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Rift inheritance controls the switch from thin- to thick-skinned thrusting and basal décollement re-localization at the subduction-to-collision transition

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1 Rift inheritance controls the switch from thin- to thick-skinned
2 thrusting and basal décollement re-localization at the subduction-
3 to-collision transition.

4

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17

18 **Keywords:** Thrust tectonics; Structural inheritance, Rifting, Continental collision

19

20 **ABSTRACT**

21 In accretionary convergent margins, the subduction interface is formed by a lower plate
22 décollement above which sediments are scraped off and incorporated into the accretionary wedge.
23 During subduction, the basal décollement is typically located within or at the base of the
24 sedimentary pile. However, the transition to collision implies the accretion of the lower plate
25 continental crust and deformation of its inherited rifted margin architecture. During this stage, the
26 basal décollement may remain confined to shallow structural levels as during subduction, or re-
27 localize into the lower plate middle-lower crust. Modes and timing of such re-localization are still

28 poorly understood. We present cases from the Zagros, Apennines, Oman, and Taiwan belts, all of
29 them involving a former rifted margin and pointing to a marked influence of inherited rift-related
30 structures on the décollement re-localization. A deep décollement level occurs in the outer sectors
31 of all of these belts, i.e. in the zone involving the proximal domain of pre-orogenic rift systems.
32 Older – and shallower – décollement levels are preserved in the upper and inner zones of the
33 tectonic pile, which include the base of the sedimentary cover of the distal portions of the former
34 rifted margins. We propose that thinning of the ductile middle crust in the necking domains during
35 rifting, and its complete removal in the hyperextended domains, hampered the development of
36 deep-seated décollements during the inception of shortening. Progressive orogenic involvement of
37 the proximal rift domains, where the ductile middle crust was preserved upon rifting, favors its
38 reactivation as a décollement in the frontal portion of the thrust system. Such décollement
39 eventually links to the main subduction interface, favoring underplating and the upward motion of
40 internal metamorphic units, leading to their final emplacement onto the previously developed
41 tectonic stack.

42

43 **1 INTRODUCTION**

44 Structural reactivation in rifted continental margins during plate convergence is a process
45 of paramount importance in the Wilson cycle (e.g., Wilson, 1966; Jackson, 1980; Cohen, 1982).
46 During convergence, reactivation of structures belonging to the continental margin of the
47 subducting lower plate marks the shift from subduction to collision. This shift typically involves an
48 initial “soft collision” stage, during which the stretched lithosphere of the distal portion of the
49 margin is subducted, and a later “hard collision” stage, which is established upon the inception of
50 poorly-stretched continental lithosphere subduction (e.g., Ballato et al., 2011). The change from soft
51 to hard collision leads to a major tectonic reorganization and commonly results in the reallocation
52 of deformation away from the subduction interface. In accretionary convergent margins, the
53 subduction interface consists of a basal décollement located within or at the base of the lower plate
54 sedimentary pile, the hanging wall of which is scraped off and incorporated into the accretionary

55 wedge (e.g., Nankai: Moore et al., 1990; or Cascadia: MacKay, 1995). With further convergence
56 and inception of the collisional phase, the basal décollement position becomes variable: it can be
57 located either within the lower plate sedimentary cover or within its middle-lower crust (e.g.,
58 Oncken et al., 1999; Lacombe and Mouthereau, 2002; Nemčok et al., 2013; Lacombe and
59 Bellahsen, 2016; Pfiffner, 2017). Evidence of the latter case is portrayed by major earthquakes
60 occurring in the crystalline basement in the frontal portions of mountain belts (e.g., the Taiwan
61 2013 Mw 6.2 Nantou earthquake, Brown et al., 2017; the Zagros 2017 Mw 7.3 Iran-Iraq
62 earthquake, Tavani et al., 2018a), which requires a basal décollement re-localization. Such process
63 apparently contradicts the simple notion that hinterland basement thrusts climb up-section foreland-
64 ward (Fig. 1A), but points out that the inherited crustal rheology across the previous rifted margin
65 may represent a fundamental control in the development of the décollement and in its propagation
66 during plate convergence (e.g., Lacombe and Bellahsen, 2016; Lescoutre and Manatschal, 2020)
67 (Fig. 1B). Consistently, in several mountain belts involving former rifted margins, the distal
68 sedimentary cover is usually detached from its basement to be thrust on top of the proximal and
69 necking domains of the rifted margin. There, thin-skinned thrusting eventually evolves to thick-
70 skinned by the reactivation of deeply rooted faults that deform the overlying thin-skinned system.
71 The Oman Mts. (Tarapoanca et al., 2010), the Apennines (Mazzoli et al., 2014), the European Alps
72 (Schmid et al., 2004), the Taiwan belt (Mouthereau and Lacombe, 2006; Brown et al., 2012;
73 Lacombe and Bellahsen, 2016), and the Zagros belt in the Lurestan region (Vergés et al., 2011) are
74 some well-known examples of this structural evolution. The architecture of these orogenic systems
75 suggests that the re-localization of the basal décollement at deeper structural levels occurs when the
76 necking and proximal domains of the rift system become involved in the collision.

