin (green dots in Fig. 1).

439 In order to ensure that our interpretation is consistent over the whole studied area, the
440 selected seismic horizons were propagated from a seismic line to another through the entire
441 seismic grid. The robustness of this horizon picking is confirmed by the fact that the position of
442 our Cenozoic seismic horizons fall within a time interval of less than 0.5 Myr on the dated
443 sections outcropping several 100s of km apart in different places of the basin.

444

445 *3.3. Structural mapping*

446

447 In order to highlight the structures activated during the Cenozoic, we constructed a
448 georeferenced structural map of the Cenozoic deformation through the Tarim Basin and its
449 surrounding ranges (Figs. 6 and S1). On this map, we reported the main faults and folds affecting
450 the Cenozoic sedimentary series. We identified and located the outcropping structures from the
451 available surface data (i.e. the field constraints, satellite images, digital elevation models,
452 together with the previously-published geological maps and cross-sections) (e.g., Figs. 4 and 5).
453 When it was possible, we also verified the occurrence and location of these structures on the
454 seismic profiles. In addition, we identified and located the structures buried below the Cenozoic
455 sediments from the available subsurface data (i.e. the previously-published and unpublished
456 seismic profiles and wells) (e.g., Fig. S2). Finally, we reported the position of the outcropping
457 and buried fault tips and fold axes on the map, using different line thicknesses according to the
458 scale of the structures, and different colours according to their rooting depth (red for the
459 basement, orange for the Paleozoic décollements and yellow for the Cenozoic ones) (Fig. 6).

460

461 *3.4. Time-to-depth conversion*

462

463 To build geological cross-sections through the basin, we needed to convert our seismic
464 interpretations (in two-way-time) into geological geometries (in depth). For this, we used the
465 VSP data from four boreholes distributed across the basin (red dots in Fig. 1) to establish

466 velocity models for three sediment packages: the Sinian-Paleozoic (Sinian to Carboniferous),
467 Permian-Mesozoic (Permian to Cretaceous) and Cenozoic series (Fig. 3). Indeed, for these
468 different packages, the variation of the interval velocities according to depth follows a similar
469 trend in the four wells. For each series, we approximated this variation by a velocity function
470 versus depth with two parts: (1) an upper part corresponding to a single linear regression of the
471 velocity data (solid lines in Fig. 3), and (2) a lower part corresponding to a maximum velocity
472 determined according to the main lithology of the sediment package (calcareous for the Sinian-
473 Paleozoic deposits, terrigenous for the Permian-Mesozoic and Cenozoic series) (dashed lines in
474 Fig. 3). By using this velocity model, we assumed that the three approximated velocity functions
475 are uniform at the basin scale, i.e. we neglected the local velocity variations that can exist due to
476 lateral lithological changes. The time-to-depth conversion based on these functions is therefore a
477 first-order achievement, which seems to be acceptable for this study at the scale of the whole
478 Tarim Basin. Indeed, the depths estimated for the selected horizons falls in the range of their
479 depths assessed in previously-published structural interpretations (cross-sections, isobath and
480 isopach maps where the horizon depths vary by a few kilometres) (e.g., Li et al., 1996; Chen and
481 Shi, 2003; Tang et al., 2014; Li et al., 2016e; Lu et al., 2016; Wang and Wang, 2016; Izquierdo-
482 Llavall et al., 2018; Jiang et al., 2018).

483

484 3.5. Geological transect construction

485

486 In addition to the structural map, we constructed four balanced geological transects
487 through the basin and the outer parts of the ranges (Figs. 7, 8, 9, 10, and S3). We built these
488 regional cross-sections following a standard method (e.g., Woodward et al., 1989). We chose
489 their paths perpendicular to the Cenozoic structure direction, which may change along each
490 section (Fig. 6). As much as possible, we also located them along seismic profiles to precisely
491 constrain the deformation in the basin (Figs. 1 and S2). Where this was not possible (e.g., where
492 the sections are not parallel to the seismic data such as in the centre of the basin for example), the
493 deformation could however still be constrained using the seismic profiles crossing the transects
494 in multiple locations. As for the structural map, we used the surface geology (e.g., Figs. 4 and 5)
495 to build these transects where structures are cropping out, in particular through the ranges at the
496 ends of the sections. We also used the available subsurface data to constrain the structures buried

497 below the Cenozoic sediments in the basin (e.g., Fig. S2). Finally, to balance our geological
498 transects, we used the concepts of the typical area balancing approach (e.g., Hossack, 1979;
499 Mitra and Namson, 1989), combined with a line-length balancing (e.g., Dahlstrom, 1969) for the
500 Meso-Cenozoic stratigraphic boundary. The line-length and area balancing approaches refer to
501 the basic assumption that the strata lengths and cross-sectional areas do not substantially change
502 during deformation. The combination of these approaches allowed us to propose admissible and
503 viable cross-sections over their whole length (Elliott, 1983), even where the details of the
504 structural geometries are not well imaged on seismic profiles (e.g., inside the duplex structures to
505 the south of transects A and B or to the north of transect C). This implies that the produced cross-
506 sections are restorable.

507

508 *3.6. Transect restoration*

509

510 In order to quantify the total Cenozoic shortening accommodated within the basin and the
511 outer parts of the ranges, we restored the proposed balanced geological transects at the beginning
512 of the Cenozoic (Fig. 7). This restoration was thus mainly based on the length conservation of
513 the Paleogene basal boundary, which pre-dates the Cenozoic deformation onset. Where this
514 boundary is lost by erosion without any older stratigraphic horizon that may be used as another
515 marker, we minimized the eroded length by closing the structures just above the present-day
516 topography (e.g., see the dotted lines to the south of transects C and D or to the north of transect
517 A). Moreover, where this boundary is missing in the stratigraphy (e.g., where the lower Tertiary
518 is lacking due to no deposit or erosion at that time), the restoration was made at the base of the
519 oldest Cenozoic sediments existing in the area (e.g., the Middle-Upper Miocene to the north of
520 transect A or the Plio-Quaternary to the north of transects B and D). Therefore, by following
521 those rules, the shortening amounts deduced from our structural restoration are minimum
522 estimates. However, to avoid too much approximation, we only performed this restoration
523 between the main bordering faults of the Cenozoic deposits, without including the range interiors
524 where stratigraphic markers are lacking. Eventually, the fixed reference lines used for this work
525 are located in the basin, at the turning points of the different transects (Figs. 1 and 7).

526

527 **4. Overall Cenozoic deformation**

528

529 The structural map and regional geological transects constructed to document the
530 Cenozoic deformation through the Tarim area are presented in **Figures 6 and 7**. They illustrate
531 the overall structure of the surrounding mountain ranges verging toward the basin, i.e. northward
532 along the Tibetan Plateau edge to the south and southward along the Tian Shan to the north.
533 These surrounding ranges correspond to well-developed thrust wedges to the west of the basin
534 (Western Kunlun Range and Southwestern Tian Shan) and strike-slip wedges to the east (Altyn
535 Tagh and Southeastern Tian Shan). They concentrate a large amount of deformation with thick-
536 skinned structures involving a Protero-Paleozoic basement and Paleozoic sedimentary cover in
537 their interiors, while their foothills are characterized by either thin- or thick-skinned structures
538 involving a Protero-Paleozoic basement and/or Protero-Paleozoic, Mesozoic and Cenozoic
539 sediments, respectively (**Figs. 6 and 7**).

540 In the case of the Western Kunlun and Southwestern Tian Shan compressive systems,
541 these foothills are mainly formed of wide (10s of km) and long (100s of km) thin-skinned fold-
542 and-thrust belts (Western Kunlun, Kashgar-Artux, Kepingtage, and Kuqa belts), except for the
543 central part of the Southwestern Tian Shan where a thick-skinned belt locally occurs (Kalpin
544 belt) (**Figs. 6 and 7**). Associated with these fold-and-thrust belts, which can accommodate a large
545 amount of shortening (10s of km) and thickening, deep flexures (Southwest, North and Kuqa
546 depressions) have developed in the basin with up to ~10 and ~9 km of Cenozoic sediments
547 accumulated along the front of the Western Kunlun and Southwestern Tian Shan, respectively
548 (**Fig. 7A, 7B and 7C**). In contrast, no fold-and-thrust belt and deep foreland flexure are
549 associated with the Altyn Tagh and Southeastern Tian Shan transpressional systems (**Figs. 7C**
550 **and 7D**). In their case, the foothills are narrow (a few 10s of km) and correspond to a few
551 basement-involved structures accommodating a limited amount of shortening (a few kilometres
552 at most). Along the front of these ranges, less than a few kilometres of Cenozoic sediments are
553 usually trapped.

554 In addition to this deformation within the ranges, Cenozoic deformation also occurs
555 within the Tarim. Wide (from ~50 to ~150 km) and long (100s of km) basement-cored uplifts are
556 visible to the west (Bachu uplift) and south (Tanar uplift) of the basin (**Figs. 6, 7B, and 7C**).
557 These basement-cored uplifts correspond to large pop-up structures bounded by conjugate thrust
558 systems. They are clearly parallel to the ranges bordering the basin to the south (**Fig. 6**), with a

559 main vergence toward the north, i.e. synthetic to the deformation of the Tibetan Plateau edge.
560 Indeed, this vergence can be deduced from the observation of a compressive deformation
561 accumulated on their conjugate bounding thrust systems larger to the north than to the south.
562 Besides these basement-cored uplifts, other basement-involved structures (Tazhong, Tadong, and
563 Tabei uplifts) exist further north and east in the basin (Figs. 6, 7C and 7D). However, they
564 mainly deform the Protero-Paleozoic strata, with little evidence of a significant Cenozoic activity
565 on their bounding faults. At first order, the Cenozoic deformation of the original Proterozoic
566 Tarim block therefore mainly occurs along its edges and within its southwestern interior.

567

568 **5. Southwest deformation**

569

570 To the southwest of the Tarim Basin, the northeastward-vergent Western Kunlun thrust
571 system can be divided into four structural parts with, from the southwest to the northeast: (1) the
572 inner range from the Kongur Shan normal fault and Karakax left-lateral fault to the Main Pamir
573 thrust, Kashgar-Yecheng transfer system, and Tiklik thrust system, (2) the outer range from the
574 Main Pamir thrust, Kashgar-Yecheng transfer system and Tiklik thrust system to the Western
575 Kunlun frontal blind ramp, (3) the Mazar Tagh thrust sheet and Southwest depression from the
576 Western Kunlun frontal blind ramp to the Mazar Tagh emergent thrust ramps, and (4) the Bachu
577 uplift from the Selibuya-Mazar Tagh deep fault systems to the Aqia-Tumuxiuke thrust systems
578 (Figs. 6, 7A, 7B, 8A, 8B and 9).

579

580 *5.1. Western Kunlun inner range*

581

582 The inner range is several 10s to ~100 km wide, with typical elevations between ~2500
583 and ~5500 m. It is composed of exhumed imbricates of the Western Kunlun Protero-Paleozoic
584 (Sinian to Silurian) basement and Paleozoic (Devonian to Carboniferous) cover (see section 2.2).
585 In this system, the inner range is a continuous belt stretching from the Pamir to the west to the
586 Altyn Tagh to the east. There, its northern part ends in a large basement-involved anticline and
587 syncline to the southeast of Hotan (Fig. 6). Due to limited field access and subsurface data, its
588 detailed structure is poorly constrained, but connecting the major faults within the basement to a
589 deep basal decollement seems a reasonable hypothesis (Figs. 7A, 7B, 8A and 8B), as previously

590 proposed by [Matte et al., 1996](#), [Wittlinger et al., 2004](#) or [Guilbaud et al., 2017](#). Towards the
591 northeast, the range interior is separated from the outer range by the Main Pamir thrust to the
592 northwest and by the Tiklik thrust system to the southeast, connected by the Kashgar-Yecheng
593 transfer system along the eastern edge of the Pamir salient ([Fig. 6](#)). These emergent thrust and
594 strike-slip zones also correspond to the main bordering faults of the southwestern Cenozoic
595 series of the Tarim Basin.

596

597 *5.2. Western Kunlun outer range*

598

599 At the front of the inner range, the outer range is also several 10s to ~100 km wide, with
600 typical elevations between ~1500 and ~2500 m. It consists in a fold-and-thrust belt that mainly
601 developed below the Cenozoic sediments of the Tarim Basin ([Figs. 6, 7A, 7B, 8A, 8B, S2A and](#)
602 [S2B](#)). This belt follows the range from the Pamir to the west to the southeast of Hotan to the east
603 where it abruptly ends into a lateral thrust ramp ([Fig. 6](#)). Several cross-sections have already
604 been proposed through this outer range with various and contradictory interpretations on its
605 structure (e.g., [Brunel et al., 1994](#); [Matte et al., 1996](#); [Li et al., 2012b, 2016b](#); [Jiang et al., 2013](#);
606 [Wang et al., 2013](#); [Wei et al., 2013](#); [Tang et al., 2014](#); [Cheng et al., 2016b, 2017](#); [Lu et al., 2016](#);
607 [Wang and Wang, 2016](#); [Wang et al., 2016a](#); [Guilbaud et al., 2017](#); [Thompson et al., 2017](#)). From
608 our surface and subsurface data, we derived a regionally-consistent structural model settling the
609 issue of these interpretations (see [Supplementary information S4](#) for further details and
610 discussion). We concluded that the surface and subsurface data provide clear evidence for 10s-
611 of-km-long, emergent or blind folded thrust slices made of the Western Kunlun Protero-
612 Paleozoic (Sinian to Silurian) basement and its Paleo-Mesozoic (Devonian to Cretaceous) cover
613 at the rear, whereas 10s- to 100s-of-km-long duplex structures developed within the Tarim
614 Paleo-Mesozoic (Cambrian to Cretaceous) sedimentary cover at the front ([Figs. 7A, 7B, 8A, 8B,](#)
615 [S2A and S2B](#)).

616 These structures imply a basal decollement level that progressively shallows from the
617 southwest to the northeast. At the rear of the outer belt, the major thrusts necessarily root into a
618 deep decollement within the basement since they exhumed basement units ([Figs. 7A, 7B, 8A and](#)
619 [8B](#)). This decollement probably connects with the one below the range interior as classically
620 inferred or proposed for most compressional orogens (e.g., [Muñoz, 1992](#); [Burkhard and](#)

621 Sommaruga; 1998, Lavé and Avouac, 2001). The northeastward change from the thick-skinned
622 thrust slices to the thin-skinned duplexes then requires the presence of a ramp connecting this
623 deep decollement to an intermediate one at the base of the deformed sedimentary cover (Figs.
624 7A, 7B, 8A, and 8B) (see also Matte et al., 1996; Wittlinger et al., 2004; Wang et al., 2013; Li et
625 al., 2016b, 2018b; Lu et al., 2016; Cheng et al., 2017; Guilbaud et al., 2017). According to our
626 regional seismic interpretation constraints by local borehole observations, this intermediate
627 decollement is located at a depth of about 10 to 13 km below sea level (b.s.l.), in the upper
628 Cambrian evaporitic assemblage of the Tarim sedimentary series (see also Wang et al., 2013;
629 Tang et al., 2014; Lu et al., 2016; Guilbaud et al., 2017). Remarkably, the stratigraphy of the
630 Paleozoic terrains is different on both sides of the ramp connecting the deep and intermediate
631 decollements (Figs. 7A, 7B, 8A and 8B). To the southwest, this stratigraphy is characteristic of
632 the Paleozoic Western Kunlun Range with Ordovician to Silurian metamorphic rocks covered by
633 Devonian to Permian deposits (see section 2.2), whereas to the northeast, it is typical of the
634 Paleozoic Tarim Basin with Cambrian to Permian sediments (see section 2.1) (Fig. 2). Therefore,
635 the deep ramp to the southwest of the outer range seems to be a major inherited Paleozoic
636 structure reactivated during the Cenozoic.

637 To the front of the outer belt, the formation of duplexes within the Tarim Paleo-Mesozoic
638 series requires the activation of a series of ramps connecting the intermediate decollement within
639 the upper Cambrian evaporitic assemblage to a shallower one (Figs. 7A, 7B, 8A and 8B).
640 According to the seismic profiles, together with field and well constraints, this shallow
641 decollement is located at a depth of about 6 to 9 km b.s.l. at the front of the duplexes, in the
642 Paleogene evaporitic assemblage at the base of the Cenozoic deposits (see also Li et al., 2012b,
643 2016b; Jiang et al., 2013; Wang et al., 2013; Tang et al., 2014; Cheng et al., 2017; Lu et al.,
644 2016; Wang and Wang, 2016; Guilbaud et al., 2017; Thompson et al., 2017). Below this
645 decollement, the stratigraphy of the Paleo-Mesozoic sedimentary cover seems to be the same
646 (Cambrian to Cretaceous deposits) across the ramps connecting the Paleozoic and Cenozoic
647 decollements. Therefore, these shallow ramps at the front of the outer range likely correspond to
648 structures created during the Cenozoic.

649 At the top of the blind basement-involved thrust slices and of duplexes of the outer belt,
650 the Cenozoic sedimentary series is folded into 10s-of-km-wide and up to ~180-km-long
651 anticlines and synclines (Figs. 6, 7A, 7B, 8A and 8B). Assuming an area conservation during

652 deformation to keep the cross-sections balanced, the geometry of the underlying duplexes can be
653 reconstructed from the overall envelope of these folds, which is usually well visible on the
654 seismic profiles (Fig. S2A and S2B). These structures are thus interpreted as a stack of several
655 (from four to six) horses, comprised between a floor thrust within the Paleozoic decollement
656 level and a roof thrust within the Cenozoic one (Figs. 7A, 7B, 8A and 8B). This duplex style is
657 typical of the frontal deformation in the central part of the Western Kunlun outer range, such as
658 for the Qimugen fold-thrust system, the Yecheng-Pishan anticline or the Pusikai anticline (see
659 also Jiang et al., 2013; Wang et al., 2013; Wei et al., 2013; Tang et al., 2014; Li et al., 2016b; Lu
660 et al., 2016; Wang and Wang, 2016; Guilbaud et al., 2017).

661 Near the lateral terminations of the belt, structures of the outer range vary slightly from
662 this pattern. To the northwestern termination (Yingisar anticline and Wupoer fold-thrust system),
663 duplexes are also described in the Paleo-Mesozoic cover, but they are associated with emergent
664 ramps breaching their roof thrust or rooted in the Cenozoic decollement at their top (e.g., Li et
665 al., 2012b, 2013; Thompson et al., 2015, 2017, 2018; Cheng et al., 2016b; Wang and Wang,
666 2016; Wang et al., 2016a). To the southeastern termination, the Hotan anticline corresponds to a
667 ramp anticline involving the Western Kunlun basement thrust over the Cenozoic decollement
668 (e.g., Jiang et al., 2013; Cheng et al., 2017). In the centre of the fold, this thrust sheet is
669 subsequently folded by a duplex structure that developed in the footwall of the basement ramp,
670 within the Tarim Pre-Cenozoic sedimentary cover. However, in spite of these geometrical
671 variations, the main structural features of the outer range are similar almost everywhere: (1) a
672 deep ramp rooted in the basement to the south, at the transition between the Western Kunlun and
673 Tarim Paleozoic domains, and (2) a duplex structure in the Paleo-Mesozoic Tarim sedimentary
674 series to the north, with an emergent or blind leading shallow ramp. The only exception
675 corresponds to the lateral ramp of the Aqike anticline at the eastern edge of the belt (Fig. 6).
676 Indeed, this anticline corresponds to a basement-involved breached duplex with a trailing
677 emergent ramp and no cover duplex at its front (Yang et al., 2009).

678 Although largely covered by the Cenozoic deposits coming from the Western Kunlun
679 erosion, the duplexes of the outer belt localize the topographic front of the mountain range. They
680 usually form smooth topographic features uplifted by several 100s of m above the present-day
681 Tarim sedimentation level (Fig. 6). In the northwestern and southeastern ends of the range, this
682 topographic front corresponds to emergent ramps, i.e. to an emergent deformation front.

683 However, along most of the range, the Western Kunlun frontal ramp is a blind structure since it
684 corresponds either to the leading blind ramp of a duplex (Qimugen fold-thrust system, Yecheng-
685 Pishan anticline or Pusikai anticline) or to the blind front of a ramp anticline (Hotan anticline)
686 (Figs. 7A, 7B, 8A and 8B). This raises the question of how the shortening accumulated within
687 the outer belt is accommodated and transferred upward to the Cenozoic sedimentary pile. In the
688 NNW-SSE-oriented part of the range, at the level of the Qimugen fold-thrust system, there is no
689 evidence of deformation, and thus of slip transfer, in the Cenozoic deposits of the foreland,
690 which are still attached to the underlying series (Figs. 6, 7A, 8A and S2A). This implies the
691 existence of a triangular zone at the front of this fold-thrust system, with a passive roof thrust at
692 its top within the Cenozoic decollement. It means that the Cenozoic deposits are detached above
693 the underlying duplex, and that their shortening is accommodated by an emergent backthrust at
694 their base (see also Wang and Wang, 2016; Wei et al., 2013). In contrast, to the east, in the
695 WNW-ESE-oriented part of the range, such passive roof duplexes cannot exist due to the
696 basement-involved imbricates emerging at the rear of the outer belt, which render back slip into
697 the Cenozoic decollement impossible (Figs. 7B and 8B). Moreover, there is evidence of
698 deformation, and thus of slip transfer, within the Cenozoic deposits of the foreland, such as in the
699 case of the Jiede anticline or of the Mazar Tagh emergent thrust ramps located in the Tarim
700 Basin, to the north of the Western Kunlun topographic front (Figs. 6, S2B, 7B and 8B) (see for
701 example Wang et al., 2013; Wei et al., 2013; Wang et al., 2014a; Li et al., 2016b; Lu et al., 2016;
702 Guilbaud et al., 2017; Jiang et al., 2018). Accordingly, in the eastern part of the range, the
703 shortening accumulated within the outer belt appears to be transferred toward the foreland rather
704 than toward the hinterland, with the activation of the Cenozoic decollement within the basin.

705

706 *5.3. Mazar Tagh thrust sheet and Southwest depression*

707

708 The Mazar Tagh is a striking topographic feature of the western central part of the Tarim
709 Basin corresponding to an up to ~5-km-wide and ~220-km-long single to multiple ridge, raising
710 a few 100s of m above the sedimentation level of the basin (Figs. 5A, 5C and 5D). This ridge
711 stretches ~150-200 km to the northeast of the Western Kunlun topographic mountain front,
712 parallel to the eastern part of the range (Fig. 6). Several contradictory cross-sections have been
713 proposed through the Mazar Tagh area (e.g., Wittlinger et al., 2004; Yang et al., 2007a; Pan et

714 al., 2010; Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Yang et al., 2015; Guo et al.,
715 2016; Li et al., 2016b; Li et al., 2016e; Lu et al., 2016; Guilbaud et al., 2017; Xie et al., 2017;
716 Jiang et al., 2018). From our surface and subsurface data, we derived a regionally-coherent
717 structural model in agreement with some of these previous studies (see [Supplementary](#)
718 [information S4](#) for further details and discussion).

719 According to our surface and subsurface dataset, this topographic high is clearly
720 controlled by northeastward-vergent emergent thrust ramps synthetic to the Western Kunlun
721 deformation and rooting at the base of the Cenozoic deposits ([Figs. 5B, 5D, 5E, 7B, 8B, 9 and](#)
722 [S2B](#)) (see also Yang et al., 2007a; Pan et al., 2010; Tong et al., 2012; Tang et al., 2014; Wang et
723 al., 2014a; Yang et al., 2015; Guo et al., 2016; Li et al., 2016b; Li et al., 2016e; Lu et al., 2016;
724 Guilbaud et al., 2017; Xie et al., 2017; Jiang et al., 2018). Indeed, the Paleogene evaporitic
725 assemblage acting as a major decollement level in the outer range extends northeastward into the
726 Tarim, at least until the Mazar Tagh where it outcrops at the surface ([Fig. 5D and 5E](#)). This
727 suggests that the slip transferred into this Cenozoic decollement at the front of the Yecheng-
728 Pishan, Pusikai, and Hotan anticlines of the outer belt is released at the Mazar Tagh emergent
729 thrust ramps, with erosion of the Cenozoic deposits of their hangingwalls. In agreement with
730 some of the previous studies ([Pan et al., 2010; Li et al., 2016e; Lu et al., 2016; Guilbaud et al.,](#)
731 [2017; Li et al., 2018b](#)), we thus argue that to the east of the central bend of the range, these
732 ramps correspond to the present-day emergent deformation front located in the basin, ~150-200
733 km away from the topographic mountain front, with a large frontal thrust sheet without any
734 major internal deformation.

735 From the surface and subsurface data, structures involving the Proterozoic basement and
736 Protero-Paleozoic to Mesozoic (Sinian to Jurassic) cover of the Tarim Basin also exist below the
737 shallow Mazar Tagh deformation ([Figs. 5A, 5B, S2B, 7B, 8B and 9](#)) (see also Yang et al., 2007a;
738 Pan et al., 2010; Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Guo et al., 2016; Li et
739 al., 2016b; Li et al., 2016e; Lu et al., 2016; Guilbaud et al., 2017; Xie et al., 2017; Jiang et al.,
740 2018). Despite a complex and sometimes poorly-imaged geometry, these structures appear as a
741 series of fault-related anticlines, often with backthrusts, or of minor pop-ups associated with
742 steep thrust ramps. Regardless of a certain degree of symmetry, their main vergence is clearly
743 toward the southwest since the base of the Cenozoic series is systematically higher to the
744 northeast than to the southwest of these structures ([Figs. 7B, 8B, 9 and S2B](#)) (see for example

745 Yang et al., 2007a; Pan et al., 2010; Tong et al., 2012; Wang et al., 2014a; Yang et al., 2015;
746 Guo et al., 2016; Li et al., 2016b, 2018b; Li et al., 2016e; Lu et al., 2016; Guilbaud et al., 2017;
747 Xie et al., 2017; Jiang et al., 2018). This unambiguously indicates an uplift of their northeastern
748 compartments due to a southwestward thrust movement on ramps antithetic to the Western
749 Kunlun deformation. At depth, the Mazar Tagh deep fault system needs to connect to a deep
750 decollement within the basement. At their tips, the Mazar Tagh emergent thrust ramps seem to
751 be correlated with south-facing steps in the Cenozoic decollement, indicating that the location of
752 these shallow ramps was probably controlled by the hangingwall uplift of the underlying deep
753 structures (Figs. 6, 7B, 8B, 9 and S2B).

754 To the west of the central bend of the range, the offset on these Mazar Tagh emergent
755 thrust ramps decreases so that these ramps progressively disappear westward, with no more
756 evidence of a wide thrust sheet or of any other deformation within the Cenozoic foreland
757 deposits (Figs. 7A, 8A and S2A). On the other hand, a thrust ramp is still visible in the seismic
758 profiles through the Proterozoic basement and Protero-Paleozoic to Mesozoic cover of the Tarim
759 Basin, in continuity with the WNW-ESE Mazar Tagh deep fault system (Figs. 6, 7A, 8A and
760 S2A). However, their Cenozoic activity remained limited as indicated by very smooth fault-
761 propagation folds at their tips without any offset of the overlying Cenozoic deposits.

762 In the Tarim Basin, the area extending from the outer range to the Mazar Tagh deep fault
763 system is known as the Southwest depression (Figs. 6, 7A and 7B). In this area, the Tarim
764 basement and sedimentary cover are tilted southward and the Cenozoic deposits thicken toward
765 the Western Kunlun Range to reach their maximum thickness at the front of the outer belt (Figs.
766 7A, 7B, 8A, 8B, S2A and S2B). This overall geometry of the southwestern Tarim is thus typical
767 of a flexural basin associated with the shortening and thickening accumulated in the Western
768 Kunlun thrust system. It corresponds to the region of maximal accommodation space in the
769 Tarim Basin with up to ~10 km of Cenozoic sediments trapped close to the cities of Kashgar and
770 Yecheng, i.e. where the Western Kunlun converge with the Tian Shan and at the front of the
771 central bend of the range, respectively. In this foreland system, the Western Kunlun outer belt
772 corresponds to the wedge-top depozone, while the area comprised between the front of this belt
773 and the Mazar Tagh deep fault system represents the foredeep (after the terminology of DeCelles
774 and Giles, 1996), with the specificity that the whole Cenozoic sediments of the foredeep have

775 been detached from the underlying series to form the Mazar Tagh thrust sheet to the east of the
776 central bend of the mountain range.

777

778 *5.4. Bachu uplift*

779

780 To the north of the Tarim Southwest depression, the Bachu uplift is a ~150-km-wide and
781 ~250-km-long basement-involved structure forming a structural high of the Proterozoic
782 basement and Protero-Paleozoic to Mesozoic (Sinian to Triassic) sedimentary cover of the basin,
783 below the Cenozoic sediments (Figs. 6, 7B, 9 and S2B). From its overall geometry documented
784 by seismic profiles, together with field and well data, this basement-cored structure corresponds
785 to a large pop-up bounded by two conjugate thrust fault systems necessarily rooted in a deep
786 decollement within the basement below the Tarim. Bounding this basement high to the
787 southwest, the Selibuya and Mazar Tagh deep fault systems are composed of a series of fault-
788 related anticlines and small pop-ups associated with steep southwestward-vergent thrust ramps
789 (Figs. 6, 7B, 9 and S2B) (see also Wang et al., 2014a; Yang et al., 2018b; Guo et al., 2016; Li et
790 al., 2016e; Jiang et al., 2018). To the northeast, the Aqia and Tumuxiuke fault systems display
791 large folds related to thrust ramps with a clear northeastward vergence and a hangingwall uplift
792 of a few kilometres, which is larger than the one of the Selibuya and Mazar Tagh fault systems
793 (Figs. 6, 7B, 9 and S2B) (see also Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Guo
794 et al., 2016; He et al., 2016; Li et al., 2016e; Jiang et al., 2018). Consequently, the main vergence
795 of the Bachu pop-up appears to be toward the northeast, i.e. synthetic to the Western Kunlun
796 deformation. In addition, this structure clearly developed parallel to the Western Kunlun Range,
797 with a NNW-SSE orientation to the west changing to a WNW-ESE one to the east. To the
798 southwest of the Bachu uplift, a thrust ramp also exists in lateral continuity with the WNW-ESE
799 Mazar Tagh deep fault system (Figs. 6, 7A, 8A and S2A), but its Cenozoic activity remained
800 limited compared to the one of the NNW-SSE Selibuya faults slightly further north (Tong et al.,
801 2012; Li et al., 2016e; Yang et al., 2018b). Finally, there is no evidence of flexure associated
802 with this large basement structure, neither at its front to the northeast, nor at its rear to the
803 southeast.

804 In the western part of the Bachu uplift, a strong internal deformation occurs in between
805 the bounding fault systems (Fig. 6). There, this large basement pop-up is relatively symmetric

806 with several internal conjugate steep thrusts rooted in the basement and emerging at the surface
807 (Tong et al., 2012; Tang et al., 2014; Wang et al., 2014a; Li et al., 2016e; Jiang et al., 2018;
808 Yang et al., 2018b). Some of these thrusts (Gudongshan Fault System) join the Mazar Tagh deep
809 fault system in the central bend of the Bachu structure, where a relay zone occurs between
810 several fault segments (Fig. 6). However, the internal deformation of the Bachu high tends to
811 decrease eastward of its central bend, with only a few thrust ramps and their associated folds, as
812 well as smooth undulations (Fig. 6). Concomitantly, the Bachu pop-up becomes more
813 asymmetric and acquires the geometry of a large fault-propagation fold with a backlimb broken
814 by the Mazar Tagh deep fault system (Tong et al., 2012; Guo et al., 2016; He et al., 2016; Li et
815 al., 2016e). Finally, at the eastern termination of the Bachu uplift, the offset of the conjugate
816 Tumuxiuke and Mazar Tagh deep faults rapidly decreases eastward, together with the associated
817 basement-cored folding (Fig. 6).

818 Based on seismic profiles and well data, previous studies have demonstrated that the
819 Bachu uplift is an inherited Protero-Paleozoic structure that was probably formed in extension
820 before to be reactivated in compression during the early stages of the Tarim evolution (e.g.,
821 Carroll et al., 2001; Chen and Shi, 2003; Lin et al., 2012b, 2012c; Gao and Fan, 2014; Liu et al.,
822 2015a; He et al., 2016; Wu et al., 2018). The restoration of this uplift at the beginning of the
823 Cenozoic supports this former activity. Indeed, the compressional offsets accumulated by the
824 Pre-Cenozoic deposits along the Bachu bounding and inner faults are not totally reset by this
825 restoration (Fig. 7B). During the Cenozoic, these faults were again reactivated in compression.
826 The Bachu uplift has then developed to the north of the foredeep represented by the Southwest
827 depression, taking the place of the forebulge of the Western Kunlun foreland system (after the
828 terminology of DeCelles and Giles, 1996). Indeed, initially at the surface at the beginning of the
829 Cenozoic, as indicated by a major angular unconformity, the Cenozoic deposits then gradually
830 onlapped and overlapped this reactivated uplift, which is now covered by a few 100s of m of
831 Plio-Quaternary sediments (Figs. 7B and 9) (see also Tong et al., 2012; Tang et al., 2014; Wang
832 et al., 2014a; He et al., 2016; Li et al., 2016e; Jiang et al., 2018).

833

834 *5.5. Deformation amount*

835

836 Along the two transects crossing the Western Kunlun thrust system, we quantified the
837 shortening from the main faults bounding the southwestern Tarim Cenozoic series (i.e. from the
838 Kashgar-Yecheng transfer system and Tiklik thrust system at the rear of the outer range) to the
839 front of the Western Kunlun compressive system (i.e. to the front of the Bachu uplift) (see [Figs.](#)
840 [6, 7A and 7B](#) for the fixed reference lines set at the turning points in the centre of sections A and
841 B). Accordingly, the shortening values and proportions given below do not account for the
842 deformation accommodated within the inner range. In its outer range and foreland only, the
843 Western Kunlun thrust wedge accommodates a large amount of Cenozoic compressive
844 deformation, with a total shortening of ~32 and ~35 km measured west and east of the central
845 bend of the system, respectively ([Fig. 7A and 7B](#)). These shortening estimates are in the range of
846 the other values published from previous cross-section restorations of the outer range, which are
847 comprised between 24.6 and 54 km ([Jiang et al., 2013; Cheng et al., 2016b, 2017](#)).

848 To the west, in the NNW-SSE oriented part of the range (transect A), most of the
849 measured shortening concentrates in the outer range, with ~31,5 km (~98.4% of the ~32 km of
850 total shortening) accommodated within the outer belt and ~0.5 km (~1.6%) within the basin. In
851 the outer belt, ~9.5 km (~29.7% of the ~32 km of total shortening) are accommodated by the
852 basement-involved thrust slices and ~22 km (~68.7%) by the cover duplex. Compared to these
853 values, the shortening of ~0.5 km (~1.6%) accommodated within the basin on the Mazar Tagh
854 deep fault system is almost negligible. However, it should be noted that our section A was built
855 through the western end of the Tarim Basin to cross perpendicularly the various structures
856 intersected from the Western Kunlun to the Tian Shan. Built perpendicularly to the Western
857 Kunlun Range only, it would have crossed the western part of the Bachu uplift where a few
858 additional kilometres of shortening seem to have been accommodated (e.g., [Tong et al., 2012;](#)
859 [Tang et al., 2014; Li et al., 2016e; Jiang et al., 2018; Yang et al., 2018b](#)).

860 To the east, in the WNW-ESE oriented part of the range (transect B), the shortening is
861 distributed between the outer range and the Bachu uplift, with ~30 km (~85.7% of the ~35 km of
862 total shortening) accommodated within the outer belt and ~5 km (~14.3%) within the basin. In
863 the outer belt, ~16.5 km (~47.1% of the ~35 km of total shortening) are accommodated by the
864 basement-involved thrust slices and ~13.5 km (38.6%) by the cover duplex. Consequently, these
865 ~13.5 km of shortening had to be transferred to the Mazar Tagh emergent thrust ramps through
866 the shallow Cenozoic decollement. Across the Bachu uplift, ~2 km (~5.7% of the ~35 km of total

867 shortening) are accommodated by the Mazar Tagh deep fault system and ~3 km (~8.6%) by the
868 Tumuxiuke thrust system. The resulting ~5 km of shortening accommodated by the whole Bachu
869 uplift is thus in agreement with the values of a few kilometres published earlier (Li et al., 2016e).

870 According to these estimates, the shortening accumulated in the Western Kunlun thrust
871 wedge outward from the inner range is of the same order of magnitude to the west and to the east
872 of the central bend of the mountain belt. The structure and high topography of the range interior,
873 together with the large flexure of the Southwest depression, indicate that a large amount of
874 shortening must also be accommodated by the internal part of the system. However, in the
875 absence of stratigraphic markers, it was not possible to restore and quantify this shortening from
876 the geological transects.

877 In addition, previous investigations demonstrated that a significant amount of strike-slip
878 movements is accommodated on major strike-slip fault systems in the range, such as the Karakax
879 fault or the Kashgar-Yecheng transfer system (Ding et al., 2004; Cowgill, 2010). Strike-slip
880 motion could also have occurred in the basin, for example in the western part of the Bachu uplift
881 or along the Mazar Tagh deep fault system where very steep faults and relatively symmetric
882 structures exist. However, our 2D restoration does not allow for quantifying such a deformation
883 that takes place outside of the planes of our sections.

884

885 **6. Southeast deformation**

886

887 To the southeast of the Tarim Basin, the northwestward-vergent Altyn Tagh strike-slip
888 system can be divided into three structural parts with, from the southeast to the northwest: (1) the
889 inner range from the left-lateral Altyn Tagh fault to the Tam Karaul and Northern Altyn Tagh
890 thrusts, (2) the outer range and Southeast depression from the Tam Karaul and Northern Altyn
891 Tagh thrusts to the front of the Minfeng, Ruoqiang, and Dunhuang fault systems, and (3) the
892 Tanan uplift (Figs 6, 7C, 7D, 8C, 8D, S2C and S2D).

893

894 *6.1. Altyn Tagh inner range*

895

896 The inner range is several 10s to ~100 km wide, with typical elevations between ~2500
897 and ~5500 m. It is composed of exhumed imbricates of the Altyn Tagh Protero-Paleozoic (Sinian

898 to Silurian-Devonian) basement and Paleo-Mesozoic (Carboniferous and Jurassic) cover (see
899 [section 2.3](#)). However, this inner range is discontinuous and only developed to the southwest and
900 in the centre of the Altyn Tagh system ([Fig. 6](#)). Indeed, except to the southwest where it is in
901 continuity with the southeastern part of the Western Kunlun interior, the Altyn Tagh inner range
902 disappears where its frontal thrusts (the Tam Karaul and Northern Altyn Tagh thrusts) laterally
903 join and merge into the major left-lateral Altyn Tagh fault at the rear. This fault organization
904 pattern is typical of transpressive wedges. As for the Western Kunlun Range, the detailed
905 structure of the Altyn Tagh inner range is poorly constrained, but connecting the major faults to a
906 deep basal decollement within the basement seems a reasonable hypothesis (see also [Wittlinger
907 et al., 1998](#) and [Tang et al., 2014](#)). To the northwest, the range interior is separated from the
908 outer range by the Tam Karaul and Northern Altyn Tagh emergent thrusts ([Fig. 6](#)). With the
909 Altyn Tagh fault, these thrusts correspond to the main bordering faults of the southeastern
910 Cenozoic series of the Tarim Basin.

911

912 *6.2. Altyn Tagh outer range and Southeast depression*

913

914 At the front of the inner range, the outer range is several 10s to ~60 km wide, with typical
915 elevations between ~1500 and ~2500 m. To our knowledge, no specific cross-section has ever
916 been proposed so far for this outer range. However, contrary to the Western Kunlun compressive
917 system bordering the Tarim Basin to the southwest, the Altyn Tagh transpressive system does
918 not have any fold-and-thrust belt in its outer part. According to the surface and subsurface data,
919 the outer range is mainly composed of several-10s-of-km-long basement-involved structures
920 made of the Altyn Tagh Protero-Paleozoic basement and its Paleo-Mesozoic to Cenozoic
921 (Carboniferous to Quaternary) cover ([Figs. 7C, 7D, 8C, 8D, S2C and S2D](#)). These structures are
922 related to northwestward-vergent, blind or emergent thrust ramps forming the Minfeng,
923 Ruoqiang and Dunhuang fault systems. These ramps necessarily root into a deep decollement
924 within the basement, which is probably in continuity with the one below the inner range.

925 With no sediments older than the Carboniferous (see also [XBGMR, 1985](#); [Li et al., 1996](#);
926 [Tang et al., 2014](#); [He et al., 2016](#)), the Paleozoic stratigraphy of the outer belt is similar to the
927 one of the inner range (e.g., [Sobel and Arnaud, 1999](#); [Chen et al., 2003](#)). In the outer range,
928 unconformities are also frequent in the Paleo-Mesozoic strata or at the base of the Cenozoic

929 deposits in the hanging wall of the thrust ramps (Figs. 7C, 8C and S2C) (see also XBGMR,
930 1985; Li et al., 1996; Tang et al., 2014; He et al., 2016). Moreover, the restoration of the section
931 at the beginning of the Cenozoic does not totally reset the compressional offsets accumulated by
932 the older sediments along these faults (Fig. 7C). All this implies that the latter are inherited
933 structures reactivated during the Cenozoic. In the Altyn Tagh outer range, the whole Cenozoic
934 deformation is associated with these inherited faults since there is no evidence of younger
935 structures rooted in a decollement level in the Paleozoic or Cenozoic sedimentary series.

936 Associated with these reactivated thrusts, the Paleo-Mesozoic and Cenozoic sediments
937 are gently folded into 10s-of-km-wide and up to 100s-of-km-long anticlines and synclines. To
938 the southwest, the Cenozoic deposits related to the erosion of the Altyn Tagh cover these folds
939 and the topographic mountain front is located behind, at the front of the inner range. To the
940 northeast, they form smooth topographies uplifted several 100s of m above the present-day
941 Tarim sedimentation level and localize the topographic mountain front.

942 In the Tarim Basin, the area at the front of these folds is known as the Southeast
943 depression (Figs. 6, 7C, and 7D). In this area, the Tarim basement and sedimentary cover are
944 gently tilted southeastward, with the Cenozoic deposits slightly thicker at the front of the Altyn
945 Tagh outer belt than further northwest in the basin (Figs. 7C, 7D, 8C, 8D, S2C and S2D). This
946 overall geometry of the southeast Tarim thus indicates a slight flexure associated with the
947 shortening and thickening accumulated in the Altyn Tagh strike-slip system. However, with ~3
948 km of trapped Cenozoic sediments at most, the generated accommodation space remains
949 moderate. This suggests a limited compressive deformation accommodated across the Altyn
950 Tagh Range, which is somehow expected when compared to a thrust wedge such as the Western
951 Kunlun Range for example. Moreover, the southwestern part of the Altyn Tagh foredeep is
952 broken by the Tanan basement uplift, which probably explains the limited accommodation space
953 in this foreland (Strecker et al., 2012).