77 Here we review information from several collisional systems, with an emphasis on the role
78 of inherited rift-related structures, to establish with unprecedented resolution a direct correlation
79 between the structure of collisional orogens and the detailed architecture of precursor rifted
80 margins. Finally, the transition from the first early “soft collision” stage to the later “hard collision”

81 stage is discussed as a possible factor triggering the reorganization of the orogenic wedge, leading
82 to the uplift and exhumation of metamorphic units in the interior of the belt.

83

84 **2 STRUCTURE OF THE WORK**

85 This work is organized into two main sections: case studies (section 3) and discussion (section 4).

86 The first includes four sub-sections describing each thrust belt and their corresponding

87 reconstructed rifted margins. For each illustrated thrust belt, we present a crustal-scale section and

88 the inferred architecture of the inherited rifted margin. Each case study section is further divided

89 into two parts: the first is a “geological setting”, in which the extant updated literature is reviewed,

90 followed by a “reconstruction of the rifted margin” section, in which the inherited rifted margin

91 architecture is interpreted and discussed in terms of structural domains. In the main discussion, we

92 link the geometry and kinematics of the thrust systems to the structure of the rifted margins,

93 stressing similarities and putting the different belts into an evolutionary order. We first discuss the

94 thrust belts that correspond to the initial stage of décollement re-localization, then we focus on the

95 structure of the more evolved orogens, and eventually infer the driving mechanism for the transition

96 from thin- to thick-skinned thrusting.

97

98 **3 CASE STUDIES**

99 **3.1 The Lurestan region of the Zagros**

100 ***3.1.1 Geological setting***

101 The Zagros belt developed due to the Cretaceous closure of the Neo-Tethys ocean and the

102 subsequent Cenozoic continental collision between the Arabian and Eurasian plates (e.g., Berberian

103 and King, 1981; Vergés et al., 2011; Mouthereau et al., 2012). The Main Zagros Thrust (MZT) and

104 the Main Recent Fault (MRF) form the still actively deforming suture of the orogenic system (e.g.,

105 Talebian and Jackson, 2004), dividing the former Arabian rifted margin to the SW from the

106 Sanandaj-Sirjan block to the NE (Fig. 2) (e.g., Berberian and King, 1981; Ghasemi and Talbot,

107 2006). To the SW of the suture, the High Zagros Fault (HZF) is a major regional thrust, which

108 hanging wall is mostly composed of the intensely folded and shortened deep-water sedimentary
109 cover of the distal portions of the Arabian rifted margin. The Simply Folded Belt (SFB) in the
110 footwall of the HZF extends to the Mountain Front Flexure (MFF) (Fig. 2A), a major topographic
111 feature controlled by the underlying basement-involved Mountain Front Fault (e.g., Berberian and
112 King, 1981).

113 The Lurestan Arc is the northwestern salient of the Zagros Mts. (Fig. 2A). It experienced
114 two shortening stages, as recorded by two syn-orogenic foredeep basins of Maastrichtian-to-Eocene
115 and Miocene-to-recent age (Homke et al., 2009; Saura et al., 2015). The older stage involved the
116 northeastern portion of the belt in the hanging wall of the HZF, whereas the younger one mostly
117 involved the SFB (Fig. 2). In the NE area, the tectonic pile exposed in the footwall of the suture
118 includes, from top to bottom (Fig. 2): (i) a thin ophiolite nappe, (ii) the Bisotun-Avalon nappe, and
119 (iii) the Kermanshah-Qulqula nappe (e.g., Vergés et al., 2011). The Bisotun-Avalon nappe is made
120 of sediments of the Upper Triassic to Lower Cretaceous Bisotun-Avalon carbonate platform
121 (Wrobel-Daveau et al., 2010). This carbonate platform formed on top of a crustal block separated
122 from the Arabian margin during the Triassic (Fig. 2D), as marked by the differentiation of the
123 stratigraphic successions of the two domains since that time (Fig. 2B) (Tavani et al., 2018b). The
124 Kermanshah-Qulqula nappe is made up of Lower Jurassic to Upper Cretaceous radiolarites, shales,
125 marls and limestones deposited in the Kermanshah-Qulqula basin (Gharib and De Wever, 2010);
126 this basin was originally interposed between the Arabian margin and the Bisotun-Avalon carbonate
127 platform (Fig. 2D), and partly developed on an exhumed mantle domain (Wrobel-Daveau et al.,
128 2010). This tectonic pile is covered by Paleogene to Miocene continental deposits belonging to the
129 Red Beds Fm. (Koshnaw et al., 2017), which bracket the main activity of the HZF in the
130 Maastrichtian-Paleocene interval. Apart from thin slices of serpentinized peridotites scraped off
131 from the substratum of the radiolarite basin, both the Bisotun-Avalon and Kermanshah-Qulqula
132 nappes are detached from their respective basement and have been telescoped for tens of kilometers
133 along sub-horizontal décollements located within the Triassic (Bisotun-Avalon nappe) and the
134 Lower Jurassic (Kermanshah-Qulqula nappe) successions. Further south, footwall splays of the

135 HZF occur in the northeastern portion of the SFB (Fig. 3A). These splays are rooted in a Triassic
136 décollement level, have displacements in the order of some kilometers, and are roughly coeval with
137 the HZF, since their associated folds are truncated by the HZF (Tavani et al., 2018a).

138 The SFB is characterized by low-displacement thrusts, by a mixed thin- and thick-skinned
139 style of deformation, and by a remarkably different Mesozoic sedimentary succession with respect
140 to that exposed in the hanging wall of the HZF. In detail, the Arabian margin sedimentary
141 succession in the Lurestan Arc includes 6-8 km of Mesozoic-Cenozoic clastic and carbonate rocks
142 (Fig. 2B) overlying *ca.* 4 km of Paleozoic units (Hessami et al., 2001; Vergés et al., 2011).

143 Apart from the splays of the HZF (Fig. 3A), the proximal domain of the Arabian margin was
144 deformed in a piggy-back sequence since the Miocene (Barber et al., 2018). The basement-cover
145 interface of the SFB is recognized as a regional mechanical weakness, which promoted partial
146 decoupling between the sedimentary cover and the crystalline basement (Vergés et al., 2011; Tavani
147 et al., 2018a). In detail, low-displacement reverse basement faults ramp-up to join the basement-
148 cover interface, above which the sedimentary cover is mostly shortened by folding (Fig. 2C).
149 Steeply-dipping basement thrusts are rooted into a mid-crustal décollement at about 20-25 km depth
150 (e.g., Vergés et al., 2011), and many of them are interpreted as inverted extensional faults (Tavani et
151 al., 2018a). The occurrence of the mid-crustal ductile décollement level is also evidenced by a
152 seismic gap occurring between 20 and 30 km depth (e.g., Talebian and Jackson, 2004; Tavani et al.,
153 2020). The Mountain Front Fault is a moderate- to gently-dipping active thrust at the tip of the mid-
154 crustal décollement, with a displacement in the order of 10 km (Vergés et al., 2011; Tavani et al.,
155 2018a). The development of the Mountain Front Fault occurred in a very recent stage of the thrust
156 belt evolution, approximately <3 Ma (Tavani et al., 2020), thus representing the youngest feature of
157 the belt. With the exception of this fault, displacement along basement reverse faults is rather
158 limited.