954

955 *6.3. Tanan uplift*

956

957 To the north of the Tarim Southeast depression, the Tanan uplift (also known as the
958 Qiemu uplift) is a ~50-km-wide and ~600-km-long basement-involved structure forming a
959 structural high of Proterozoic basement and Paleo-Mesozoic (Carboniferous to Cretaceous)

960 cover, below the Cenozoic sediments of the basin (Figs. 6, 7C, 8C and S2C) (see also Lin et al.,
961 2012b, 2012c; Tang et al., 2014; Li et al., 2015c; Liu et al., 2015a; He et al., 2016). From its
962 overall geometry documented by seismic profiles, together with field and well data, this
963 basement-cored structure corresponds to a large pop-up bounded by two conjugate fault systems
964 necessarily rooted in a deep decollement within the basement below the Tarim. To the south, the
965 Niya fault system is composed of a few conjugate thrust faults that locally delimitate a small
966 pop-up, at the tip of a southward-vergent steep thrust ramp (Figs. 6, 7C, 8C and S2C) (see also
967 Tang et al., 2014; Li et al., 2015c). To the north, the Cherchen fault system displays fault-related
968 folds and steep thrust ramps with a clear northward vergence and a hangingwall uplift of up to
969 ~1.5 km, which is larger than that of the Niya fault system (Figs. 6, 7C, 8C and S2C) (see also
970 Lin et al., 2012b, 2012c; Tang et al., 2014; Li et al., 2015c; Liu et al., 2015a; He et al., 2016;
971 Jiang et al., 2018). Consequently, the Tanan pop-up has clearly developed parallel to the Altyn
972 Tagh Range over its entire length (Fig. 6), and its main vergence appears to be toward the
973 northwest, i.e. synthetic to the Altyn Tagh deformation. However, in spite of this clear
974 northwestward vergence, there is no evidence of flexural subsidence at the front of the Tanan
975 uplift (Figs. 7C, 8C and S2C).

976 In its central part, where the two conjugate bounding fault systems are present, this large
977 pop-up is relatively symmetric (Figs. 6, 7C, 8C and S2C) (see also Li et al., 2015c). However,
978 where the Niya fault system vanishes laterally, the Tanan pop-up becomes more asymmetric and
979 acquires the geometry of a large fault-propagation fold (Liu et al., 2015a; Jiang et al., 2018).
980 Finally, at the lateral terminations of the Tanan uplift, the offset of the Cherchen fault system
981 decreases and the associated basement-cored fold progressively flattens toward the eastern part
982 of the Tarim Basin to the northeast and the eastern end of the Western Kunlun Range to the
983 southwest (Fig. 6). In this area, the Hotan-Qira normal fault system is a series of conjugate
984 normal faults described and interpreted by Avouac and Peltzer (1993) as related to the
985 subsidence of the Tarim Basin loaded by the Western Kunlun mountain range. However, seismic
986 profiles in this area (Fig. 1) led us to re-evaluate these faults as extrado normal faults above the
987 southwestward vanishing basement-cored fold of the Tanan uplift (see also Li et al., 2018b).

988 Based on seismic profiles, together with field and well data, previous studies have
989 demonstrated that the Tanan uplift is an inherited Protero-Paleozoic compressive structure active
990 during the early stages of the Tarim evolution (e.g., Lin et al., 2012b, 2012c; Gao and Fan, 2014;

991 [Lin et al., 2015a](#); [Liu et al., 2015a](#); [He et al., 2016](#)). Here again, the restoration of this uplift at
992 the beginning of the Cenozoic supports a former activity since the compressional offsets
993 accumulated by the Pre-Cenozoic deposits along the Tanan bordering faults are not totally reset
994 ([Fig. 7C](#)). As for the Bachu uplift, these faults were then reactivated in compression during the
995 Cenozoic. However, although well visible on seismic profiles all along the Altyn Tagh Range,
996 the Cherchen fault system has not been reactivated significantly all along its entire length.
997 Indeed, at the eastern end of the basin, this fault system does not yet present any geometrical
998 evidence for a significant cumulative Cenozoic deformation ([Figs. 7D, 8D and S2D](#)) (see also
999 [Morin et al., 2019](#)). Yet, reactivation of this part of the fault system could be currently ongoing
1000 since some clues of recent activity exist at its eastern termination ([Chang et al., 2014a](#)).

1001 Where it is reactivated, the Cherchen fault system now corresponds to the present-day
1002 hidden to emergent deformation front of the Altyn Tagh wedge, located in the basin ~100-150
1003 km away from the topographic mountain front. Remarkably, this structure also localizes a major
1004 stratigraphic change between the Altyn Tagh Paleozoic domain with Silurian to Devonian
1005 metamorphic and plutonic rocks covered by Carboniferous deposits (see [section 2.3](#)) to the
1006 south, and the Tarim Paleozoic domain with Cambrian to Permian sediments (see [section 2.1](#)) to
1007 the north ([Figs. 7C and 8C](#)) (e.g., [Lin et al., 2012b, 2012c](#); [Tang et al., 2014](#); [Li et al., 2015c](#); [Liu](#)
1008 [et al., 2015a](#); [He et al., 2016](#)). Accordingly, the Cherchen fault system is a major inherited
1009 structure equivalent to the deep basement ramp below the Western Kunlun outer belt, where a
1010 similar change between two stratigraphic domains is localized. Together, these structures
1011 therefore represent a major tectonic palaeo-boundary between the Paleozoic orogenic systems of
1012 the Tibetan edge to the south and the contemporaneous Tarim Basin to the north.

1013 This tectonic palaeo-boundary was thus reactivated during the Cenozoic all along the
1014 northern edge of the Tibetan Plateau. In the foreland of the Western Kunlun compressive system,
1015 the deformation also propagated as far north as the Bachu uplift ([Figs. 6, 7B and S2B](#)), whereas
1016 in the foreland of the Altyn Tagh transpressive system, there is no evidence of Cenozoic activity
1017 further north than the Tanan uplift ([Figs. 6, 7C and S2C](#)), even though some inherited Protero-
1018 Paleozoic structures, known as the Tazhong and Tadong uplifts, also exist in the central part of
1019 the basin ([Figs. 6, 7C, 7D, S2C and S2D](#)) (see also [Lin et al., 2012b, 2012c](#); [Tang et al., 2014](#);
1020 [Lin et al., 2015c](#); [Liu et al., 2015a](#); [He et al., 2016](#)). Indeed, there is little evidence of a
1021 significant Cenozoic activity on the bounding faults of these basement highs, and if deformation

1022 occurs in this part of the basin, it remains very limited. In this undeformed area of the Tarim, the
1023 Cenozoic deposits are typically less than ~2 km thick.

1024

1025 *6.4. Deformation amount*

1026

1027 Along the two transects crossing the Altyn Tagh strike-slip system, we quantified the
1028 shortening from the main faults bounding the southeastern Tarim Cenozoic series (i.e. from the
1029 Tam Karaul and Northern Altyn Tagh thrust faults at the rear of the outer range) to the centre of
1030 the basin (see [Figs. 6, 7C and 7D](#) for the fixed reference lines set at the turning points in the
1031 centre of sections C and D). Accordingly, the shortening values and proportions given just below
1032 do not account for the deformation accommodated within the inner range. In its outer range and
1033 foreland only, the Altyn Tagh strike-slip wedge accommodates a small amount of Cenozoic
1034 compressive deformation, with a total shortening of ~0.9 and ~0.3 km measured to the west and
1035 to the east of the system, respectively ([Fig. 7C and 7D](#)).

1036 To the west (transect C), the measured shortening is distributed between the outer range
1037 and the Tanan uplift, with ~0.3 km (~33.3% of the above ~0.9 km of total shortening)
1038 accommodated within the outer belt and ~0.6 km (~66.7%) within the basin. Across the Tanan
1039 uplift, ~0.2 km (~22.2% of the ~0.9 km of total shortening) are accommodated by the Niya fault
1040 system and ~0.4 km (~44.5%) by the Cherchen thrust and associated faults. To the east (transect
1041 D), the entire measured ~0.3 km shortening concentrates in the outer range and is accommodated
1042 by the local basement thrust ramps.

1043 According to these estimates, the shortening accumulated within the Altyn Tagh strike-
1044 slip wedge outward from the inner range is very limited. Moreover, it decreases from the
1045 southwest to the northeast to become almost negligible at the eastern termination of the system.
1046 The structure and high topography of the inner range indicate that most of the shortening must
1047 rather be accommodated in this internal part of the system. However, even though we were not
1048 able to restore and quantify this deformation given the lack of stratigraphic markers in the range
1049 interior, the slight flexure associated with the Altyn Tagh wedge suggests a limited cumulative
1050 shortening and thickening, as expected for such a strike-slip system. Indeed, previous studies
1051 demonstrated that a large amount of strike-slip deformation is accommodated on major strike-
1052 slip faults in the range such as the Altyn Tagh fault (e.g., [Wang, 1997](#); [Cowgill et al., 2000](#),

1053 2003; Sobel, 2001; Yue et al., 2001; Meng et al., 2001; Yin et al., 2002; Gehrels et al., 2003;
1054 Ding et al., 2004; Cheng et al., 2015, 2016a). Strike-slip motion may also have occurred in the
1055 basin, along the Niya and Cherchen bounding fault systems of the Tanan uplift, where very steep
1056 thrusts and relatively symmetric structures exist (see also Avouac and Peltzer, 1993; Yin et al.,
1057 2002). However, our 2D restoration does not allow for quantifying these movements, which take
1058 place outside of the planes of our sections.

1059

1060 **7. Northwest deformation**

1061

1062 To the northwest of the Tarim Basin, the southeastward-vergent Southwestern Tian Shan
1063 thrust system can be divided into three structural parts with, from the northwest to the southeast:
1064 (1) the inner range from the Nikolaev Line and Narat fault to the Muziduke thrust and Southern
1065 Tian Shan fault system, (2) the outer range from the Muziduke thrust and Southern Tian Shan
1066 fault system to the Kashgar deformation front, Kepingtage thrust, Kalpintagh fault, and Kuqa
1067 deformation front, and (3) the Southwest and North depressions at the front of the Kashgar
1068 deformation front, Kalpintagh thrust and Kuqa deformation front (Figs. 6, 7A, 7B, 7C, 8A, 8B,
1069 10C, S2B and S2C).

1070

1071 *7.1. Southwestern Tian Shan inner range*

1072

1073 The inner range is several 10s to ~200 km wide, with typical elevations between ~2500
1074 and ~7400 m. It is composed of exhumed imbricates of the Southern Tian Shan Proterozoic
1075 basement, Paleozoic (Devonian to Permian) plutons, and Paleozoic-Mesozoic (Silurian to
1076 Cretaceous) cover (see section 2.4). It also comprises a few compressive intra-mountainous
1077 basins filled by Cenozoic sediments, such as the Bayanbulak Basin for example (Fig. 6). In the
1078 Southwestern Tian Shan system, the inner range is a continuous belt crossed by the Talas
1079 Fergana fault to the southwest and in continuity with the Southeastern Tian Shan interior to the
1080 northeast (Fig. 6). Despite numerous surface and subsurface studies, its detailed structure
1081 remains controversial. To the southwest, connecting the major basement faults, such as the
1082 Muziduke thrust and Southern Tian Shan fault system, to a deep basal decollement seems a
1083 reasonable hypothesis (Figs. 7A, 7B, 10A and 10B) (see also Buslov et al., 2007; Gao et al.,

1084 2013; Jia et al., 2015). However, alternative interpretations have been proposed with these major
1085 faults crossing the entire crust or branching onto the Southern Tian Shan fault system (e.g., Allen
1086 et al., 1999; Loury et al., 2017; Makarov et al., 2010; Xiao et al., 2013). To the northeast, the
1087 Southern Tian Shan fault system is much steeper and possibly cuts through the whole crust since
1088 its location at the surface correlates with a step in the Moho at depth (Figs. 7C and 10C) (Li et
1089 al., 2007; Qi et al., 2009; Wen et al., 2017). Alternatively, this major basement fault system has
1090 also been suggested to connect at depth into a deep basal decollement (e.g., Yin et al., 1998;
1091 Poupinet et al., 2002). Toward the southeast, the range interior is then separated from the outer
1092 range by the Muziduke thrust to the southwest and the Southern Tian Shan fault system to the
1093 northeast (Fig. 6). These emergent thrust zones also correspond to the main bordering faults of
1094 the northwestern Cenozoic series of the Tarim Basin.

1095 Remarkably, in the western and central parts of the Southwestern Tian Shan, the
1096 stratigraphy of the Paleozoic terrains is different on both sides of the Muziduke thrust and
1097 Southern Tian Shan fault system (Figs. 7A, 7B, 10A and 10B). To the northwest, this
1098 stratigraphy is characteristic of the Southern Tian Shan Paleozoic Range with Silurian to
1099 Devonian metamorphic and plutonic rocks covered by Devonian to Permian deposits (see section
1100 2.4), whereas to the southeast, it is typical of the Paleozoic Tarim Basin with Cambrian to
1101 Permian sediments (see section 2.1) (Fig. 2). Therefore, the Muziduke thrust and Southern Tian
1102 Shan fault system seem to be major inherited Paleozoic structures that have been reactivated
1103 during the Cenozoic. Such a major stratigraphic change also occurs in the northeastern part of the
1104 Southwestern Tian Shan, but it seems to be located slightly to the south, within the outer range.

1105

1106 *7.2. Southwestern Tian Shan outer range and Kuqa depression*

1107

1108 At the front of the inner range, the outer range is also several 10s to ~100 km wide, with
1109 typical elevations between ~1000 and ~2500 m. It consists in four different fold-and-thrust belts
1110 that developed below and above the Cenozoic sediments of the Tarim Basin (Figs. 6, 7A, 7B, 7C
1111 and S2C). These four belts present different styles of deformation all along the range. They are
1112 presented below from the southwest to the northeast.

1113

1114 7.2.1. Kashgar-Artux fold-and-thrust belt

1115

1116 To the southwestern end of the Southwestern Tian Shan outer range, the Kashgar-Artux
1117 fold-and-thrust belt is a ~150-km-long belt located between the Muziduke thrust and the Kashgar
1118 deformation front (Fig. 6). Several cross-sections have already been proposed across this belt
1119 (e.g., Scharer et al., 2004; Sobel et al., 2006; Heermance et al., 2007, 2008). The surface and
1120 subsurface data provide clear evidence for 10s-of-km-long, emergent or blind, folded thrust
1121 slices made of the Tarim Paleo-Mesozoic (Cambrian to Cretaceous) sedimentary cover at the
1122 rear, whereas several-km-wide and 10s-of-km-long fault-propagation or detachment folds
1123 developed within the Tarim Cenozoic (Paleogene to Quaternary) deposits at the front.

1124 Together with the basement-involved imbricates of the inner range, these structures imply
1125 a basal decollement level that progressively shallows from the northwest to the southeast.
1126 Indeed, in the range interior, the major thrusts probably root in a deep decollement within the
1127 basement. At the transition between the inner and outer ranges, the change from thick-skinned to
1128 thin-skinned thrust slices that operates within the Paleo-Mesozoic cover then requires the
1129 presence of a ramp connecting this deep basement decollement to an intermediate one, most
1130 probably within the upper Cambrian evaporitic formation found in the western Tarim
1131 sedimentary series (e.g., Heermance et al., 2007, 2008). However, the poor quality of the seismic
1132 profiles and the lack of deep well data do not allow to precisely constrain the depth of this
1133 Paleozoic decollement in this part of the basin.

1134 Toward the front of the outer belt, the formation of thrusts and folds within the Tarim
1135 Cenozoic cover then required the activation of a second ramp connecting this intermediate
1136 decollement at the base of the deformed Paleo-Mesozoic series to a shallow one at its top.
1137 According to the seismic profiles, together with field and well constraints, this shallow
1138 decollement corresponds to the Paleogene evaporitic assemblage located at the base of the
1139 Cenozoic sediments, at a depth of about 6 to 9 km b.s.l. in this area (e.g., Scharer et al., 2004;
1140 Sobel et al., 2006; Heermance et al., 2007, 2008). In the Kashgar-Artux belt, there is no evidence
1141 for a structural inheritance such as stratigraphic changes or non-restorable offsets across faults.
1142 Accordingly, all the structures developed in this belt likely correspond to structures created
1143 during the Cenozoic.

1144 Although partially covered by the Cenozoic deposits coming from the Southwestern Tian
1145 Shan erosion, the frontal folds of the Kashgar-Artux belt form smooth topographies uplifted

1146 several 10s to 100s of m above the present-day Tarim sedimentation level (Fig. 6). Along this
1147 belt, the local topographic front of the mountain range is thus localized by the detachment fold
1148 associated with the Kashgar deformation front, beyond which the Cenozoic deposits of the
1149 foreland have remained undeformed and attached to the underlying series.

1150

1151 7.2.2. Kepingtage fold-and-thrust belt

1152

1153 In the southwestern part of the Southwestern Tian Shan outer range, the Kepingtage fold-
1154 and-thrust belt is a ~230-km-long belt comprised between the Muziduke thrust to the northwest
1155 or Southern Tian Shan fault system to the northeast, and the Kepingtage thrust (Fig. 6). Several
1156 cross-sections have already been proposed through the Kepingtage belt (e.g., Yin et al., 1998;
1157 Allen et al., 1999; Turner et al., 2010; Tang et al., 2014; Yang et al., 2018b). In agreement with
1158 the previous interpretations, we concluded that the surface and subsurface data provide clear
1159 evidence for a few 10s-of-km-wide and 10s-of-km-long thrust sheets made of the Tarim
1160 Paleozoic (Cambrian to Permian) sedimentary cover, with fault-propagation folds related to
1161 emergent thrust ramps at their front (Figs. 7 and 10A). These thrust sheets also include Cenozoic
1162 (Paleogene to Quaternary) sediments deposited in up to ~4-km-deep piggy-back basins.
1163 Associated with these compressive structures, a set of strike-slip faults perpendicular to the thrust
1164 sheets also exists in the Kepingtage belt, including the left-lateral Piqiang fault (Fig. 6) (see
1165 Allen et al., 1999; Turner et al., 2010, 2011; Yang et al., 2018b).

1166 As in the Kashgar-Artux belt, the transition from the basement imbricates of the inner
1167 range to the sedimentary cover thrust sheets of the Kepingtage belt implies a basal decollement
1168 level that progressively shallows from the northwest to the southeast. Indeed, the major thrusts of
1169 the range interior probably root in a deep decollement within the basement, while the change
1170 from thick-skinned to thin-skinned thrust slices that involves the Paleozoic and Cenozoic cover
1171 requires the presence of a ramp connecting this deep decollement to the intermediate one within
1172 the upper Cambrian evaporitic formation of the western Tarim sedimentary series (Fig. 2) (e.g.,
1173 Allen et al., 1999; Turner et al., 2010). According to the seismic profiles and field constraints,
1174 this Paleozoic decollement is located here at a depth of about 4 to 6 km b.s.l. (Figs. 7A and 10A)
1175 (see also Allen et al., 1999; Turner et al., 2010; Yang et al., 2018b). In the Kepingtage belt, the
1176 whole deformation is accommodated by the thrust sheets that grew above this decollement level,

1177 since there is no evidence of structures rooted in a shallower one within the Cenozoic
1178 sedimentary series.

1179 Based on field data, previous studies have documented unconformities, as well as SW-
1180 NE thickness and facies changes, in the Paleozoic sedimentary series of the Kepingtage belt (e.g.
1181 [Allen et al., 1999](#); [Turner et al., 2010, 2011](#); [Chang et al., 2012b](#)). Moreover, the strike-slip faults
1182 offsetting the belt are aligned with the faults present in the northwestern part of the Bachu uplift
1183 to the southeast. These observations suggest that the Kepingtage area was the place of a large
1184 inherited structure, in continuity with the one of the Bachu uplift, which has also been
1185 reactivated during the Cenozoic (e.g. [Allen et al., 1999](#); [Carroll et al., 2001](#); [Turner et al., 2010,](#)
1186 [2011](#); [Yang et al., 2018b](#)). However, the reactivated strike-slip faults appear to be perpendicular
1187 to the compressive structures of the Kepingtage belt, for which there is no evidence of a
1188 structural inheritance such as stratigraphic changes or non-restorable offsets across the faults.
1189 Therefore, the shallow thrusts rooting into the Paleozoic decollement likely correspond to faults
1190 created during the Cenozoic above a large basement-involved structure, whose inheritance rather
1191 controls the occurrence of orthogonally-oriented strike-slip deformation (e.g., [Turner et al., 2010,](#)
1192 [2011](#); [Yang et al., 2018b](#)).

1193 Although partially covered by the Cenozoic deposits resulting from the erosion of the
1194 Southwestern Tian Shan, the frontal part of the Kepingtage thrust sheets form sharp topographies
1195 uplifted several 100s of m to more than ~1 km above the present-day Tarim sedimentation level
1196 ([Fig. 6](#)). In this area, the local topographic mountain front is thus localized by the Kepingtage
1197 thrust ramp emerging at the front of the belt.

1198

1199 7.2.3. Kalpin belt

1200

1201 In the central part of the Southwestern Tian Shan outer range, the Kalpin belt is a ~80-
1202 km-long belt located between the Southern Tian Shan fault system and the Kalpintagh fault ([Fig.](#)
1203 [6](#)). As far as we know, there has been no published cross-section across this belt. According to
1204 the surface and subsurface data, this belt is composed of 10s-of-km-long, emergent and folded
1205 thrust slices and small pop-ups related to a few steep thrust ramps. These structures involved the
1206 Tarim Proterozoic basement and its Protero-Paleozoic (Sinian to Permian) cover ([Figs. 7B and](#)
1207 [10B](#)). They also individualize two basins containing less than ~2 km of Cenozoic (Plio-

1208 Quaternary) deposits. At depth, these structures possibly connect to a deep decollement within
1209 the basement. In the Kalpin belt, like in the Kepingtage one, there is no evidence of structures
1210 rooting into a shallow decollement level within the Cenozoic sedimentary series.