159

160 ***3.1.2 Reconstruction of the rifted margin***

161 The Arabian margin involved in the Zagros Belt underwent two-stage rifting, during the
162 Permian-Triassic and the Early-Middle Jurassic intervals (Tavani et al., 2018b). However, the first
163 stage poorly affected the Lurestan region and the tectono-stratigraphic architecture of the margin
164 mostly results from the Early-Middle Jurassic extensional tectonics, coeval with the development of
165 the Kermanshah-Qulqula basin (Tavani et al., 2018b). During this extensional phase, large areas
166 evolved from shallow to deep-water depositional environments (Figs. 2B, 3A) (e.g., Barrier and
167 Vrielynck, 2008), with largely underfilled Middle and Late Jurassic basins where deep-water
168 conditions persisted until the Late Cretaceous onset of convergence. Middle to Upper Jurassic rocks
169 are poorly exposed in the SFB, hence a clearer marker of the drowned Jurassic tracts is provided by
170 the extensively exposed Lower Cretaceous facies, which are here used to infer the limits between
171 the different domains of the rifted margin.

172 In detail, three domains can be recognized (Fig. 2B). To the SW, shallow-water carbonate
173 sedimentation spans from the Permian until the Late Cretaceous (Fig. 2A, B). This large area did
174 not experience remarkable crustal thinning and corresponds to the unstretched Arabian Plate (Fig.
175 2D). The central part of the study area corresponds to the proximal domain of the Arabian Jurassic
176 rifted margin, which records crustal stretching and drowning. This is witnessed by the deep-water
177 Lower Cretaceous Garau Fm., which is made up of limestones, marls, and shales (Jassim and Goff,
178 2006). Radiolarian-rich beds in its lower portion are well known (Fig. 3B) and testify for a
179 paleobathymetry near the carbonate compensation depth (CCD) and basin starvation. Here we
180 firstly report that in the northernmost area, the Garau Fm. also includes m-thick radiolarites (Fig.
181 3C), thus representing even deeper facies (below the CCD), transitional between the typical
182 radiolarian-rich Garau Fm. and the radiolarites of the Kermanshah-Qulqula basin. In our
183 interpretation, this transitional facies is indicative of an area that was part of the necking domain of
184 the Jurassic rift, transitional between the stretched proximal domain and the hyperextended
185 Kermanshah-Qulqula radiolarite basin, in which local mantle exhumation is reported (Wrobel-
186 Daveau et al., 2010). The width of the Kermanshah-Qulqula radiolarite basin and of the Bisotun-
187 Avalon extensional ribbon are from Verges et al. (2011).

189 **3.2 The southern Apennines**

190 **3.2.1 Geological setting**

191 The Late Cretaceous to Quaternary Apennines belt developed within the framework of
192 Africa-Eurasia plate convergence (e.g., Dewey et al., 1989; Doglioni, 1991; Faccenna et al., 2014).
193 In this subduction-to-collision system, Eurasia represented the overriding plate, whereas the
194 downgoing plate included the Jurassic to Early Cretaceous Alpine Tethys ocean and its southern
195 rifted margin, i.e. the Adria promontory of Africa. Remnants of the overriding plate are extensively
196 outcropping in southernmost peninsular Italy and in northern Sicily, in the so-called Calabria-
197 Peloritani Arc (Somma et al., 2005, and references therein). These continental crust units overthrust
198 ophiolitic rocks (Liguride complex; Ogniben, 1969), exposed along the tectonic suture. In the
199 southern Apennines, convergence was characterized by a roughly E- to NE-ward migration of the
200 thrust front and the related foreland basin (e.g., Menardi Noguera and Rea, 2000; Patacca and
201 Scandone, 2007) and by the progressive tectonic incorporation of the Mesozoic-Tertiary domains
202 established on the Adria promontory of Africa during the Late Triassic to Early Jurassic rifting
203 stage. In summary, Triassic to Jurassic divergence between Gondwana and Eurasia led to the
204 opening of the Alpine Tethys, of which Adria formed part of the SE rifted margin. There, a suite of
205 horst and graben/half-graben structures formed upon rifting, in which carbonate platforms and
206 pelagic basins developed. In the southern Apennines, these Mesozoic domains are, from west to
207 east (Fig. 4D): the Apennine carbonate platform, the Lagonegro pelagic basin, and the Apulian
208 carbonate platform (Ogniben, 1969; D'Argenio et al., 1975; Patacca and Scandone, 2007; Cosentino
209 et al., 2010; Santantonio and Carminati, 2011).

210 During the Eurasia-Adria Cenozoic convergence, deep-water sediments and slices of
211 ophiolites were scraped off from the downgoing Tethys oceanic plate and accreted to form the
212 Liguride complex (Ogniben, 1969; Knott, 1987). During the middle Miocene, this accretionary
213 complex was thrust eastward for tens of kilometers and placed on top of the Apennine Platform and
214 its Miocene syn-orogenic cover (e.g., Roure et al., 1991; Menardi Noguera and Rea, 2000; Scrocca,

215 2010). To date, only a few exposures of slope and/or basin facies located to the west of the
216 Apennine Platform are reported in the area (Capri Island, ISPRA 2014; Curzi et al., 2020),
217 indicating that the entire necking and hyperextended domains between the Apennine Platform and
218 the Liguride basin has been overthrust by the Liguride complex. During the middle to late Miocene,
219 the tectonic pile was detached along a basal décollement placed at the base of the sedimentary cover
220 of the Apennine platform and was thrust NE-ward on top of the Lagonegro basin units. Finally,
221 during the Pliocene, the sedimentary cover of the Lagonegro basin was also detached from its pre-
222 Mesozoic basement and placed on top of the Apulian Platform (Roure et al., 1991; Menardi
223 Noguera and Rea, 2000). The latter is presently buried below the thin-skinned thrust belt and rises
224 to the NE, being exposed in the foreland region. The décollement between the allochthonous units
225 and the buried Apulian Platform unit is marked by a fluid-saturated, clay-rich mélange zone of
226 variable thickness, reaching up to ca. 1500 m (Mazzoli et al., 2001; Fig. 4). The above described
227 thin-skinned tectonic pile has been re-deformed by subsequent late Pliocene to early Pleistocene
228 shortening, controlled by steeply-dipping reverse faults rooted in the Apulian Platform basement
229 (Mazzoli et al., 2014) (Fig. 4C). Immediately to the SW of these faults, seismic data indicate the
230 occurrence of a slope domain between the Apulian Platform and the Lagonegro basin (Menardi
231 Noguera and Rea, 2000). This slope domain is currently in the footwall to the Lagonegro basin
232 basal thrust, which in turn is to the footwall of the Apennine Platform basal thrust. Importantly, the
233 Moho across this area rises up from 32-34 km to the NE to 26-28 km to the SW (Menardi Noguera
234 and Rea, 2000; Di Bucci et al., 2006; Mazzoli et al., 2014). It is noteworthy that thick-skinned
235 thrusting has not developed in the Apennine Platform, which was originally interposed between two
236 areas of hyper-extended crust. Deeply rooted thrusts linked to a deep crustal décollement have
237 developed only when a significant portion of the proximal domain of the Apulian platform has been
238 involved in the orogenic pile.

239

240 ***3.2.2 Reconstruction of the rifted margin***

241 Triassic to Early Jurassic extension in the Adria crustal block led to the formation of a
242 rifted margin made of a suite of carbonate platforms and pelagic basins. Our reconstruction of the
243 margin is shown in Figure 4D. The Lagonegro basin is here interpreted as a hyperextended domain
244 due to the occurrence of deep-water radiolarites deposited below the CCD in its central parts (in
245 which the whole Jurassic column consists of 70-80 m of radiolarian cherts and siliceous argillites;
246 Mazzoli et al., 2001, and references therein), with a commonly inferred width of nearly 150 km
247 (Menardi Noguera and Rea, 2000; Scrocca et al., 2005). On the other hand, the pre-orogenic width
248 of the Apennine carbonate platform is estimated to be nearly 60 km (Menardi Noguera and Rea,
249 2000; Mazzoli et al., 2014); this value is slightly underestimated, since the transition between the
250 Apennine carbonate platform and the basin located to the SW is generally not included in cross
251 sections. In this sense, as exposures of slope and/or basinal facies occur a few tens of km to the SW
252 of the coastline (e.g., Capri Island, ISPRA 2014), the maximum original width for the Apennine
253 carbonate platform (including the surrounding slope domains) should not exceed 100 km.

254 An important feature for the rifted margin reconstruction is the step in the Moho observed
255 along the SW portion of the regional cross section (Fig. 4C). The Moho step has been previously
256 interpreted as associated with a SW-dipping reverse fault affecting the upper mantle and the crust
257 (Menardi Noguera and Rea, 2000). However, we propose a different interpretation coherent with
258 the main message of this work and contractional slip transfer considerations. The step is here
259 interpreted as a NE-dipping lower crust extensional shear zone related to Mesozoic crustal thinning.
260 Our interpretation is strongly supported by the two following observations: (i) the Moho step is
261 placed exactly underneath the slope domain interposed between the Apulian shallow-water
262 carbonate platform and the Lagonegro deep-water basin, i.e. exactly where crustal extension is to be
263 expected; and (ii) in agreement with the transition between the two different depositional
264 environments, the pre-orogenic crustal thickness reduces from 30 km to the NE (Apulian Platform
265 cover plus the middle and lower crust) down to less than 20 km to the SW (Lagonegro basin cover
266 plus the middle and lower crust); such a crustal thinning requires significant extension in the area.
267 Within this framework, the Moho step would be placed within or immediately to the NE of the