1211 According to field data, the Kalpin area, which is known as the Kalpin uplift, appears to
1212 be a lateral extension of the large inherited Kepingtage basement structure (e.g. [Allen et al.,](#)
1213 [1999](#); [Carroll et al., 2001](#); [Turner et al., 2010](#); [Chang et al., 2012b](#)). Consequently, in contrast
1214 with the neo-formed compressive structures of the thin-skinned Kepingtage fold-and-thrust belt,
1215 the basement-involved structures of the thick-skinned Kalpin belt are probably related to
1216 inherited basement faults reactivated during the Cenozoic, such as possibly for the frontal
1217 Kalpintagh thrust fault.

1218 Although partially covered by the Cenozoic deposits coming from the erosion of the
1219 Southwestern Tian Shan, the structures of the Kalpin belt now form sharp topographies uplifted
1220 several 100s of m up to ~1.5 km above the present-day Tarim sedimentation level ([Fig. 6](#)). In this
1221 area, the local topographic front of the mountain range is localized by the Kalpintagh fault
1222 emerging at the front of the belt.

1223

1224 7.2.4. Kuqa fold-and-thrust belt and depression

1225

1226 In the northeastern part of the Southwestern Tian Shan outer range, the Kuqa fold-and-
1227 thrust belt is a ~370-km-long belt located between the Southern Tian Shan fault system and the
1228 Kuqa deformation front ([Fig. 6](#)). Numerous contradictory cross-sections have been proposed
1229 through this belt (e.g. [Burchfield et al., 1999](#); [Chen et al., 2004](#); [He et al., 2005](#); [Qi et al., 2009](#);
1230 [Li et al., 2012a](#); [Wu et al., 2014](#); [Zhang et al., 2014b](#); [Li et al., 2016d](#); [Wen et al., 2017](#); [Zhao and](#)
1231 [Wang, 2016](#); [Wang et al., 2017](#); [Izquierdo-Llavall et al., 2018](#); [Neng et al., 2018](#)). From our
1232 surface and subsurface dataset, we derived a regionally-consistent structural model settling the
1233 issue of these contradictory interpretations (see [Supplementary information S4](#) for further details
1234 and discussion). At the rear of the belt, the surface and subsurface data provide evidence for 10s-
1235 of-km-long, emergent or blind, folded thrust slices constituted of the Southern Tian Shan
1236 Proterozoic basement, Paleozoic (Devonian to Permian) plutons and Paleo-Mesozoic (Silurian to
1237 Cretaceous) cover. At the front, 10s-of-km-long duplex structures involve the Tarim Paleo-
1238 Mesozoic (Cambrian? to Cretaceous) sedimentary cover ([Figs. 7C, 10C and S2C](#)). The surface

1239 and subsurface data also reveal several-km-wide and up to ~200-km-long fault-propagation or
1240 detachment folds that developed within the Tarim Cenozoic (Paleogene to Quaternary) deposits
1241 at the top and the front of the basement-involved thrust slices and Paleo-Mesozoic cover
1242 duplexes (Figs. 7C, 10C and S2C). Finally, these folds individualize a few piggy-back basins,
1243 such as the ~30-km-wide and ~130-km-long Kuqa depression where ~9 km of Cenozoic
1244 sediments are stacked nearby the city of Baicheng (Fig. 6, 7C and 10C).

1245 These structures imply a basal decollement level that progressively shallows from the
1246 north to the south. At the rear of the Kuqa belt, the major thrusts necessarily connect to a deep
1247 decollement within the basement or directly onto the Southern Tian Shan fault system. The
1248 southeastward change from the thick-skinned thrust slices to the thin-skinned duplexes then
1249 requires the presence of a ramp transmitting slip from depth to an intermediate decollement at
1250 the base of the deformed Paleo-Mesozoic sedimentary cover (Figs. 7C and 10C). However, in
1251 this part of the basin, the quality of the seismic profiles and the lack of deep well data do not
1252 allow for precisely constraining the age, lithology and depth of this intermediate decollement that
1253 remains unclear (compare the cross-sections of Chen et al., 2004; He et al., 2005; Qi et al., 2009;
1254 Li et al., 2012a; Wen et al., 2017; Wang et al., 2017; Izquierdo-Llavall et al., 2018; Neng et al.,
1255 2018; and Supplementary information S4 for more details on the different interpretations).

1256 Nonetheless, the stratigraphy of the Paleozoic terrains appears to be different on both
1257 sides of the ramp connecting the deep and intermediate decollements (Figs. 7C and 10C). To the
1258 northwest, this stratigraphy is characteristic of the Southern Tian Shan Paleozoic domain with
1259 Silurian to Devonian metamorphic and plutonic rocks directly covered by Devonian to Permian
1260 deposits (see section 2.4) (Fig. 2). To the southeast, the Proterozoic basement of the Tarim Basin
1261 could be topped by a lower Paleozoic series, even if the latter is truncated by an unconformity
1262 located at the base of the overlying Mesozoic sediments (Fig. 2). The occurrence within this
1263 Paleo-Mesozoic cover of a sedimentary level acting as an intermediate decollement in the Kuqa
1264 belt, but which does not exist at the surface further north, is an additional indication of a
1265 significant stratigraphic change between these two Paleozoic domains with Southern Tian Shan
1266 vs. Tarim stratigraphic affinities. Therefore, the ramp connecting the deep and intermediate
1267 decollements to the northwest of the Kuqa belt seems to be a major inherited Paleozoic structure
1268 reactivated during the Cenozoic, similar to the Muziduke thrust and the Southern Tian Shan fault
1269 system further west. Together, these structures thus represent a major structural palaeo-boundary

1270 between the Paleozoic orogenic systems of the Southern Tian Shan to the north and the
1271 contemporaneous Tarim Basin to the south.

1272 At the front of the Kuqa belt, the formation of duplexes within the Tarim Paleo-Mesozoic
1273 cover and of thrusts and folds within the Cenozoic deposits requires the activation of blind ramps
1274 connecting the intermediate decollement to a shallow one between the two sedimentary series
1275 (Figs. 7C, 10C and S2C). Based on the field constraints, seismic profiles, and well data, this
1276 shallow decollement is located at a depth of about 5 to 8 km b.s.l. at the front of the duplexes,
1277 within the Paleogene evaporitic assemblage of the Cenozoic deposits (see also Chen et al., 2004;
1278 He et al., 2005; Li et al., 2012a; Wu et al., 2014; Tian et al., 2016; Izquierdo-Llavall et al., 2018;
1279 Neng et al., 2018). Below this decollement, the interpretation and reconstruction of the
1280 underlying duplexes may be problematic because of a strong component of salt tectonics, as well
1281 as a seismic imaging degraded by the presence of evaporites (e.g., Chen et al., 2004; He et al.,
1282 2005; Li et al., 2012a; Wu et al., 2014; Li et al., 2016d; Tian et al., 2016; Wang et al., 2017;
1283 Izquierdo-Llavall et al., 2018; Neng et al., 2018). However, the stratigraphy of the Paleo-
1284 Mesozoic cover seems to be the same across the ramps connecting the Paleozoic and Cenozoic
1285 decollements. Therefore, these shallow ramps in the centre of the Kuqa belt likely correspond to
1286 structures created during the Cenozoic.

1287 Above the basement-involved thrust slices and Paleo-Mesozoic cover duplexes, the slip is
1288 transferred toward the top and the front of the Kuqa belt through the Cenozoic decollement.
1289 Within the Cenozoic deposits, long fault-propagation or detachment folds developed in
1290 association with emergent or blind thrust ramps, often with a strong component of salt tectonics
1291 (Figs. 7C and 10C) (e.g. Chen et al., 2004; He et al., 2005; Hubert-Ferrari et al., 2007; Li et al.,
1292 2012a; Wu et al., 2014; Li et al., 2016d; Tian et al., 2016; Zhao and Wang, 2016; Wang et al.,
1293 2017; Izquierdo-Llavall et al., 2018; Neng et al., 2018). These folds are located just at the front
1294 of the structures involving the basement and the Paleozoic cover below the Cenozoic
1295 decollement, or a few 10s of km further south. In this case, these most frontal folds individualize
1296 the Kuqa depression that corresponds to a piggy-back basin developed within a wide syncline at
1297 their rear (Figs. 6, 7C and 10C).

1298 Although largely covered by the Cenozoic deposits resulting from the erosion of the
1299 Southwestern Tian Shan, these frontal folds localize the topographic front of the mountain range.
1300 At the surface of the basin, they usually form smooth topographies uplifted several 100s of m

1301 above the present-day Tarim sedimentation level (Fig. 6). Along the belt, the topographic front is
1302 thus localized by the detachment folds at the front of the Kuqa belt. Further southeast, the
1303 Cenozoic deposits of the foreland have remained undeformed and attached to the underlying
1304 series.

1305

1306 *7.3. Southwest and North depressions*

1307

1308 The Tarim Basin at the front of the Southwestern Tian Shan outer range can be
1309 considered in terms of different sub-areas, which are presented below from the southwest to the
1310 northeast.

1311 The Tarim Basin sub-area located at the front of the Kashgar-Artux belt corresponds to
1312 the northwestern termination of the Southwest depression (Fig. 6). There, this depression
1313 corresponds to the region where the accommodation space is maximal in the Tarim Basin, with
1314 up to ~10 km of Cenozoic sediments nearby Kashgar. Indeed, the added loads of the Western
1315 Kunlun and Southwestern Tian Shan thrust wedges probably explain the very strong flexural
1316 subsidence of this specific area. In this foreland system, the Western Kunlun outer belt and the
1317 Kashgar-Artux belt correspond to two facing wedge-top depozones, and the Southwest
1318 depression located between these belts to a triangular foredeep, according to the terminology of
1319 DeCelles and Giles (1996).

1320 At the front of the Kepingtage belt, there is no such depression (Figs. 7A and 10A).
1321 Indeed, this belt is thrust over the western part of the Bachu uplift, which clearly prevents the
1322 development of a flexural foredeep (Fig. 6). Moreover, the Bachu high seems to persist and be
1323 also active below the Kepingtage belt (see also Allen et al., 1999; Turner et al., 2010), possibly
1324 interfering with its deformation. In particular, this may explain why the Kepingtage wedge-top
1325 depozone (terminology of DeCelles and Giles, 1996) is mainly developed above the
1326 sedimentation level, with deeply-eroded anticlines at the front of the thrust sheets and shallow
1327 piggy-back basins at their top.

1328 Finally, the Tarim Basin sub-area located at the front of the Kalpin and Kuqa belts is
1329 known as the North depression (Fig. 6). In this area, the Tarim basement and sedimentary cover
1330 are tilted northward so that the Cenozoic deposits thicken toward the Southwestern Tian Shan
1331 and reach their maximum thickness at the front of its outer belt (Figs. 7B, 7C, 10B, 10C, S2B

1332 and S2C). This overall geometry of the northern Tarim is thus typical of a flexural basin,
1333 associated here with the shortening and thickening accumulated in the Southwestern Tian Shan
1334 thrust system. It corresponds to a region of high accommodation space, with up to ~7 km of
1335 Cenozoic sediments to the south of Aksu, i.e. at the front of the Kalpin belt (Figs. 7B, 10B and
1336 S2B). Moreover, to the northeast, the North depression is bounded by the Kuqa belt, which in
1337 turn surrounds the Kuqa depression where up to ~9 km of Cenozoic deposits are trapped close to
1338 the city of Baicheng (Figs. 7C and 10C). In this foreland system, the Kuqa belt and its associated
1339 depression correspond to the wedge-top depozone, the North depression at the front of this belt
1340 represents the foredeep, and the undeformed central area of the basin constitutes the forebulge
1341 (terminology of DeCelles and Giles, 1996).

1342 At the transition between these foredeep and forebulge, an inherited Protero-Paleozoic
1343 basement-cored structure called the Tabei uplift exists. However, there is little evidence of a
1344 significant Cenozoic activity on the faults associated with this basement high, except to its very
1345 eastern part. There, its southern bounding fault appears to have been reactivated as a normal
1346 fault, most probably in response to the stresses generated by the flexural bending of the northern
1347 Tarim Basin. However, this Cenozoic deformation occurring in the northern part of the basin
1348 remains very limited (Figs. 7C, 10C and S2C).

1349

1350 7.4. Deformation amount

1351

1352 Along the transects crossing the Southwestern Tian Shan thrust system, we quantified the
1353 shortening from the main faults bounding the northwestern Tarim Cenozoic series (i.e. from the
1354 Muziduke thrust and Southern Tian Shan fault system at the rear of the outer range) to the centre
1355 of the basin (see Figs. 6, 7A, 7B and 7C for the fixed reference lines set at the turning points in
1356 the centre of sections A, B and C). Accordingly, the shortening values and proportions given just
1357 below do not account for the deformation accommodated within the inner range. In its outer
1358 range only, the Southwestern Tian Shan thrust wedge already accommodates a large amount of
1359 Cenozoic compressive deformation, with a total shortening of ~36 and ~22 km measured to the
1360 southwest and to the northeast, i.e. through the Kepingtage and Kuqa belts, respectively (Figs.
1361 7A and 7C). These shortening amounts are in the range of the other values published from
1362 previous cross-section restorations of these systems, which range from 23 to 50 km for the

1363 Kepingtage belt (e.g., Yin et al., 1998; Allen et al., 1999; Yang et al., 2007b; Turner et al., 2010),
1364 and from ~10 to ~42 km for the Kuqa one (e.g., Yin et al., 1998, Burchfield et al., 1999; Li et al.,
1365 2012a; Zhang et al., 2014b; Tian et al., 2016; Wen et al., 2017; Izquierdo-Llavall et al., 2018).
1366 Unfortunately, in the absence of a preserved Mesozoic-Cenozoic stratigraphic marker, it was not
1367 possible to properly quantify the Cenozoic deformation accumulated in the central part of the
1368 Southwestern Tian Shan compressive system, i.e. through the Kalpin belt, since eroded material
1369 cannot be reconstructed. The proposed restoration therefore only provides a very low and
1370 probably meaningless minimum estimate of ~4 km for the total shortening across this belt.

1371 In the Southwestern Tian Shan, the whole quantified shortening is accommodated in the
1372 outer range with no shortening within the basin. To the southwest, in the Kepingtage belt
1373 (transect A), the ~36 km of total shortening are a minimum estimate due to the eroded front of
1374 the cover thrust sheets, which accommodate 100% of the deformation. To the northeast, in the
1375 Kuqa belt (transect C), ~19 km (~86.4% of the ~22 km of total shortening there) are
1376 accommodated by the basement-involved thrust slices and ~3 km (~13.6%) by the cover
1377 duplexes.

1378 According to these estimates, the shortening accumulated within the Southwestern Tian
1379 Shan thrust wedge outward from the inner range decreases from west to east. The structure and
1380 high topography of the range interior, together with the large flexure of the Southwest and North
1381 depressions, indicate that a large amount of shortening must also be accommodated in this
1382 internal part of the system. However, the absence of stratigraphic markers impedes here the
1383 quantification of this deformation using a structural restoration approach.

1384 In addition to shortening, strike-slip deformation may be accommodated on strike-slip
1385 fault systems in the range (e.g., Yin et al., 1998; Zubovitch et al., 2010). Strike-slip motion may
1386 also have occurred on faults at the margins of the basin, for example on the steep Kalpintagh
1387 fault (Li et al., 2015d). However, our 2D restoration does not allow for quantifying such
1388 movements, which take place outside of the planes of our sections.

1389

1390 **8. Northeast deformation**

1391

1392 To the northeast of the Tarim Basin, the southward-vergent Southeastern Tian Shan
1393 strike-slip system can be divided into two parts with, from the north to the south: (1) the inner

1394 range from the right-lateral Baluntay fault to the right-lateral Korla fault, and (2) the outer range
1395 from the Korla fault to the Lop Nur fault system (Figs. 6, 7D, 10D and S2D).

1396

1397 *8.1. Southeastern Tian Shan inner range*

1398

1399 The inner range is several 10s to ~140 km wide, with typical elevations between ~1000
1400 and ~4800 m. It is composed of exhumed imbricates of the Southern Tian Shan Proterozoic
1401 basement, Paleozoic (Devonian to Permian) plutons, and Paleozoic-Mesozoic (Silurian to
1402 Jurassic) cover (see section 2.4). These terrains are in continuity with the Southwestern Tian
1403 Shan interior further west (Fig. 6). A few intra-mountainous basins filled by Cenozoic sediments,
1404 such as the Korla Basin for example, are also included in this part of the range (Fig. 6). There,
1405 the steep inherited Paleozoic faults have been mainly reactivated as right-lateral strike-slip faults
1406 that delimitate some ridges separating the basins (e.g., XBGMR, 1976, 1985; Yin et al., 1998;
1407 Allen et al., 1994; Li et al., 2016a), a deformation style typical of a transpressive system. In this
1408 system, the major faults appear to cut through the whole crust since their location at the surface
1409 correlates with steps in the Moho at depth (Li et al., 2016a). To the south, this inner range is
1410 separated from the outer range by the Korla fault. However, this strike-slip fault zone does not
1411 always correspond to the northeastern boundary of the Cenozoic series of the Tarim Basin,
1412 which is most often located further south (Fig. 6).

1413 In this part of the Southern Tian Shan, the main stratigraphic change between the
1414 Paleozoic range and the Paleozoic Tarim Basin is localized on the Kuruk Tagh fault system
1415 (Figs. 7D and 10D) (e.g., XBGMR, 1976, 1985; Charvet et al., 2011). Accordingly, this fault
1416 zone is a major inherited structure equivalent to the thrust ramps at the front of the basement
1417 imbricates of the Southwestern Tian Shan outer belt. The Kuruk Tagh fault system thus
1418 corresponds to the eastward extension of the major deformation palaeo-boundary between the
1419 Paleozoic orogenic systems of the Southern Tian Shan to the north and the contemporaneous
1420 Tarim Basin to the south.

1421

1422 *8.2. Southeastern Tian Shan outer range*

1423

1424 At the front of the inner range, the outer range is a few to several 10s of km wide, with
1425 typical elevations between ~800 and ~1000 m. To our knowledge, no cross-section across this
1426 outer range has ever been published. However, like the Altyn Tagh transpressional system
1427 bordering the Tarim Basin to the southeast, the Southeastern Tian Shan strike-slip wedge does
1428 not have any fold-and-thrust belt in its outer part. According to the surface and subsurface data,
1429 the outer range is composed of 100s-of-km-long basement-involved structures made of the Tarim
1430 Proterozoic basement and its Protero-Paleozoic to Cenozoic (Sinian to Quaternary) cover (Figs.
1431 7D and 10D). These structures are related to emergent or blind faults, such as the Lop Nur fault
1432 system. Although very steep to nearly vertical, these faults generally have a north block uplifted
1433 by a few 10s to 100s of m, which indicates a southward vergence (Figs. 7D, 10D and S2D). At
1434 depth, all these structures possibly connect to a deep basal decollement within the basement (e.g.,
1435 Charvet et al., 2011).

1436 Unconformities, together with thickness and facies changes, were documented from
1437 surface observations in the Protero-Paleozoic (Sinian to Devonian) sedimentary cover of the
1438 outer range, such as in the Tarim Basin to the south (e.g., Tang et al., 2008; Lin et al., 2012b; Wu
1439 et al., 2018; Morin et al., 2019). These observations imply that the Lop Nur fault system and the
1440 associated basement-cored uplift, which is known as the Kuruk Tagh uplift, are inherited
1441 Protero-Paleozoic structures of the Tarim Basin reactivated during the Cenozoic. In the
1442 Southeastern Tian Shan outer range, the whole Cenozoic deformation is associated with these
1443 inherited faults, since there is no evidence of younger structures rooting into a Paleozoic or
1444 Cenozoic decollement level in the sedimentary series at their front.

1445 The large basement-involved structure of the outer range now forms a smooth topography
1446 uplifted by a few 100s of m above the present-day Tarim sedimentation level, and localizes the
1447 mountain front (Fig. 6). Located several kilometres further south, the Lop Nur fault system,
1448 which corresponds to the present-day deformation front of the Southeastern Tian Shan range, is
1449 hidden by onlapping Plio-Quaternary deposits (Figs. 6, 7D, 10D and S2D). At the front of this
1450 fault system, there is no evidence of a significant flexural subsidence, with less than ~1 km of
1451 Cenozoic sediments trapped in this part of the basin (Figs. 7D, 10D and S2D) (see also Morin et
1452 al., 2019). There, some inherited Protero-Paleozoic structures also exist, but there is little
1453 evidence of a significant Cenozoic activity on their bounding faults, so that any eventual
1454 Cenozoic deformation in this area would have remained very limited. However, towards the very

1455 eastern end of the Tarim Basin, other likely inherited structures may have been slightly
1456 reactivated (Fig. 6).

1457

1458 *8.3. Deformation amount*

1459

1460 Along the transect crossing the Southeastern Tian Shan strike-slip system, we quantified
1461 the shortening from the main bordering fault of the northeastern Tarim Cenozoic series (i.e. from
1462 the Korla fault in the outer range) to the centre of the basin (see Figs. 6 and 7D for the fixed
1463 reference lines set at the turning points in the centre of section D). Accordingly, the shortening
1464 values given just below do not account for the deformation accommodated within the inner
1465 range. In its outer range only, the Southeastern Tian Shan strike-slip wedge accommodates a
1466 very limited amount of Cenozoic compressive deformation, with a possible total shortening of
1467 less than several 10s of m. However, in the absence of a preserved Mesozoic-Cenozoic
1468 stratigraphic marker, it was not possible to properly quantify the Cenozoic shortening, since
1469 eroded material cannot be reconstructed. Nevertheless, this shortening is entirely accommodated
1470 by basement-involved structures.