268 boundary between the former necking and proximal domains (Fig. 4D). No information instead is
269 available to define the position and width of the necking domains that bounded the Apennine
270 Platform to the NE and SW. Similarly, the width of the hyperextended domain between the Neo
271 Tethys and the Apennine Platform shown in figure 4C is highly speculative.

272

273 **3.3 Oman**

274 **3.3.1 Geological setting**

275 The Oman mountain belt (also known as Al-Hajar mountains) developed following the Late
276 Cretaceous closure of the Neo-Tethys ocean between the Arabian and Eurasian plates, and by
277 subsequent shortening during the Cenozoic (Glennie et al., 1973; Tarapoanca et al., 2010; Hansman
278 et al., 2017). Ocean closure was initiated by intra-oceanic NE-directed subduction (Robertson and
279 Searle, 1990; Jolivet et al., 2016), followed by the generation of the Semail ophiolites in a supra-
280 subduction environment at 96-95 Ma (e.g., Robertson and Searle, 1990; Rioux et al., 2012). The
281 obducted Semail ophiolites occupy the uppermost portion of the Oman tectonic pile (Fig. 5). The
282 thin metamorphic sole at the base of the Semail ophiolite peridotite records amphibolite facies
283 metamorphism at 95-92 Ma (i.e. 54-46 km depth, Searle and Cox, 2002). Obduction initiated at 95-
284 93 Ma (Hacker, 1994), when the Semail ophiolites were thrust over the various domains of the
285 distal portion of the former Arabian rifted margin. The highly deformed and partly dismembered
286 remnants of the Arabian margin are presently sandwiched between the Semail ophiolites and the
287 Arabian shelf, which includes the stable continent and the proximal domain of the rift. The
288 northeastern portion of the Arabian shelf was involved in the subduction, as recorded by blueschist
289 to eclogite facies metamorphic rocks exposed at the Saih Hatat tectonic window (SH in Fig. 5A)
290 (Chemenda et al., 1996; Breton et al., 2004). During convergence, the sedimentary cover of the
291 distal domains of the Arabian margin was progressively detached from its basement, incorporated
292 into the tectonic wedge, and thrust on top of the proximal domain of the Arabian rifted margin.
293 During the foreland-ward migration of the allochthonous wedge, the Arabian foreland underwent
294 SW-migrating flexural bending with uplift and erosion of the rifted margin, as witnessed by the

295 development of a Turonian unconformity (Boote et al., 1990; Cooper et al., 2014) dating the onset
296 of deformation related to convergence in the Arabian shelf. The wedge emplacement onto this Late
297 Cretaceous foredeep ended by the Campanian (Warburton et al., 1990), almost coevally with the
298 eclogite-facies metamorphism recorded in the Saih Hatat rocks at around 79 Ma (Warren et al.,
299 2003). Stable shelf conditions alternated with emersion periods during the early Maastrichtian to
300 Eocene time interval, leading to the sedimentation of an unconformable post-kinematic sequence
301 (Post-Nappe in figure 5) covering the Cretaceous wedge to the NE and progressively becoming
302 conformable over the foredeep sediments to the SW.

303 During the Cenozoic, a second shortening phase occurred, responsible for the deformation
304 of the Upper Cretaceous to Cenozoic Post-Nappe package (Corradetti et al., 2020) and for the final
305 uplift and doming of the Jabal Akhdar and Saih Hatat structural culminations (Mount et al., 1998;
306 Breton et al., 2004; Cooper et al., 2014; Hansman et al., 2017). Late deformation is well constrained
307 by thermochronological data (Mount et al., 1998; Saddiqi et al., 2006; Tarapoanca et al., 2010;
308 Hansman et al., 2017; Corradetti et al., 2020) and absolute radiometric dating of veins from tectonic
309 windows and surrounding tectonic units (Grobe et al., 2018; Hansman et al., 2018). Despite the
310 large debate on the timing and modes of exhumation of the Oman mountains, available data point
311 toward a rapid exhumation starting from the early Eocene, at 45-40 Ma, (Tarapoanca et al., 2010;
312 Jacobs et al., 2015; Hansman et al., 2017), and a later reactivation at 20-15 Ma (Jacobs et al., 2015;
313 Corradetti et al., 2020). Eocene uplift and unroofing relate to a major shortening event that affected
314 the whole nappe edifice, triggering flexuring of the lower plate (e.g., Homewood et al., 1986) and
315 development of a new foredeep (Tarapoanca et al., 2010; Corradetti et al., 2020). Such shortening
316 event is associated with thick-skinned thrusting, most-likely resulting from positive inversion of
317 pre-existing basement faults inherited from the Permian-Triassic extensional phase (Boote et al.,
318 1990; Tarapoanca et al., 2010; Hansman et al., 2017). Notably, in the area crossed by our section no
319 evidence of thick-skinned tectonics during the Cretaceous stage has been documented so far for the
320 Arabian shelf.

321

322 3.3.2 *Reconstruction of the rifted margin*

323 The Arabian rifted margin in the Oman region formed during the Permo-Triassic magma-
324 rich rifting (e.g., Robertson and Searle, 1990). The different paleogeographic domains of the
325 margins are the Arabian Shelf to the SW, and the Hawasina basin to the NE, the latter including the
326 Sumeini slope, and the Hamrat Duru and Umar deep-water basins separated by remnants of
327 platform carbonates such as the Kawr Group (Fig. 5D) (Béchenec et al., 1988; Rabu et al., 1993).
328 Accordingly, we here interpret these paleogeographic domains in terms of structures of a rifted
329 margin. In detail, the reconstruction of the Arabian margin includes, from SW to NE: (i) the
330 proximal domain (NE portion of the Arabian shelf) characterized by a Neo-Proterozoic to
331 Cretaceous sedimentary pile exposed in the Jabal Akhdar and the Saih Hatat culminations and in the
332 Musandam peninsula (JA, SH, and MU in figure 5A), where shallow water conditions persisted
333 during the Mesozoic; (ii) the Sumeini slope, where deep water facies (including > 10 m thick
334 packages of radiolarites) are intercalated with distal calcarenites during the Permian-Cretaceous
335 time span. We infer that the occurrence of these thick packages of radiolarites is indicative of deep-
336 water conditions (close to the CCD depth) and basin starvation, which we correlate with the necking
337 domain. Thus, the NE portion of the Sumeini slope in our reconstruction coincides with the necking
338 domain; (iii) the distal portion of the margin, mostly made by the deep-water sediments resting on
339 top of Permian-Triassic syn-rift volcanic rocks. In detail, the latter domain is divided in two basins:
340 the Hamrat Duru and Umar basins, separated by a narrow domain (Kawr), in which shallow-water
341 carbonate platform conditions persisted at least during the Triassic rifting, constituting an elevated
342 feature (horst) throughout the Jurassic-Cretaceous times (Béchenec et al., 1988; Rabu et al., 1993).
343 The occurrence of the Kawr carbonate platform within the distal portion of the margin mimics the
344 structure of the Bisotun-Avalon carbonate platform of the Zagros and of the Apennine Platform of
345 the southern Apennines. In agreement, we suggest that the Kawr carbonate platform represents a
346 continental ribbon (sensu Péron-Pinvidic and Manatschal 2010) (Fig. 5D). The preserved geological
347 record of this ribbon is not sufficiently detailed to allow estimating the position of the necking
348 domains surrounding it and, in agreement, these necking domains are not indicated. More generally,

349 the width of the different domains reported in figure 5D remain speculative (albeit in full agreement
350 with their typical width documented worldwide; Chenin et al., 2017), as the far-travelled and highly
351 re-imbricated nature of the thrust sheets of the Oman belt prevents accurate reconstructions. Indeed,
352 only the cumulative width of the entire Hawasina basin is generally reconstructed, being at least 400
353 km-wide (Béchenec et al., 1988; Cooper, 1988).