1471 Even if the compressive deformation accumulated within the outer range of the
1472 Southeastern Tian Shan strike-slip wedge is poorly constrained, it seems to be limited. Moreover,
1473 the structure and low topography of the inner range suggest that the compressive deformation
1474 accommodated across this internal part of the system should also be limited. Even though the
1475 absence of stratigraphic markers impedes any restoration and quantification of the deformation
1476 within the inner range, the non-existence of a flexure associated with the Southeastern Tian Shan
1477 supports limited shortening and thickening, typical of such a strike-slip wedge. Instead, a large
1478 amount of strike-slip movements has likely been accommodated on major strike-slip faults of the
1479 range, such as the Kuruk Tagh fault system or the Korla fault. Strike-slip motion may also have
1480 occurred in the basin, along the very steep Lop Nur faults for instance. However, 2D restoration
1481 does not allow for quantifying such deformation, which take place outside of the plane of our
1482 section.

1483

1484 **9. Discussion**

1485

1486 *9.1. Neo-formed structures and particular case of the Mazar Tagh thrust sheet*

1487

1488 In the outer part of the ranges surrounding the Tarim Basin, a significant fraction of the
1489 deformation (up to 100%) is accommodated by structures neo-formed during the Cenozoic
1490 (Table 1). However, these neo-formed structures are restricted to the sedimentary cover (thin-
1491 skinned deformation) at the front of the Western Kunlun Range and Southwestern Tian Shan,
1492 where the amount of compressive deformation is most important (Fig. 11). In contrast, neo-
1493 formed structures are absent at the front of the Altyn Tagh or Southeastern Tian Shan
1494 transpressive systems. In addition, the development of these structures is also strongly correlated
1495 with the occurrence of Paleozoic and Cenozoic favorable decollement levels in the sedimentary
1496 cover.

1497 Among these neo-formed structures, the Mazar Tagh thrust sheet, which is located at the
1498 front of the Western Kunlun, has remarkable dimensions (Figs. 6, 7B, 8B and 11). With a north-
1499 south width of ~150-200 km and an east-west length of ~300 km, it could even be the largest
1500 active thrust sheet in the world without internal deformation (see also Wang et al., 2014c, Suppe
1501 et al., 2015; Guilbaud et al., 2017). Consequently, the whole foredeep sedimentary pile (i.e. the
1502 sediments of the Southwest depression) is detached at the base of the Cenozoic series and
1503 transported northwards. The Mazar Tagh thrust sheet is therefore oversized compared to classical
1504 thrust sheets. To the first order, the size of a thrust sheet is controlled by the ratio between its
1505 internal resistance (depending on its thickness) and its basal friction (depending on the nature
1506 and thickness of its basal decollement level) (e.g., Liu et al., 1992; Mandal et al., 1997; Smit et
1507 al., 2003; Hatcher, 2004). Accordingly, the gypsiferous Paleocene-Eocene decollement level
1508 should thus be very efficient (i.e. with a low basal friction) to allow for the development of the
1509 Mazar Tagh thrust sheet (Suppe et al., 2015). However, this is probably not the only explanation
1510 for such an oversized thrust sheet. Indeed, the large thickness (up to ~10 km) of sediments
1511 trapped in the Southwest depression may have also drastically increased the rigidity of this thrust
1512 sheet, leading to the emergence of its frontal ramps far away from the range topographic front,
1513 where flexure, accommodation and sedimentation are reduced (e.g., Mugnier et al., 1997; Fillon
1514 et al., 2013).

1515 Finally, an obvious spatial correlation exists between the location of the Mazar Tagh
1516 emergent thrust ramps rooting into the Cenozoic decollement and the Mazar Tagh deep faults

1517 below it (Fig. 6). In fact, these ramps appear to emerge towards the surface from existing
1518 structural steps in this decollement, caused by the hangingwall uplift of the underlying basement-
1519 cored structures (Figs. S2B, 7B, 8B and 9). Therefore, the deformation of the Cenozoic
1520 decollement by the deep Mazar Tagh thrusts most probably controlled the subsequent
1521 localization of the shallow ramps. Potentially, the Mazar Tagh thrust sheet could have been even
1522 wider if this basement structures did not exist.

1523

1524 *9.2. Structural inheritance*

1525

1526 In the ranges surrounding the Tarim Basin, metamorphic and stratigraphic evidences
1527 indicate that the basement-cored structures (thick-skinned deformation) are inherited from the
1528 orogens that developed around the Tarim block during the Paleozoic (e.g., Windley et al., 1990;
1529 Allen et al., 1993, 1994; Matte et al., 1996; Jolivet et al., 2010, 2013; Glorie et al., 2011;
1530 Macaulay et al., 2014; this study). Within the inner ranges, these inherited structures absorb all
1531 the deformation, while in the outer ranges they accommodate from 0 to 100% of the shortening
1532 (Fig. 11; Table 1). Within the basin, at the front of the Altyn Tagh Range, the Tanan uplift also
1533 corresponds to an inherited structure from the large Paleozoic orogenic system of the southern
1534 margin of the Tarim block. Together with the most frontal basement ramp below the Western
1535 Kunlun outer belt, this structure corresponds to a major structural palaeo-boundary between the
1536 Paleozoic ranges to the south and the contemporaneous Tarim Basin to the north (Fig. 11). To
1537 the north, the structural palaeo-boundary between the Paleozoic Tian Shan and the Tarim Basin
1538 also presents evidence of a Cenozoic activity along its whole length. Accordingly, the Paleozoic
1539 structures all around the Tarim block have been reactivated during the Cenozoic orogeny (Fig.
1540 11).

1541 In the outer part of the Tian Shan and within the basin, structures inherited from the
1542 Protero-Paleozoic history of the Tarim block were also reactivated (e.g., Allen et al., 1999;
1543 Carroll et al., 2001; Turner et al., 2010, 2011; Chang et al., 2012b; this study). The Kalpin and
1544 Kuruk Tagh uplifts are now part of the Southwestern and Southeastern Tian Shan outer ranges,
1545 respectively (Fig. 11). On the other hand, in the western part of the Tarim Basin, the Bachu uplift
1546 was rejuvenated in the foreland of the Western Kunlun (Fig. 11).

1547 The Cenozoic reactivation of all those Paleozoic and Protero-Paleozoic structures
1548 confirms the major role of structural inheritance in continental deformation, as already widely
1549 demonstrated in Asia (e.g., [Molnar et al., 1981](#); [Windley et al., 1990](#); [Matte et al., 1996](#); [De](#)
1550 [Grave et al., 2007](#); [Jolivet et al., 2010](#); [Jolivet et al., 2013](#); [Tang et al., 2014](#)). However, all the
1551 inherited Protero-Paleozoic structures in the basin have not been reactivated. If the Tanan and
1552 Bachu uplifts show evidence of a significant Cenozoic deformation, such is not the case for the
1553 Tazhong, Tadong and Tabei uplifts ([Figs. 6, 7 and 11](#)). The location and orientation of the Tanan
1554 and Bachu structures may have played a decisive role in their reactivation, with a compressive
1555 vs. strike-slip main deformation component similar to the one of the local edge of the Tibetan
1556 Plateau. Indeed, the Tanan uplift, which is probably reactivated with a prevailing strike-slip
1557 component, is oriented parallel to the Altyn Tagh and located just to the front of this range. The
1558 Bachu uplift, which is reactivated with a prevailing compressive component, is oriented parallel
1559 to the Western Kunlun Range and developed where the forebulge should have been located in
1560 the foreland of this orogenic system. In contrast, the Tazhong, Tadong and Tabei uplifts were not
1561 reactivated even though their orientations are comparable to the ones of the surrounding ranges,
1562 probably because they are located further from the Tibetan Plateau edge.

1563 In spite of all the reactivated structures in the Tarim basin and surrounding ranges, the
1564 experimental and numerical models designed to understand the Cenozoic deformation of Asia
1565 usually only focus on the influence of stress boundary conditions and lithospheric rheology, but
1566 neglect structural inheritance (e.g. [Tapponnier et al., 1982](#); [Vilotte et al., 1984](#); [England and](#)
1567 [Houseman, 1985](#); [Davy and Cobbold, 1988](#); [Peltzer and Tapponnier, 1988](#); [Burg et al., 1994](#);
1568 [Neil and Houseman, 1997](#); [Wang et al., 2007](#); [Dayem et al., 2009](#); [Calignano et al., 2015](#)).
1569 Ideally, these various parameters should be all considered in the modelling. In the future, testing
1570 the effect of different stress boundary conditions on the reactivation of the inherited basement
1571 structures in the area could be a step forward, as a comprehensive regional study is now available
1572 to compare with model results.

1573

1574 9.3. *Strike-slip faults and potential deformation partitioning*

1575

1576 This study mainly deals with the compressive deformation in the Tarim region since 2D
1577 restoration does not allow for quantifying strike-slip movements. However, strike-slip

1578 deformation may occur in various places in the ranges (e.g., Wang, 1997; Cowgill et al., 2000,
1579 2003; Sobel, 2001; Yue et al., 2001; Meng et al., 2001; Yin et al., 2002; Gehrels et al., 2003;
1580 Ding et al., 2004; Cowgill, 2010; Cheng et al., 2015, 2016a; Wang et al., 2016b) and in the basin,
1581 such as on the fault systems bounding the Tanan and Bachu uplifts (e.g., Avouac and Peltzer,
1582 1993; Yin et al., 2002; Li et al., 2016e; Yang et al., 2018b; this study). Indeed, the central part of
1583 the Tanan uplift corresponds to a symmetric basement pop-up bounded by steep ($> 50^\circ$) thrusts
1584 (Figs. 6, 7C and 8C), an overall geometry characteristic of strike-slip structures (e.g., Casas et
1585 al., 2001; McClay and Bonora, 2001 and references therein; Woodcock and Rickards, 2003). The
1586 Bachu uplift also corresponds to an asymmetric (to the east) or symmetric (to the west) basement
1587 pop-up with steep ($> 60^\circ$) bounding thrusts, in particular in the case of the Seyibula or Mazar
1588 Tagh deep fault systems (Figs. 6, 7B and 9). Vertical faults may even be observed in these
1589 systems (Li et al., 2016e). The existence of these potential strike-slip structures may indicate
1590 deformation partitioning through the Bachu uplift, with a prevailing compressive component to
1591 the north (on the Aqia-Tumuxiuke thrust systems) and a prevailing strike-slip one to the south
1592 (on the Seyibula-Mazar Tagh deep fault systems) (Fig. 6). As for the structure orientation, this
1593 configuration is similar to the one of the Western Kunlun Range where deformation is
1594 partitioned between compressive structures to the front, and the Kashgar-Yecheng transfer
1595 system and the Karakax strike-slip fault to the rear.

1596 Additionally, a few studies proposed a prolongation of the Talas Fergana strike-slip fault
1597 (Shache-Yangdaman fault) inside the Tarim Basin, in front of the Pamir Range, southeastward
1598 down to the city of Yecheng (Wei et al., 2013; Blayney et al., 2016, 2019; Bande et al., 2017b).
1599 However, we find no evidence of deformation associated with such a fault prolongation, neither
1600 at the surface, nor on the seismic data in spite of some available seismic profiles crossing the
1601 supposed fault location (Figs. 1, S2A, 7A and 8A). Indeed, the Shache basement fault below the
1602 Jiede anticline nearby Yecheng has a limited lateral extension (Fig. 6), since it is not observed on
1603 the seismic lines available laterally. Consequently, this structure cannot be a prolongation of the
1604 Talas Fergana fault. A horsetail splay termination of this fault into the Southwestern Tian Shan
1605 thrusts, as suggested by Jia et al., (2015), seems to be a more reasonable interpretation, with
1606 regards to our dataset in the basin.

1607 Finally, the sharp eastern edges of the Western Kunlun Range and Bachu uplift in map
1608 view could suggest the occurrence of a major NNE-SSW strike-slip fault linking both

1609 terminations. However, we did not observe any evidence for such a fault in the numerous seismic
1610 lines crossing the area (Fig. 1). The eastern extremities of the Western Kunlun Range and Bachu
1611 uplift rather correspond to a progressive, but locally abrupt, eastward decline of their structures
1612 (Fig. 6).

1613

1614 9.4. Shortening

1615

1616 In the basin and the outer part of the surrounding ranges, we quantified the Cenozoic
1617 shortening from structural restorations (Fig. 7; Table 1). However, the associated estimates
1618 correspond to the deformation accommodated from the margins of the basin to its centre, and
1619 therefore consist in minimum values with respect to the total shortening accumulated in the
1620 ranges. Indeed, our restorations do not account for the shortening absorbed across the inner
1621 ranges. In order to assess the total deformation across the ranges, we used a crustal budget
1622 approach (for the methodology see Fig. 12A and 12B, as well as the Supplementary information
1623 S5 and references therein). This approach is based on the conservation of the crustal matter
1624 during deformation. Practically, it corresponds to an area balancing of the crust of the ranges
1625 between the present-day and initial states of deformation (e.g., Ménard et al., 1991; Avouac et
1626 al., 1993; Teixell, 1998). Eventually, the compressive deformation across the inner ranges can be
1627 inferred from the difference between the shortening estimates across the entire and the outer
1628 ranges (Table 1).

1629 In the Western Kunlun compressive system, we estimate a total crustal shortening of
1630 $\sim 49.5 +23/-17.5$ km to the west (transect A) and $\sim 83 \pm 23$ km to the east (transect B) (Fig. S6A1
1631 and S6B1; Table 1). This compressive deformation is essentially accommodated within the entire
1632 (inner and outer) range when compared to the basin, with $\sim 99\%$ ($49 +23/-17.5$ km) and $\sim 94\%$
1633 (78 ± 23 km) of this total shortening concentrated across the range for transects A and B,
1634 respectively (Table 1). These estimates are within the wide range of values, from ~ 42 to ~ 272
1635 km, already suggested for this orogen (Lyon-Caen and Molnar, 1984; Rumelhart et al., 1999;
1636 Cowgill et al., 2003). However, the highest of these suggested shortening values are based on the
1637 probably oversimplified assumption that the Pamir-Western Kunlun Range was initially straight
1638 and subsequently acquired its present-day bend during the Cenozoic orogeny (see Chen et al.,
1639 2018 and references therein). In the inner range, $\sim 35.4\%$ ($17.5 +23/-17.5$ km) and $\sim 57.8\%$ ($48 \pm$

1640 23 km) of the total shortening is absorbed to the west (transect A) and to the east (transect B),
1641 respectively (Table 1). In addition, ~63.6% (~31.5 km) and ~36.2% (~30 km) of the total
1642 shortening are accumulated within the outer range to the west (transects A) and to the east
1643 (transect B) (Fig. 7A and 7B; Table 1). The different shortening amounts absorbed across the
1644 inner range, when the shortening amount within the outer range is similar, could thus reflect
1645 varying proportions of strike-slip vs. compressive deformation in the interior of the mountain
1646 belt, with more strike slip deformation in its NNW-SSE part than in its WNW-ESE one. In the
1647 Tarim Basin, a significant amount of deformation is also accommodated by the Bachu uplift that
1648 absorbs ~6% (~5 km) of the total compressive deformation in the Western Kunlun system along
1649 transect B (Fig. 7B; Table 1).

1650 Across the Altyn Tagh transpressive system, we estimate a total crustal shortening of
1651 $\sim 38.6 \pm 18$ km to the west (transect C) and 24 ± 9 km to the east (transect D) (Fig. S6C1 and
1652 6D1; Table 1). Here again, this compressive deformation is essentially accommodated within the
1653 entire (inner and outer) range when compared to the basin, with ~98.5% (38 ± 18.0 km) and
1654 100% (24 ± 9 km) of this total shortening accommodated in the range for transects C and D,
1655 respectively (Table 1). The main part of this compressive deformation is absorbed within the
1656 inner range, with ~97.7% (37.7 ± 18 km) and ~98.8% (23.7 ± 9 km) concentrated across
1657 transects C and D, respectively (Table 1). The outer range only accumulates ~0.8 to ~1.2% (~0.3
1658 km) of the total shortening (Fig. 7C and 7D; Table 1). To the west, a limited amount of
1659 deformation is also accommodated within the basin where the Tanan uplift accommodates
1660 ~1.5% (~0.6 km) of the total compressive deformation across the Altyn Tagh system (Fig. 7C;
1661 Table 1). To our knowledge, these values are the first shortening estimates provided for this
1662 strike-slip wedge.

1663 Across the Southwestern Tian Shan compressive system, we estimate a total crustal
1664 shortening of $54 +24/-18$ km to the west (transect A), 53 ± 23 km in the centre (transect B), and
1665 44 ± 15 km to the east (transect C) (Fig. S6A2, S6B2 and S6C2; Table 1). This compressive
1666 deformation is totally accommodated within the range when compared to the basin (Table 1).
1667 Many other shortening values, ranging from ~80 to ~280 km, have been proposed for the Tian
1668 Shan (e.g., Molnar and Tapponnier, 1975; Chen et al., 1991; Avouac et al., 1993; Allen et al.,
1669 1994). However, these values correspond to estimates of shortening across the whole range
1670 (Northern, Central and Southern Tian Shan), and cannot be therefore easily compared with our

1671 results. In the inner range, ~33.3% (18 +24/-18 km) and ~50% (22 ± 15 km) of the total
1672 shortening are accumulated to the west (transect A) and to east (transect C), respectively (Table
1673 1). In the outer range, at least ~66.7% (~36 km) and ~50% (~22 km) of the compressive
1674 deformation is absorbed across the Kepingtage and Kuqa belts, respectively, to the west (transect
1675 A) and to the east (transect C) of the Southwestern Tian Shan (Fig. 7A and 7C; Table 1).

1676 In the Southeastern Tian Shan transpressive system, we estimate a total crustal shortening
1677 of 15 +20/-15 km (transect D) that may be entirely accommodated in the inner range (Fig. S6D2;
1678 Table 1). Indeed, in the outer range, the compressive deformation seems to be negligible, with
1679 less than several 10s of m of shortening (Fig. 7D; Table 1). To our knowledge, these values are
1680 the first estimates of shortening provided for the Southeastern Tian Shan strike-slip system.

1681 In this study, we do not have enough constraints to quantify the shortening on each side
1682 of the Paleozoic sutures in the ranges surrounding the Tarim Basin, and thus to isolate the exact
1683 contribution of compressive deformation affecting the original Proterozoic Tarim block.
1684 However, based on our shortening assessments, we find that several 10s of km of Cenozoic
1685 compressive deformation have been accommodated on the margins of the Tarim block.
1686 Nevertheless, this deformation is lower in the eastern strike-slip systems (Altyn Tagh and
1687 Southeastern Tian Shan) than in the western compressive ones (Western Kunlun and
1688 Southwestern Tian Shan). Accordingly, regarding this geological evidence and the quantified
1689 shortening, a continental crust dipping down to ~80 km below the Western Kunlun Range (e.g.,
1690 Lyon-Caen and Molnar, 1984; Wittlinger et al., 2004; this study) is plausible. In contrast, a
1691 continental subduction implying a crust dipping down to ~100 km below the Altyn Tagh Range
1692 (Wittlinger et al., 1998) seems unlikely given the limited number of compressive structures and
1693 the associated very low shortening.

1694 Finally, in addition to the Cenozoic deformation on the Tarim block margins, some
1695 basement-involved deformation also occurs within the western and southern parts of this block.
1696 Indeed, in spite of an overall flat topography due to an important sediment filling, a non-
1697 negligible amount of compressive deformation (up to a few % of the total shortening) is also
1698 accommodated within the Tarim Basin across the Bachu and Tanan uplifts. Accordingly, the
1699 Tarim block, including its central part occupied by the present-day Tarim Basin, does not
1700 correspond to a totally rigid and undeformable heterogeneity within the Asian compressive
1701 system.

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9.5. Crustal structure and implications for the deformation of Central Asia

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In order to propose a crustal-scale cross-section across the Tarim Basin and the surrounding mountain ranges, we combined our geological transects with Moho depths inferred from the deep geophysical data (gravimetry, tomography, seismology, rheology, etc.) available in the area, the basin geometry and isostatic calculations (Figs. 12 and S6; Supplementary information S5 and references therein). Below the ranges (e.g., below the Western Kunlun and Southern Tian Shan), we connected the major faults (e.g., the Karakax fault and Southern Tian Shan fault system) with well-documented offsets of the Moho (e.g., Jiang et al., 2004; Wittlinger et al., 2004; Li et al., 2007). However, in the centre of the basin, the deep geophysical data do not precisely constrain the Moho geometry below the Bachu and Tanan basement-cored uplifts, raising the question of where they root at depth (Fig. 12C). The involvement of the basement in these structures and their large wavelength (respectively ~150 km and ~50 km of width) suggest that they are large-scale structures that may be deeply rooted into the middle crust, the base of the crust or into the mantle (Fig. 13).

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Such large foreland basement-cored uplifts exist in other orogenic systems (e.g., Himalayas, Laramides, Andes, Alps, etc.), where their rooting depth has also been discussed and investigated (e.g., Lacombe and Mouthereau, 2002; Yin et al., 2010; Yeck et al., 2014; Yonkee and Weil, 2015; Lacombe and Bellahsen, 2016). Usually, these structures are interpreted as developing above a mid-crustal decollement level (e.g., Lacombe and Mouthereau, 2002; Yeck et al., 2014; Yonkee and Weil, 2015; Lacombe and Bellahsen, 2016). Based on these other case examples and on the rheology of the local continental lithosphere (Fig. 12C), we thus propose that the Bachu and Tanan uplifts also root into a decollement level within the crust (Fig. 13A and 13B). Indeed, some geophysical interpretations (yield strength profiles and earthquake frequency at depth) suggest the existence of two possible decollement levels below the Tarim Basin, in the middle and at the base of the crust (Fig. 12C) (Sloan et al., 2011; Zhang et al., 2013b; Liu et al., 2015b; Chang et al., 2016b; Deng and Tesauero, 2016). The Bachu and Tanan uplifts may therefore deform above one of these crustal decollements (Fig. 13A and 13B). The main northward structural vergency of these basement highs would then indicate that the activated

1732 decollement transfers deformation from the northern edge of the Tibetan Plateau to the centre of
1733 the basin.

1734 However, in the Himalayan system, the Shillong Plateau is a foreland basement-cored
1735 uplift that may root into the mantle, as suggested by underlying possible Moho offsets (Yin et al.,
1736 2010). To investigate this possibility for the Bachu and Tanan basement uplifts, new deep
1737 geophysical data are necessary to specifically constrain the local Moho geometry. Indeed,
1738 accurate geophysical images of these areas are still lacking (e.g., Fig. 3 of Wittlinger et al.,
1739 2004). Nevertheless, rooting these structures directly into the mantle raises the problem of the
1740 foreland-ward transfer of the deformation associated with the Moho offset imaged below the
1741 outer range of the Western Kunlun (Wittlinger et al., 2004) (Fig. 13C). Consequently, we favor
1742 our previous interpretations of a crustal decollement level below the Tarim.

1743 These interpretations imply a decoupling between the deforming crust and an underlying
1744 rigid mantle. This is in agreement with the tomographic studies proposing a cold and rigid
1745 mantle below the basin (Deng and Tesauro, 2016; Deng et al., 2017). Moreover, the deformation
1746 pattern of the Tarim block can be compared with results of experimental models studying the
1747 influence of the depth of a rigid lithospheric heterogeneity on continental deformation
1748 (Calignano et al., 2015). Indeed, the deformation localization around the Tarim Basin and the
1749 associated flexures are in agreement with models simulating a strong upper mantle (Figs. 5, 8c
1750 and 9b of Calignano et al., 2015). Therefore, the geological, geophysical and experimental
1751 observations all suggest that the rigidity of the Tarim block is supported by the mantle, while
1752 deformation is transmitted above a crustal decollement below the Tarim Basin, from the Tibetan
1753 Plateau to the Tian Shan and further north.

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1755 **10. Conclusion**

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1757 On the basis of a structural map and of four balanced geological transects, we
1758 characterized the Cenozoic deformation across the Tarim Basin and the surrounding ranges.
1759 From structural restorations and crustal budgets, we also quantified the compressive component
1760 of this deformation. This comprehensive study thus provides a consistent regional overview of
1761 how the Tarim block participates to the Cenozoic deformation of Asia.

1762 To the west, up to 83 ± 23 and $54 +24/-18$ km of total crustal shortening is
1763 accommodated across the Western Kunlun and Southwestern Tian Shan compressive systems,
1764 respectively. On the other hand, to the east, only up to 38 ± 18 and $15 +20/-15$ km of total crustal
1765 shortening is accommodated across the Altyn Tagh and Southeastern Tian Shan transpressive
1766 systems, respectively. The crustal shortening is therefore higher in the thrust wedges to the west,
1767 than in the strike-slip ones to the east. Accordingly, continental subduction is plausible below the
1768 Western Kunlun Range, whereas it seems unlikely below the Altyn Tagh.

1769 Most of this Cenozoic compressive deformation of the Tarim block (from ~94% to
1770 100%) is concentrated within the surrounding ranges. Indeed, a shortening of up to 78 ± 23 km
1771 (with ~30 km in the outer range only) is accommodated in the Western Kunlun Range, $54 +24/-$
1772 18 km (~36 in the outer range) in the Southwestern Tian Shan, 38 ± 18 km (~0.3 in the outer
1773 range) in the Altyn Tagh Range, and $15 +20/-15$ km (~0 km in the outer range) in the
1774 Southeastern Tian Shan. In spite of an overall flat topography due to an important sediment
1775 filling, a non-negligible amount of compressive deformation is also accommodated within the
1776 Tarim Basin by large basement-cored uplifts with a vergence synthetic to the deformation of the
1777 Tibetan Plateau edge. To the west, the Bachu uplift absorbs ~6% (~5 km) of the total crustal
1778 shortening of the Western Kunlun compressive system. To the south, the Tanan uplift
1779 accumulates ~1.5% (~0.6 km) of the total crustal shortening of the Altyn Tagh transpressive
1780 system. Strike-slip deformation may have also occurred on the bordering faults of these uplifts,
1781 but these displacements still need to be quantified. Consequently, the Tarim block, including its
1782 central part occupied by the present-day Tarim Basin, does not correspond to a totally rigid and
1783 undeformable heterogeneity within the Asian compressive system.

1784 In the ranges around the Tarim Basin, thin-skinned structures neo-formed during the
1785 Cenozoic orogeny accommodate a significant fraction of the compressive deformation (up to
1786 ~66.7%). However, the main part of the deformation (from ~33.3% to 100 %) is rather
1787 accommodated by reactivated thick-skinned structures inherited from the Protero-Paleozoic
1788 history of the Tarim block. The structural inheritance thus appears to be at least as important as
1789 the stress boundary conditions or the lithospheric rheology in controlling the Cenozoic
1790 deformation of the Tarim area.

1791 Finally, in the centre of the basin, the Bachu and Tanan uplifts probably root into a deep
1792 crustal decollement level. This decollement implies decoupling between the deforming crust and

1793 an underlying rigid mantle, in agreement with the geological and geophysical data available in
1794 the region, and in line with experimental observations. Therefore, the rigidity of the Tarim block
1795 seems to be supported by the mantle, while deformation is transmitted above a decollement in
1796 the crust from the Tibetan Plateau to the Tian Shan and further north.

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2802
2803

2804 **Figure captions**

2805
2806 **Figure 1:** Geographical and geological setting of the Tarim area and location of the data used in
2807 this study. Only the main structures outcropping at the surface are represented. References for
2808 the data from previous studies are indicated in the bottom-right insert. Green long-dashed lines
2809 locate the regional balanced geological transects A, B, C and D presented in **Figure 7**. Green
2810 short-dashed lines at terminations of transects B and C correspond to the extended pathway of
2811 the synthetic crustal-scale cross-sections presented in **Figure 13**. ATF: Alтын Tagh Fault, BF:
2812 Baluntay Fault, KFX: Karakax Fault, KGS: Kongur Shan normal fault, NF: Narat Fault, NL:
2813 Nikolaev Line, TFF: Talas Ferghana Fault, VSP: Vertical Seismic Profile. In this study, the
2814 Tarim block refers to the area comprised in between the Paleozoic ophiolitic sutures (thick red
2815 lines) displayed in the surrounding ranges, rather than to the present-day Tarim Basin only.
2816 These sutures located within the Western Kunlun, Alтын Tagh and Southern Tian Shan ranges

2817 correspond to the Oyttag-Kudi, Lapeiquan-Hongliugou, and South Tian Shan sutures,
2818 respectively.

2819

2820 **Figure 2:** Simplified chronostratigraphic chart of the Tarim Basin, Western Kunlun Range and
2821 Southern Tian Shan. Black arrows represent the main decollement levels in the Tarim area, while
2822 colour lines correspond to the seismic horizons picked for this study.

2823

2824 **Figure 3:** Velocity functions used for time-to-depth conversion of the seismic interpretations.
2825 For each sediment package (Sinian-Paleozoic, Permian-Mesozoic and Cenozoic), these velocity
2826 functions are composed of two parts: (1) an upper part (solid line) corresponding to a single
2827 linear regression of the vertical seismic profiles of P-wave data (brown, green and yellow
2828 crosses) from four wells distributed across the Tarim Basin (red dots in Fig. 1), and (2) a lower
2829 part (dashed line) corresponding to a maximum velocity determined according to the main
2830 lithology of the sediment package (calcareous for the Sinian-Paleozoic deposits, terrigenous for
2831 the Permian-Mesozoic and Cenozoic series).

2832

2833 **Figure 4:** Example of surface data and the associated detailed work on the Qimugen fold-thrust
2834 system (see Fig. 1 for location). (A) Satellite image (Landsat images with a 28-m resolution) of
2835 the Qimugen hills. (B) Geological and structural map of the same area constructed from field
2836 data, satellite images and seismic profiles. (C) Field picture of the Qimugen hills taken from the
2837 south. (D) Field picture of the Qimugen hills taken from the southeast. (E) Surface geological
2838 cross-section of the Qimugen hills with a vertical exaggeration (VE) of 2.

2839

2840 **Figure 5:** Example of surface data and the associated detailed work on the eastern part of the
2841 Mazar Tagh emergent thrust ramps and deep fault system (see Fig. 1 for location). (A) Satellite
2842 image (Landsat images with a 28-m resolution) of the Mazar Tagh ridge. (B) Geological and
2843 structural map of the same area constructed from field data, satellite images and seismic profiles.
2844 (C) Field picture of the Mazar Tagh ridge taken from the south. (D) Field picture of the eastern
2845 termination of the Mazar Tagh ridge taken from the northeast. The light-colored deposits to the
2846 right correspond to the Cenozoic (Paleocene-Eocene) evaporitic decollement of the southwestern
2847 Tarim Basin, which crops out at the top of the local Mazar Tagh emergent thrust ramp. (E)

2848 Surface geological cross-section of the eastern termination of the Mazar Tagh ridge with a
2849 vertical exaggeration (VE) of 2 (modified from Pan et al., 2010).

2850

2851 **Figure 6:** Structural map of the Cenozoic deformation of the Tarim Basin and surrounding
2852 ranges. Simplified and abbreviated structure labels are spelled out in the upper right cartoon. (A)
2853 Structures outcropping at the surface only. (B) Structures outcropping at the surface and hidden
2854 below the Cenozoic sediments of the Tarim Basin. Structures rooted in the basement are
2855 represented in red, while structures rooted into a sedimentary decollement are displayed in
2856 orange (for structures developed above Paleozoic decollements) or in yellow (for structures
2857 developed above Cenozoic decollements). Green solid lines locate the regional balanced
2858 geological transects A, B, C and D presented in Figure 7. Green dashed lines at terminations of
2859 transects B and C correspond to the extended pathway of the synthetic crustal-scale cross-
2860 sections presented in Figure 13. In this study, the Tarim block refers to the area comprised
2861 between the Paleozoic ophiolitic sutures (thick black lines) located in the surrounding ranges,
2862 rather than to the present-day Tarim Basin only.

2863

2864 **Figure 7:** Regional balanced geological transects across the Tarim Basin and surrounding ranges
2865 (see Figs. 1 or 6 for location, Fig. 6 for abbreviation labels, and Fig. S3 for a 1:2,000,000-scale
2866 version). Below each transect, a restored state at the beginning of the Cenozoic is presented. For
2867 these restorations, the fixed reference lines (red pins) are located at the turning point in the centre
2868 of sections, and the lighter thrust slices in the range interiors are not considered. Shortening
2869 amounts (black bold numbers) measured on each side of the fixed reference lines are indicated to
2870 the northern and southern ends of the restored transects.

2871

2872 **Figure 8:** Details of the regional balanced geological transects across the southern part of the
2873 Tarim Basin and the Western Kunlun and Altyn Tagh ranges (see Fig. 7 for location and Fig. 6
2874 for abbreviation labels).

2875

2876 **Figure 9:** Details of the regional balanced geological transect B across the central part of the
2877 Tarim Basin (see Fig. 7 for location and Fig. 6 for abbreviation labels).

2878

2879 **Figure 10:** Details of the regional balanced geological transects across the northern part of the
2880 Tarim Basin and the Southern Tian Shan Range (see Fig. 7 for location and Fig. 6 for
2881 abbreviation labels).

2882

2883 **Figure 11:** Tectonic map of the Cenozoic deformation of the Tarim Basin and surrounding
2884 ranges. Structures rooted in the basement are represented in red, while structures rooted into a
2885 sedimentary decollement are displayed in orange (for structures developed above a Paleozoic
2886 decollement) or in yellow (for structures developed above a Cenozoic decollement). Areas
2887 comprising Proterozoic and (Protero-)Paleozoic inherited basement structures reactivated during
2888 the Cenozoic orogeny are indicated in light and dark grey, respectively. Regions with Cenozoic
2889 neo-formed cover structures are shown in orange (for structures developed above a Paleozoic
2890 decollement) or in yellow (for structures developed above a Cenozoic decollement). Green solid
2891 lines locate the regional balanced geological transects A, B, C and D presented in Figure 7.
2892 Green long-dashed lines at terminations of these transects correspond to the extended pathway of
2893 the simplified crustal-scale cross-sections presented in Figure S6, which are used to assess the
2894 shortening accommodated across the ranges from crustal budgets (see Fig. 12 and
2895 Supplementary information S5 for methodology). Green arrows and values at the ends of these
2896 cross-sections represent and scale with the most probable (solid line) and possible maximum
2897 (short-dashed line and bracketed numbers) values for the total Cenozoic shortening estimated
2898 along these transects (Table 1). Ptz: Proterozoic, Pz: Paleozoic, Mz: Mesozoic, Cz: Cenozoic. In
2899 this study, the Tarim block refers to the area comprised between the Paleozoic ophiolitic sutures
2900 (thick black lines) located in the surrounding ranges, rather than to the present-day Tarim Basin
2901 only.

2902

2903 **Figure 12:** Components for the crustal-scale cross-sections presented in Figures 13 and S6 (see
2904 Fig. 6 for abbreviation labels). (A) Example of an elevation profile extracted from the SRTM3
2905 DEM and averaged every 50 km in the range (green dashed line). This average topography
2906 corresponds to the main input for assessing the Moho depth below the ranges by isostatic
2907 calculations. (B) Example of a simplified crustal-scale cross-section used to estimate the
2908 shortening accommodated across the ranges surrounding the Tarim Basin from crustal budgets
2909 (see Supplementary information S5 for methodology). To assess the Moho depth below the

2910 ranges by isostatic calculations, a reference in a stable place (i.e. with no Cenozoic deformation
2911 or flexure) is required. This stable isostatic reference was taken in the centre of the basin,
2912 between kilometres 300 and 350 along transect C, where there is no structures activated during
2913 the Cenozoic (Figs. 7C and S6C1). A_{range} : crustal area below the range, A_{sed} : area of the
2914 Cenozoic sediments coming from the range erosion (i.e. crustal area of the range lost by erosion),
2915 H_0 : initial crustal thickness at the beginning of Cenozoic, L_1 : present-day length of the range, L_0 :
2916 initial crustal length at the beginning of Cenozoic, d : density. (C) Intermediate state of
2917 construction for the synthetic crustal-scale cross-sections. Structures from the surface down to 20
2918 km deep are constrained by the regional balanced geological transects (Fig. 7B). Structures
2919 below 20 km deep are inferred from the isostatic calculations and the basin geometry achieved in
2920 this study, and from the deep geophysical data available in the area (e.g., Li et al., 2002b; Jiang
2921 et al., 2004; Wittlinger et al., 2004). Examples of a lithospheric yield strength profile (from
2922 Zhang et al., 2013b) and of earthquake frequency according to depth (from Sloan et al., 2011)
2923 below the Tarim Basin. ϵ : deformation rate, q : heat flux, n : number of earthquakes.

2924

2925 **Figure 13:** Possible synthetic crustal-scale cross-sections across the Tarim Basin, Western
2926 Kunlun Range and Southern Tian Shan (see Fig. 6 for location and abbreviation labels). (A) and
2927 (B) Cross-sections with the basement-cored uplifts in the basin rooted into a crustal decollement
2928 (in the middle or at the base of the crust, respectively). These interpretations are favoured due to
2929 their consistency with the geological and geophysical data available so far in the area (see text
2930 for more explanations). (C) Alternative cross-section with the large basement-cored uplifts in the
2931 basin rooted into the mantle. In this study, the Tarim block refers to the area comprised between
2932 the Paleozoic ophiolitic sutures (thick black lines) located in the surrounding ranges, rather than
2933 to the present-day Tarim Basin only.

2934

2935 **Table 1:** Shortening estimates across the Tarim Basin and surrounding ranges. Shortening values
2936 are provided for (1) the entire (inner plus outer) ranges, (2) the inner ranges, (3) the outer ranges,
2937 (4) the basin, and (5) the entire orogenic systems. Shortening values for the entire ranges (1) are
2938 derived from crustal budgets (Figs. 12 and S6; Supplementary material S5). Those for the outer
2939 ranges (3) and for the basin (4) are derived from structural restorations (Fig. 7). For the outer
2940 ranges, the contributions to the shortening of the (Protero-)Paleozoic inherited (inh.) and

2941 Cenozoic neo-formed (neo.) structures are indicated in brackets. Shortening values for the inner
2942 ranges (2) correspond to the differences between the shortenings across the entire ranges (1) and
2943 those accumulated in their outer parts (3). Finally, the total shortening amounts accommodated
2944 across the whole orogenic systems (5) are obtained by adding the shortening values
2945 accommodated across the entire ranges (1) and in the basin (4). Values given for shortenings
2946 derived from crustal budgets correspond to the most probable values according to our regional
2947 balanced geological transects (Cenozoic sediment thicknesses) and other published parameters
2948 (densities and sediment porosity) used for calculations. Uncertainties given for these values
2949 correspond to possible errors on the crustal areas below the range (coming from the densities)
2950 and on the crustal areas lost by erosion (coming from the sediment porosity) (see [Supplementary](#)
2951 [information S5](#) for more explanations).

2952

2953

2954 **Supplementary materials**

2955

2956 **Figure S1:** Georeferenced structural map (in GeoTIFF format) of the Cenozoic deformation of
2957 the Tarim Basin and surrounding ranges (see [Fig. 6](#) for abbreviation labels). All structures are
2958 represented, whether they are outcropping at the surface or hidden below the Cenozoic sediments
2959 of the basin. Structures rooted in the basement are represented in red, while structures rooted into
2960 a sedimentary decollement are displayed in orange (for structures developed above a Paleozoic
2961 decollement) or in yellow (for structures developed above a Cenozoic decollement).

2962

2963 **Figure S2:** Examples of seismic profiles and interpretations used to construct the regional
2964 balanced geological transects presented in [Figures 7, 8, 9, 10, and S3](#) (see [Fig. 6](#) for abbreviation
2965 labels). The location of these profiles is given in the top map insert (same legend as in [Fig. 1](#)).

2966

2967 **Figure S3:** Regional balanced geological transects across the Tarim Basin and surrounding
2968 ranges at a scale of 1:2,000,000 (see [Fig. 6](#) for location and abbreviation labels). The main data
2969 used to construct these cross-sections are specified above each transect. References are listed in
2970 [Supplementary information S4](#).

2971

2972 **Information S4:** Discussion about some structural interpretation issues.

2973

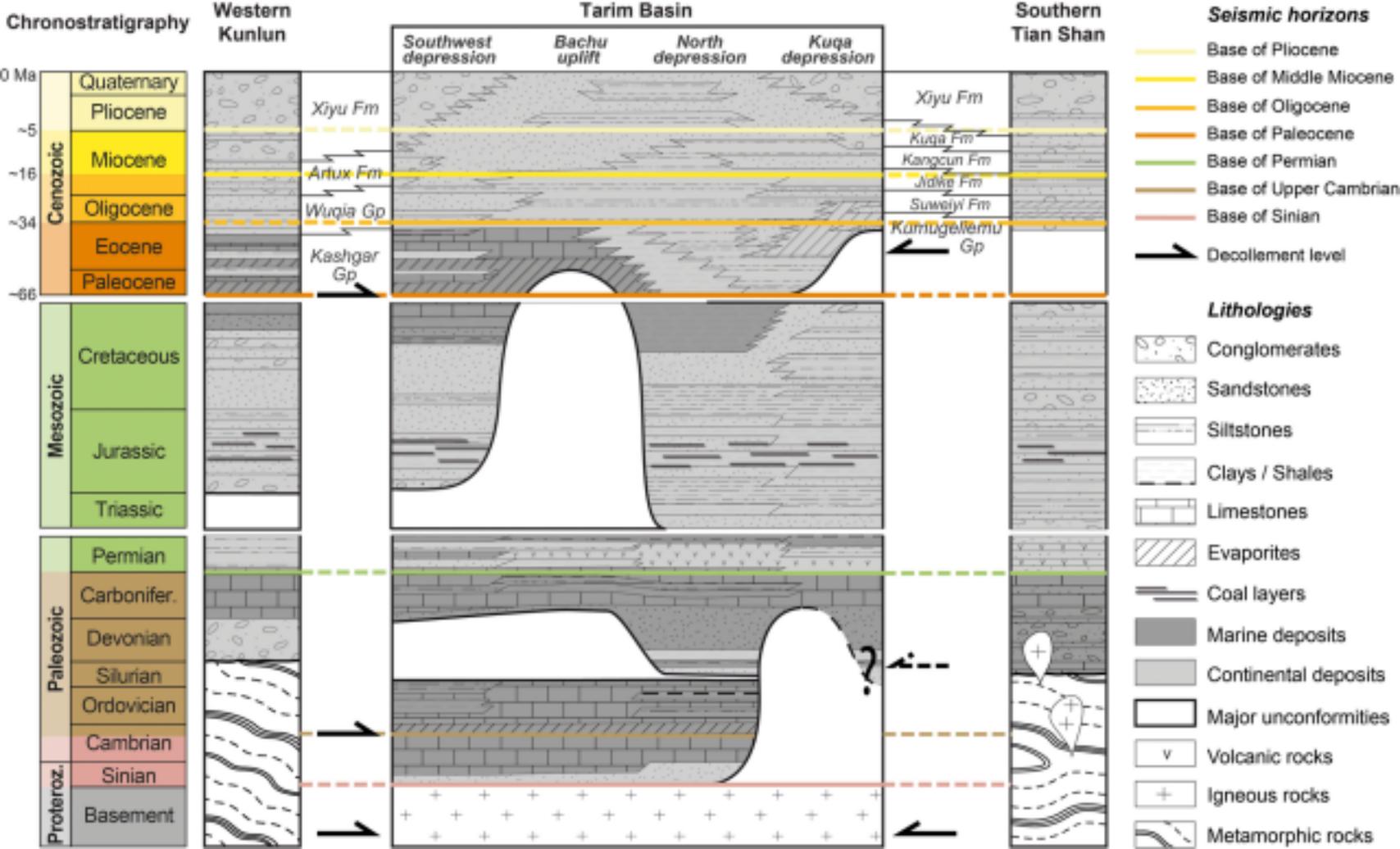
2974 **Information S5:** Methodology used to assess the shortening accommodated in the ranges
2975 surrounding the Tarim Basin from crustal budgets.

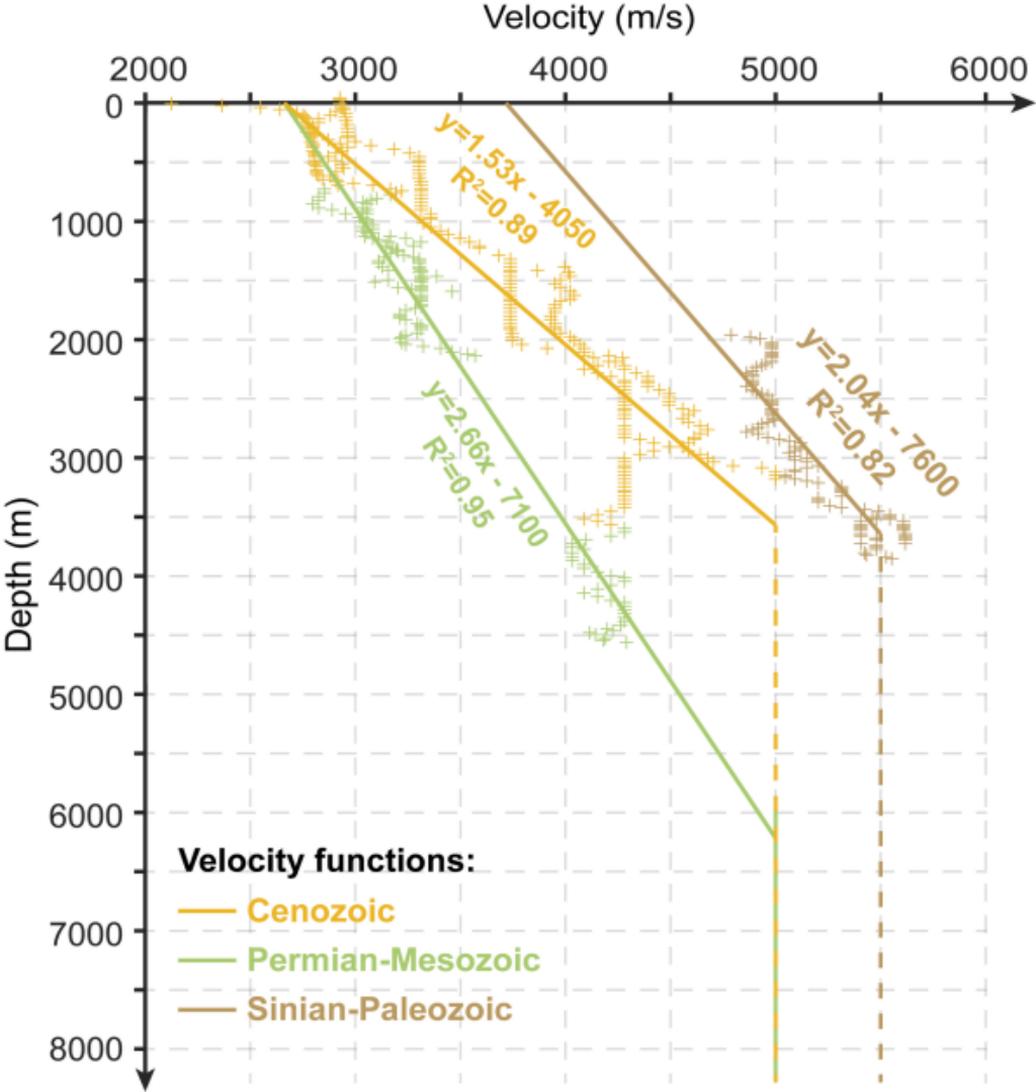
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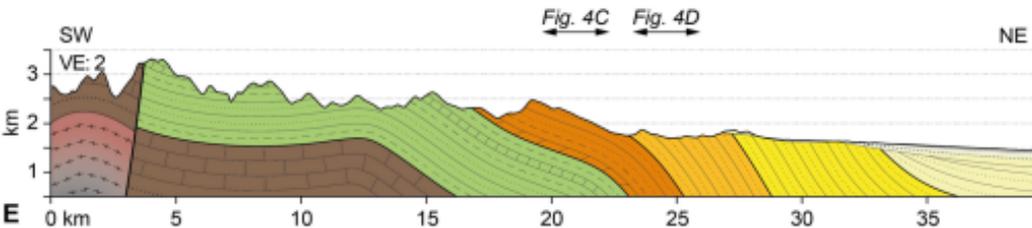
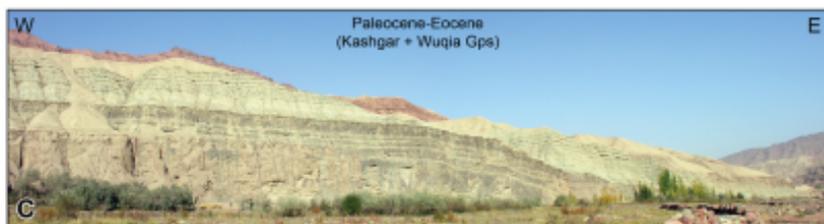
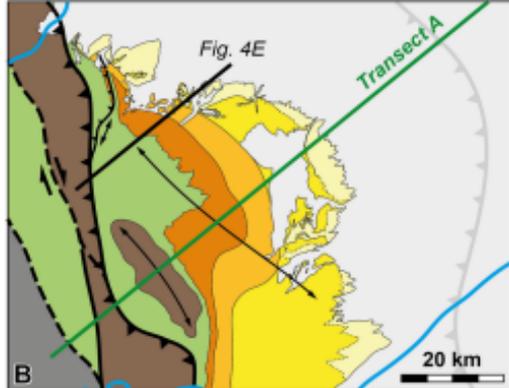
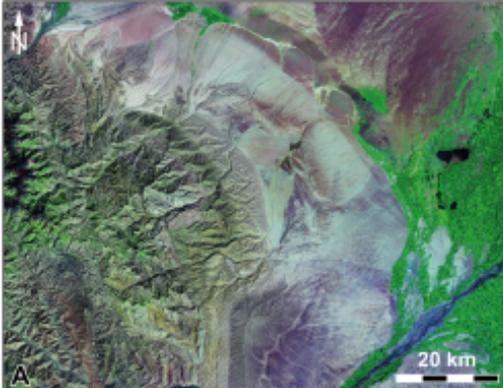
2977 **Figure S6:** Simplified crustal-scale cross-sections used to assess the shortening accommodated
2978 across the ranges surrounding the Tarim Basin from crustal budgets (see Fig. 12 and
2979 **Supplementary information S5** for methodology). Green solid lines in the top-centre map insert
2980 (same legend as in Fig. 11) locate these cross-sections, green dashed lines in between correspond
2981 to the initial pathway of the regional balanced geological transects (see also Figs. 1, 6 and 11)
2982 presented in **Figures 7 and S3**. Orange rectangles on the cross-sections correspond to the Moho
2983 locations published from deep geophysical data, with the associated references quoted to the
2984 right or to the left. Orange dashed rectangles are the out-of-trend Moho locations with the
2985 associated references marked with an asterisk. All these references are listed in **Supplementary**
2986 **information S5**. A_{range} : crustal area below the range, A_{sed} : area of the Cenozoic sediments coming
2987 from the range erosion (i.e. crustal area of the range lost by erosion), H_0 : initial crustal thickness
2988 at the beginning of Cenozoic, L_1 : present-day length of the range, L_0 : initial crustal length at the
2989 beginning of Cenozoic. In this study, the Tarim block refers to the area comprised between the
2990 Paleozoic ophiolitic sutures located in the surrounding ranges, rather than to the present-day
2991 Tarim Basin only.

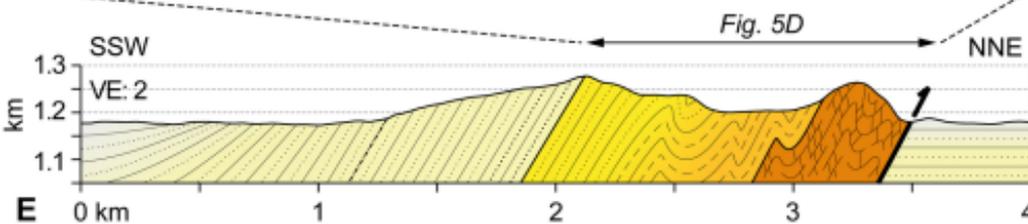
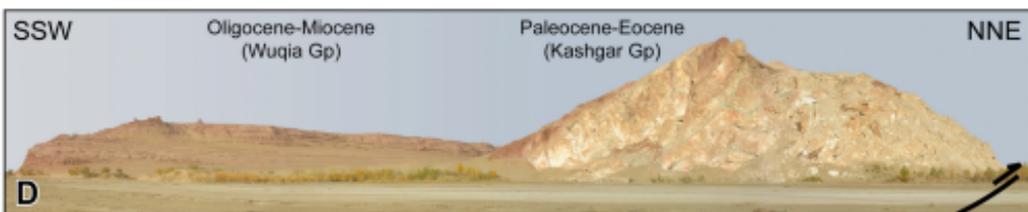
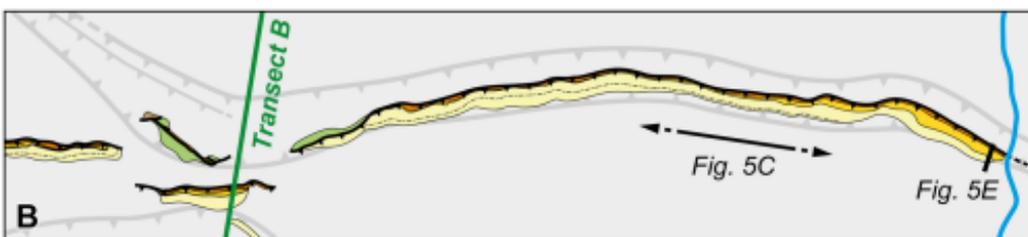
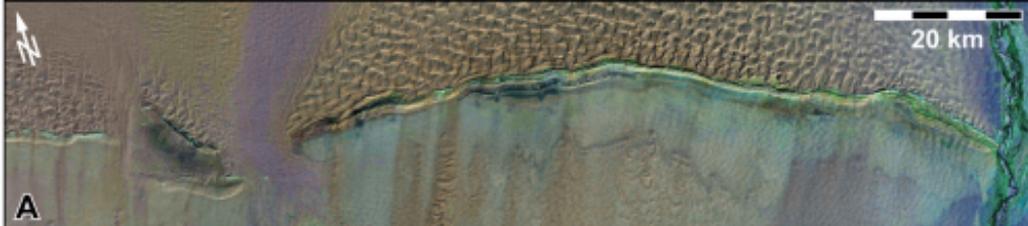
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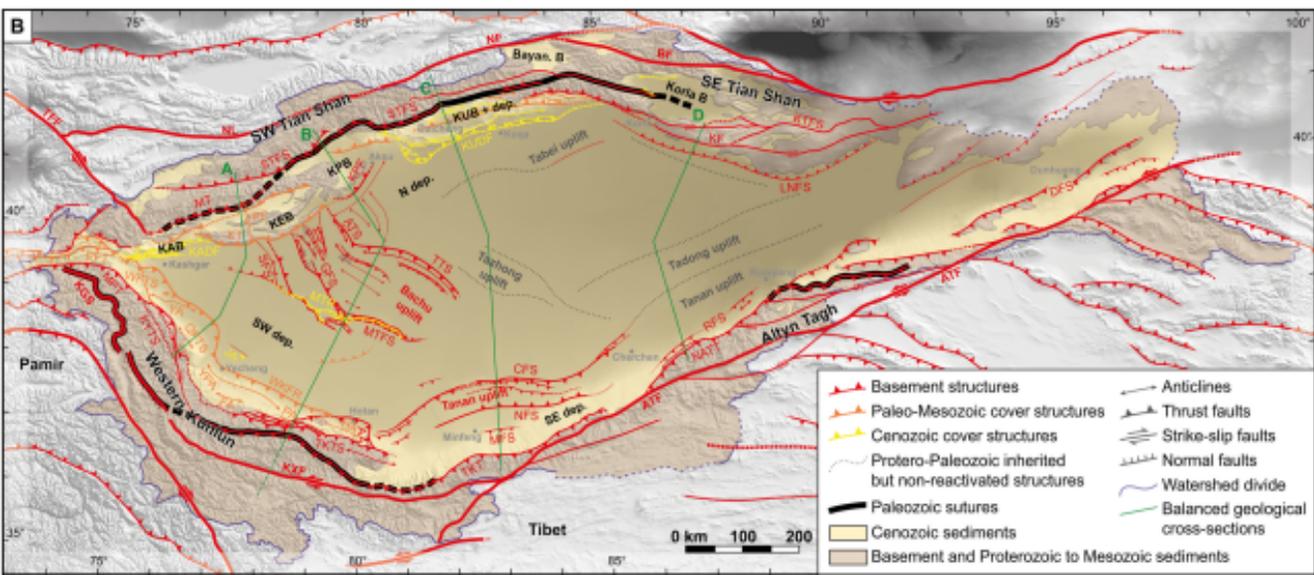
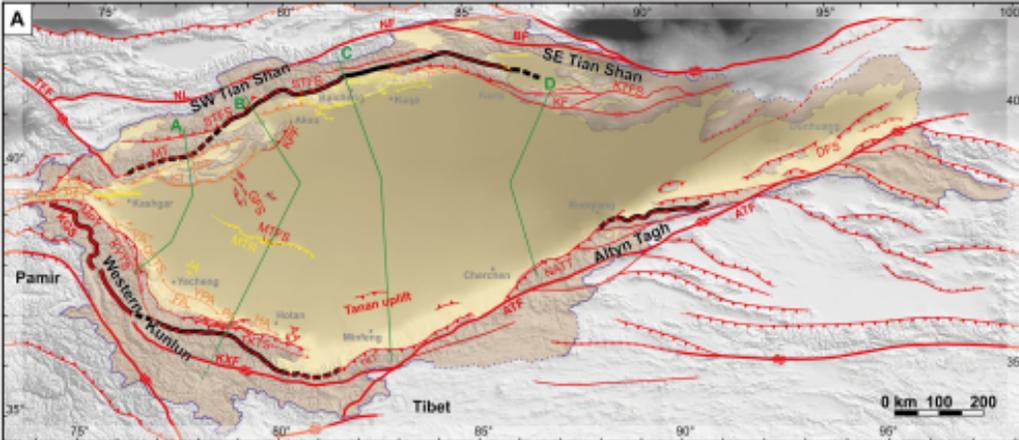
2993











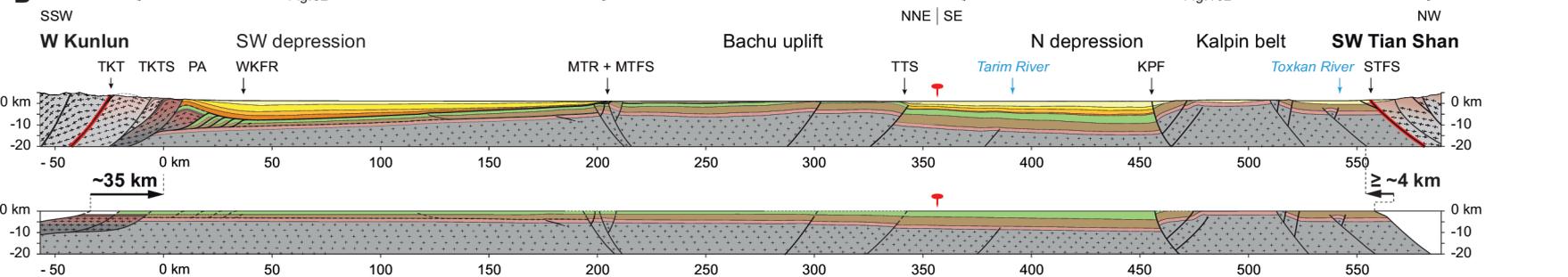
Western Kunlun / Altyn Tagh ranges

-  Devonian to Carboniferous sediments
-  Sinian to Silurian meta-sediments
-  Proterozoic basement

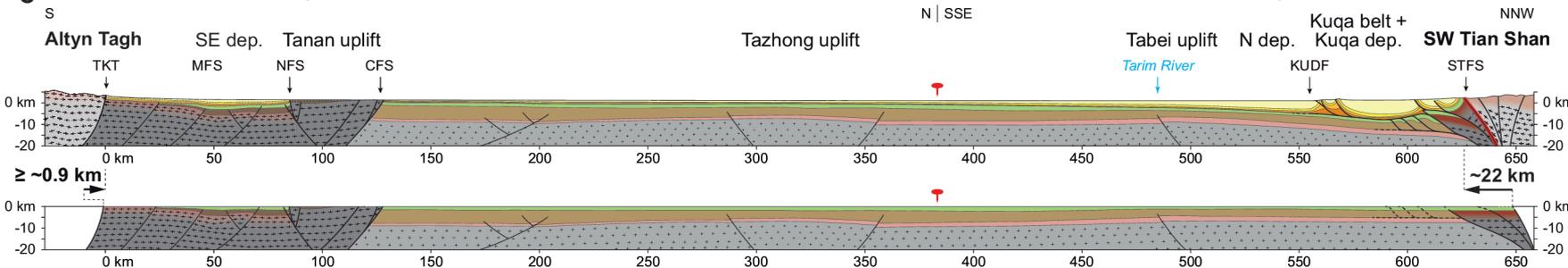
Southern Tian Shan

-  Proterozoic and Paleozoic basement and sediments
-  Paleozoic sutures

B



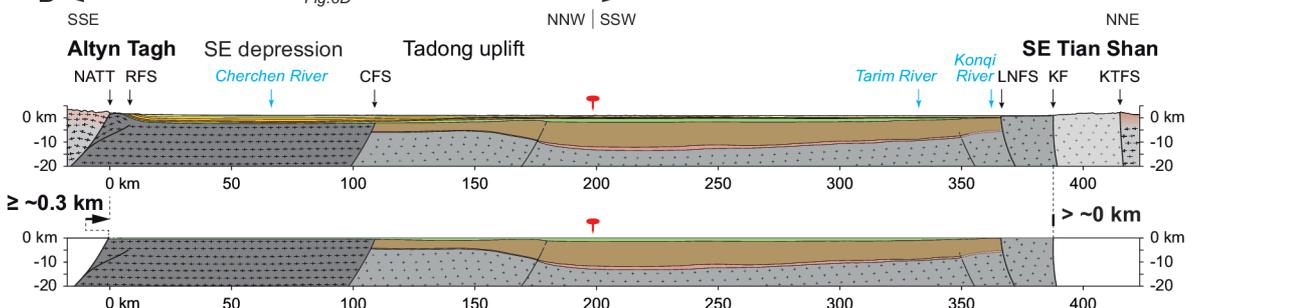
C



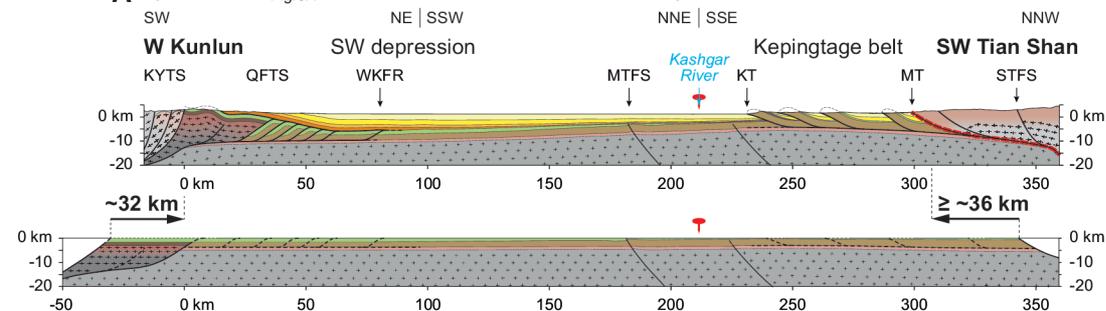
Tarim Basin

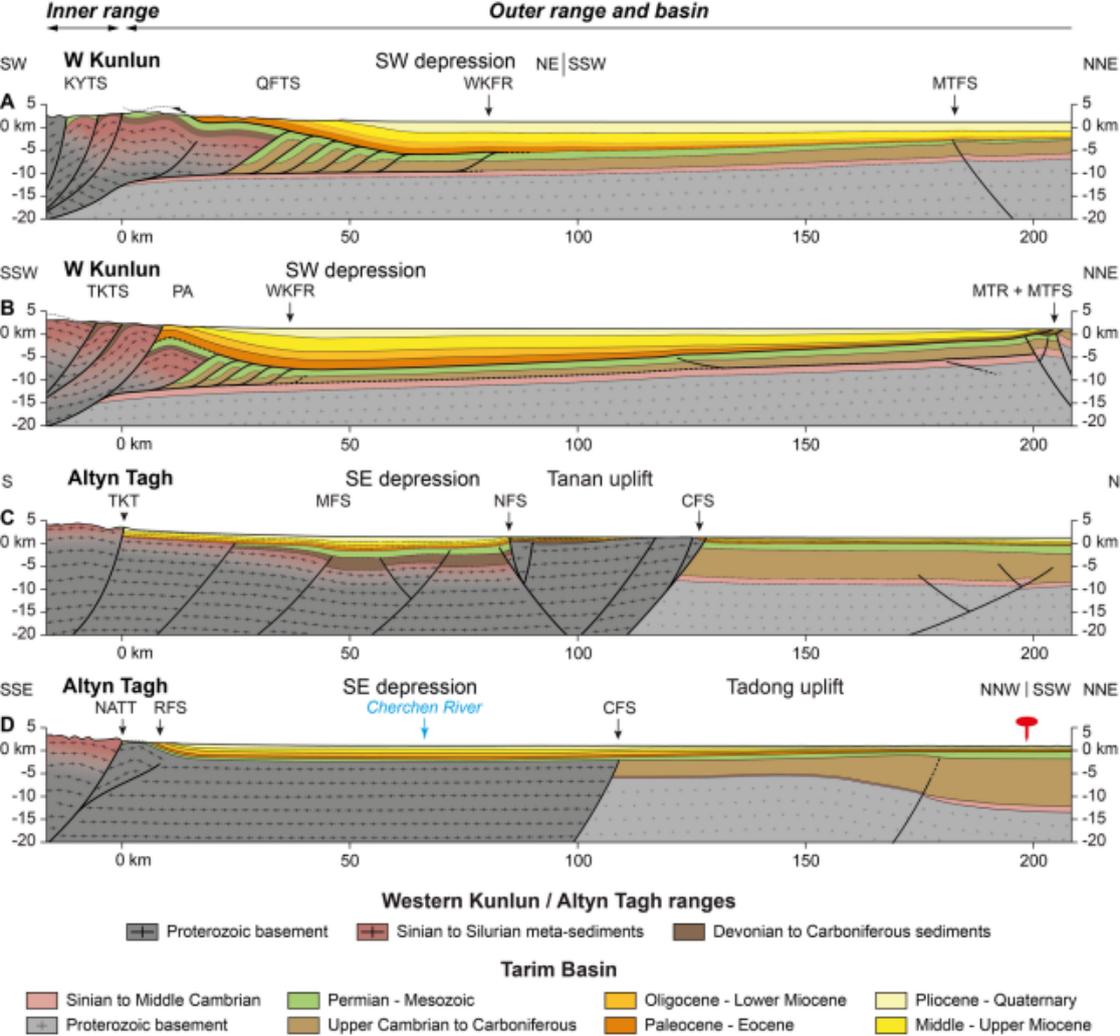
-  Pliocene - Quaternary
-  Middle - Upper Miocene
-  Oligocene - Lower Miocene
-  Paleocene - Eocene
-  Permian - Mesozoic
-  Upper Cambrian to Carboniferous
-  Sinian to Middle Cambrian
-  Proterozoic basement

D



A ← Fig. 8A → Fig. 10A →



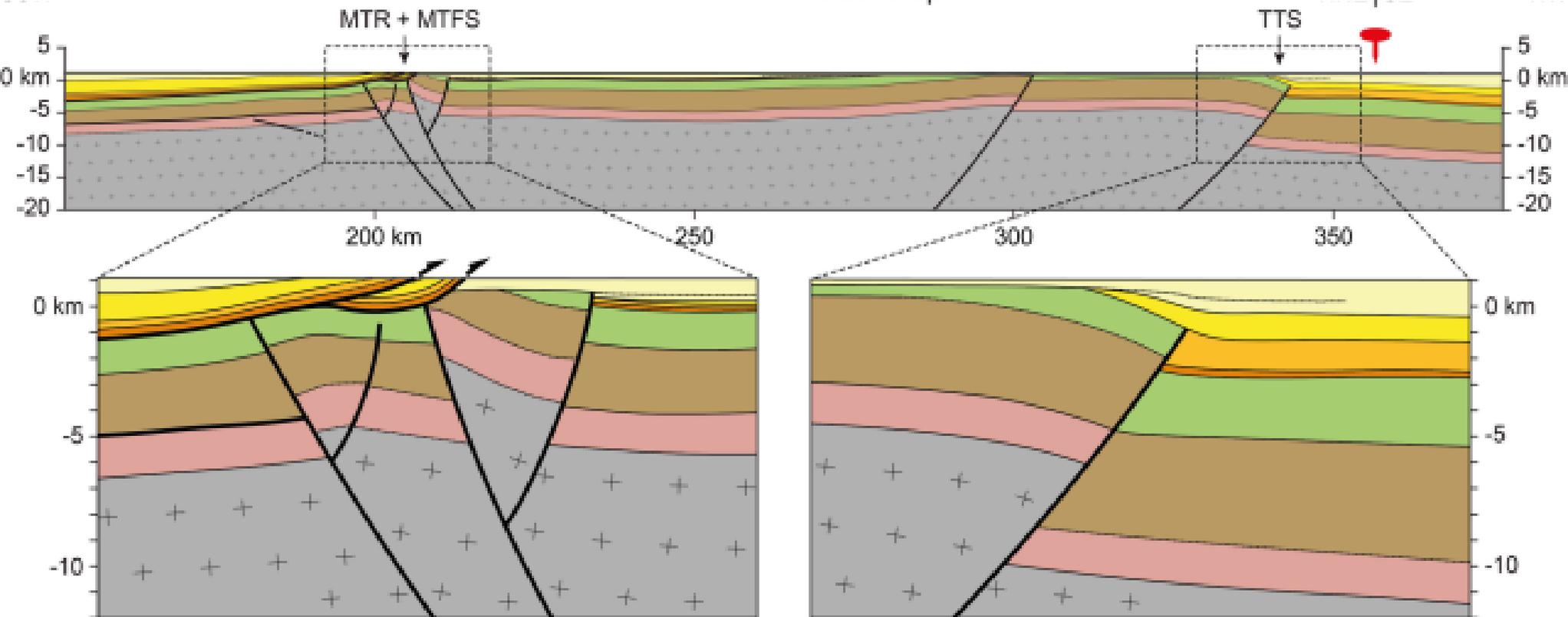


SSW

Bachu uplift

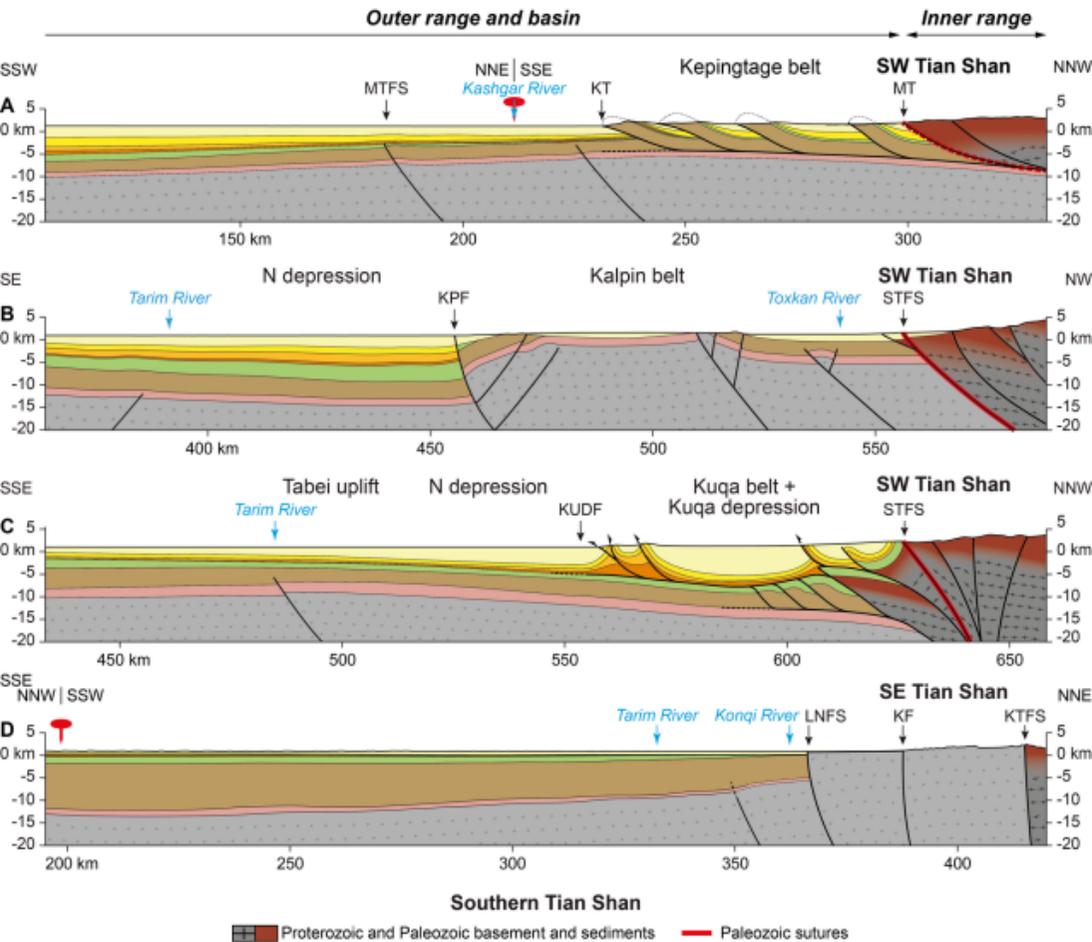
NNE | SE

NW



Tarim Basin



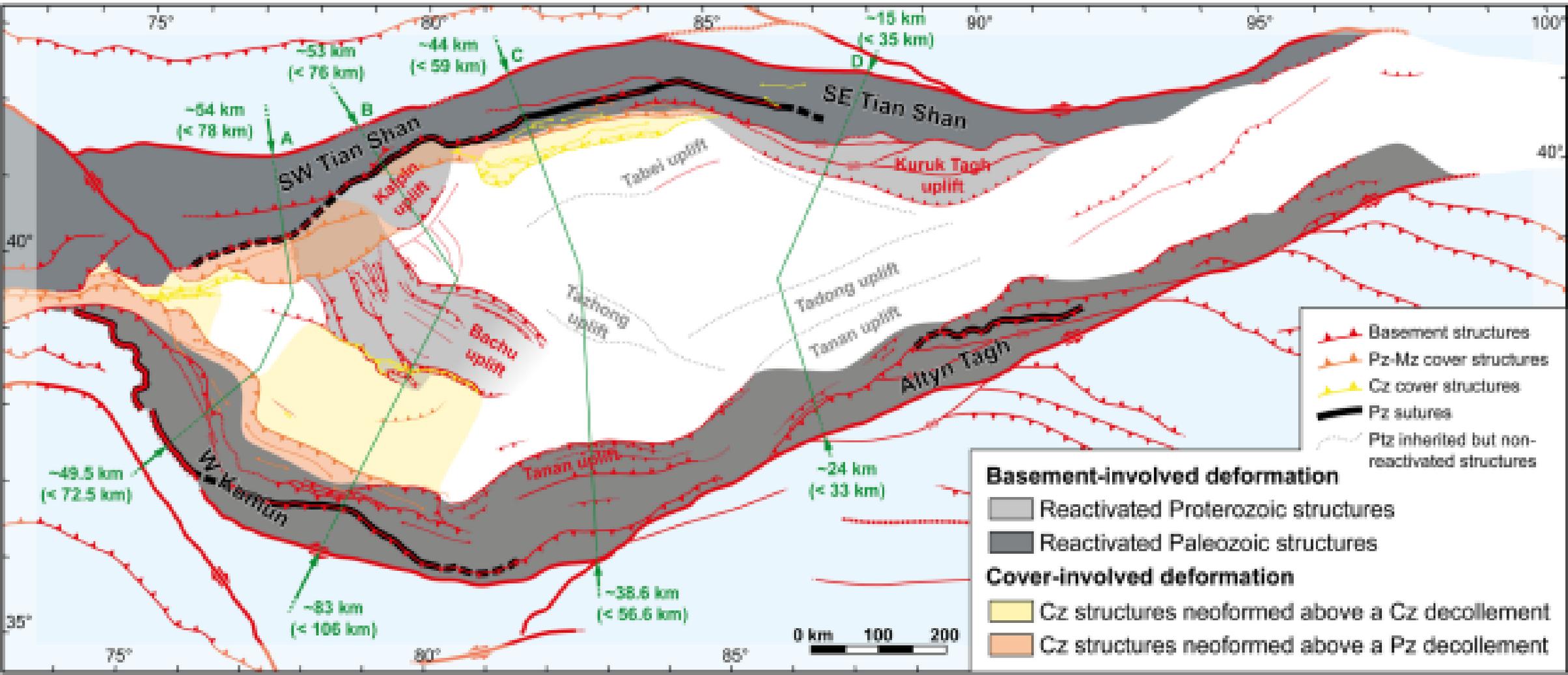


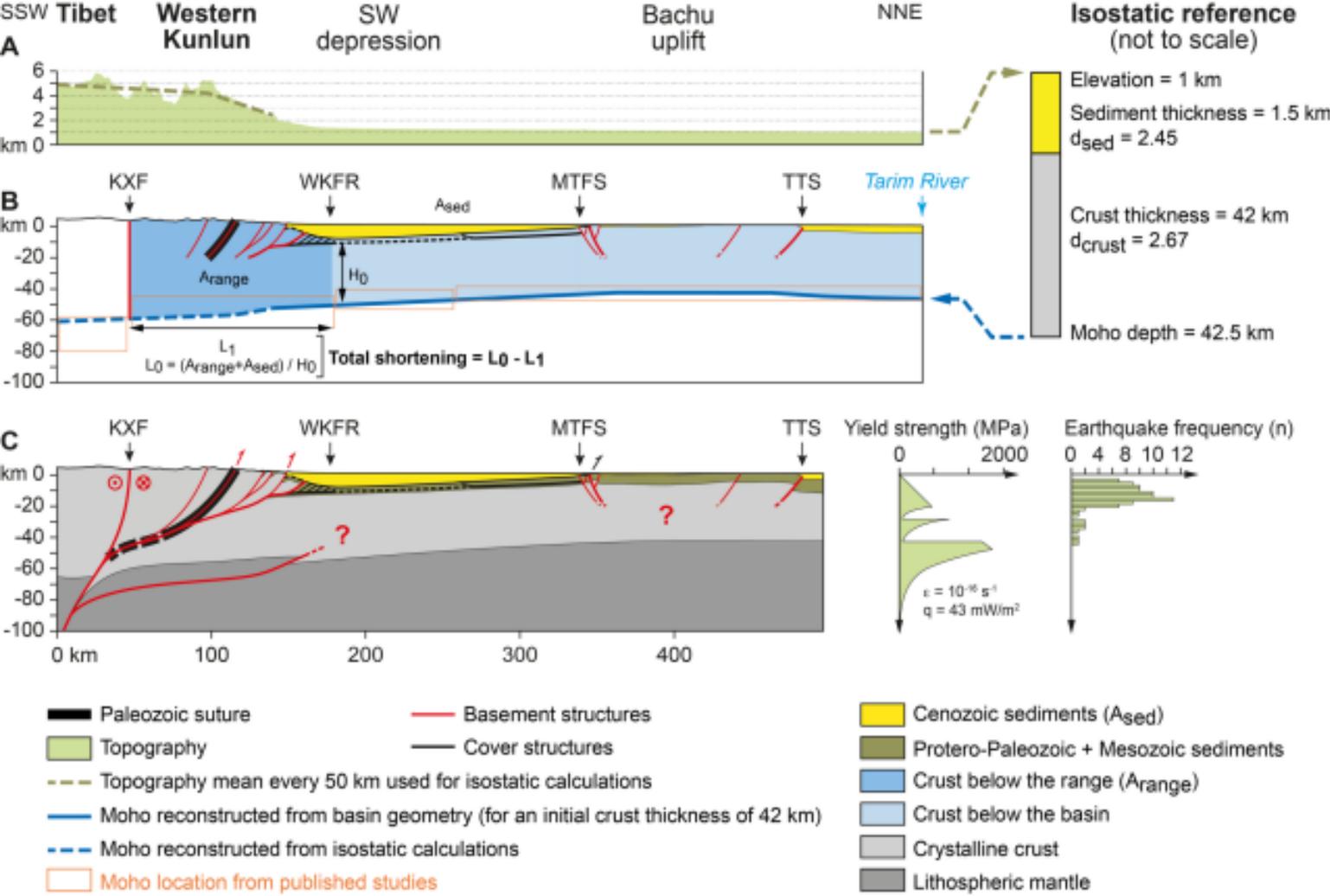
Southern Tian Shan

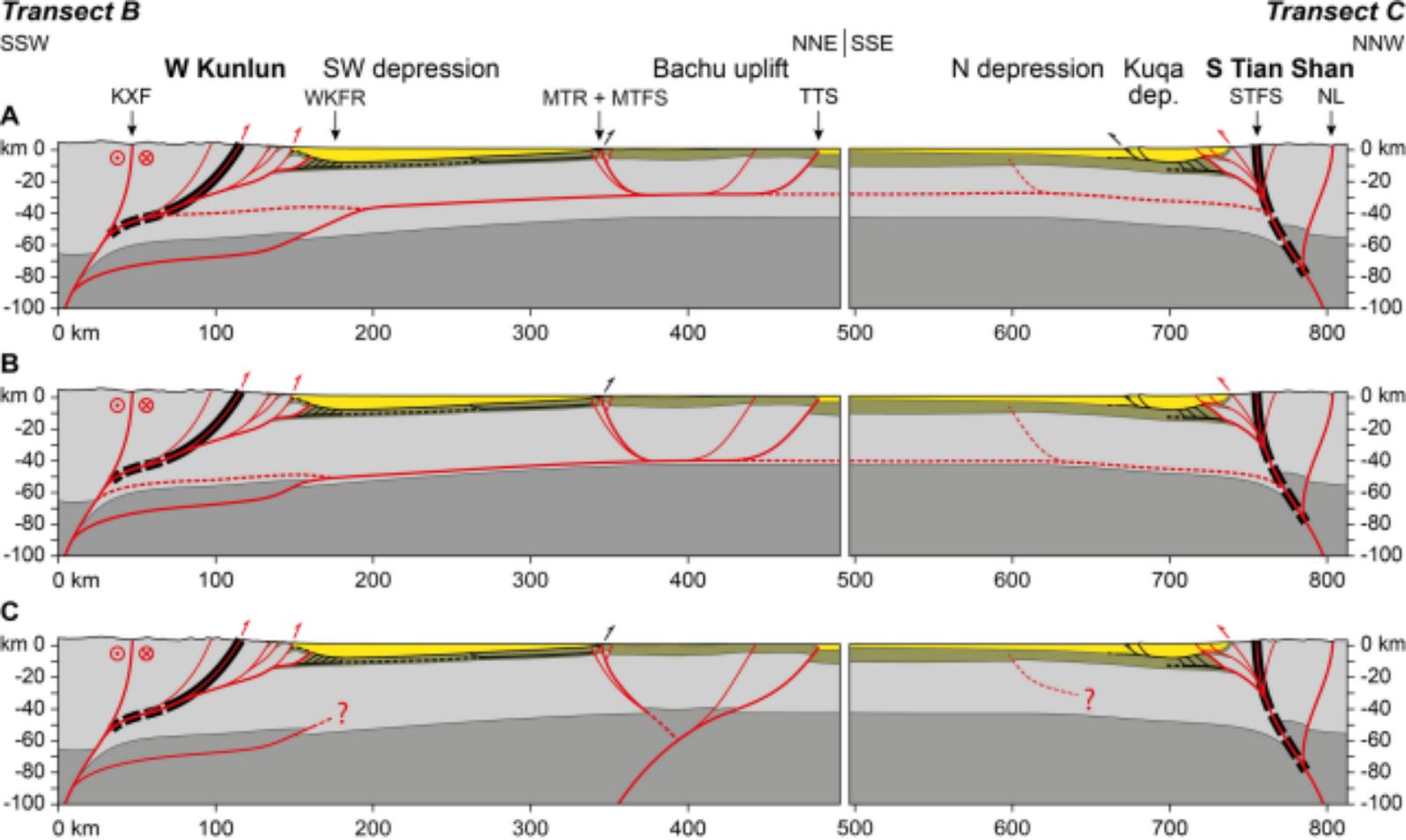
Proterozoic and Paleozoic basement and sediments
 Paleozoic sutures

Tarim Basin

 Sinian to Middle Cambrian	 Permian - Mesozoic	 Oligocene - Lower Miocene	 Pliocene - Quaternary
 Proterozoic basement	 Upper Cambrian to Carboniferous	 Paleocene - Eocene	 Middle - Upper Miocene







Range			Shortening (km)				
			1	2	3	4	5
			Entire range (crustal budg.)	Inner range (= 1 - 3)	Outer range (restor.)	Basin (restor.)	Total (= 1 + 4)
N Tibet	W Kunlun	Transect A (A1)	49 +23/-17.5	~17.5 +23/-17.5	~31.5 (inh. 9.5 + neo. 22)	~0.5	~49.5 +23/-17.5
		Transect B (B1)	78 ±23	~48 ±23	~30 (inh. 16.5 + neo. 13.5)	~5	~83 ±23
	Altyn Tagh	Transect C (C1)	38 ±18	~37.7 ±18	≥ ~0.3 (inh. 0.3+ neo. 0)	≥ ~0.6	~38.6 ±18
		Transect D (D1)	24 ±9	~23.7 ±9	≥ ~0.3 (inh. 0.3+ neo. 0)	0	24 ±9
S Tian Shan	SW Tian Shan	Transect A (A2)	54 +24/-18	~18 +24/-18	≥ ~36 (inh. 0+ neo. 36)	0	54 +24/-18
	SW Tian Shan	Transect B (B2)	53 ±23	/	≥ ~4 (inh. 4+ neo. 0)	0	53 ±23
	SW Tian Shan	Transect C (C2)	44 ±15	~22 ±15	~22 (inh. 19 + neo. 3)	0	44 ±15
	SE Tian Shan	Transect D (D2)	15 +20/-15	/	> ~0	0	15 +20/-15