354

355 **3.4 The Taiwan mountain belt**

356 **3.4.1 Geological setting**

357 The Taiwan mountain belt is forming since the late Miocene (6.5 Ma, Lin et al., 2003) as a result of
358 the oblique collision between the nearly NE-SW oriented Eurasian rifted margin and the
359 approximately N-S trending Luzon volcanic arc on the Philippine Sea Plate (e.g., Teng, 1990;
360 Sibuet and Hsu, 2004; Huang et al., 2006; Byrne et al., 2011; Brown et al., 2012) (Fig. 6). In more
361 detail, the Eurasian rifted margin formed during the opening of the South China Sea during the
362 Oligocene–Miocene, whereas collision took place after an initial phase of E-directed subduction of
363 the South China Sea oceanic lithosphere below the Philippine Sea plate, which started about 15 Ma
364 ago and is still active south of Taiwan (e.g., Huang et al., 2006). From its more internal (i.e.,
365 eastern) to its more external (i.e., western) parts, the Taiwan mountain belt comprises five fault-
366 bounded domains: the Coastal Range, the Central Range, the Hsuehshan Range, the Western
367 Foothills, and the Coastal Plain (Fig. 6A,C). The Coastal Range is made up of volcanic rocks of the
368 Luzon arc and sediments deposited within intra-arc basins, and it is juxtaposed against the Central
369 Range along the Longitudinal Valley Fault (LVF, Fig. 6), which is thought to represent the plate
370 boundary within the collision zone (e.g., Shyu et al., 2008). The Central Range comprises from west
371 to east (Fig. 6C): (i) the Lushan slate belt, made of syn- to post-rift Eocene to Miocene sediments
372 (Clark et al., 1993; Beyssac et al., 2007; Simoes et al., 2007; Brown et al., 2012), which underwent
373 low-grade syn-orogenic metamorphism (Beyssac et al., 2007); (ii) the Tananao Complex, a
374 greenschist facies metamorphic complex (e.g., Chen et al., 2017) made of Mesozoic pre-rift
375 basement rocks of the Eurasian rifted margin (i.e. marbles and schists), and (iii) the Yuli belt, the

376 remnants of a Miocene accretionary prism with exotic blueschist blocks (e.g., Chen et al., 2017) that
377 reached high-pressure metamorphic conditions during subduction (Keyser et al., 2016). The
378 Tananao Complex is thrust over the Lushan belt along the Chinma thrust. Exhumation of the
379 Tananao Complex along the Chinma thrust took place during the early to late Pleistocene (Dorsey,
380 1987; Lee et al., 2006; Brown et al., 2012) and some authors have linked its development to
381 underplating and nappe stacking processes (e.g., Simoes et al., 2007; Molli and Malavieille, 2011)
382 (Fig. 6C). The Central Range is juxtaposed against the Hsuehshan Range across the Lishan Fault in
383 the north (Clark et al., 1993; Brown et al., 2012) and against the Western Foothills across the
384 Chaochou Fault in the south (Mouthereau et al., 2002; Tang et al., 2011, Fig. 6A). The Hsuehshan
385 Range comprises variably metamorphosed Eocene and Oligocene clastics (Beysac et al., 2007;
386 Simoes et al., 2012) that were deposited within the Hsuehshan basin on the Eurasian rifted margin
387 (Teng and Lin, 2004; Huang et al., 2006; Brown et al., 2012) (Fig. 6C,D). High-temperature
388 metamorphism of this portion of the belt, that represents an outlier in the eastward-increasing
389 metamorphic trend, has been interpreted as associated with the rifting stage, rather than with
390 mountain building (Beysac et al., 2007). The Western Foothills consist of imbricated Eocene to
391 Miocene syn- to post-rift sediments and younger syn-orogenic sediments (Lacombe et al., 1999;
392 Yue et al., 2005; Brown et al., 2012) (Fig. 6C). Finally, the Coastal Plain is made up of weakly
393 deformed Pliocene to Holocene syn-orogenic sediments of the foreland basin.

394 The Taiwan mountain belt displays a hybrid, thick- and thin-skinned, structural style. The
395 basal décollement of the Western Foothills is located nearly at the base of the syn-orogenic
396 sediments. Conversely, in the more internal domains, rocks of the pre-rift basement are involved in
397 the deformation and crop out in the Tananao Complex (Fig. 6C). Surface geology (Clark et al.,
398 1993; Mouthereau et al., 2002; Brown et al., 2012; 2017) and the shallowing of high P-wave
399 velocities taking place along deeply rooted clusters of earthquake hypocenters in the interior of the
400 mountain belt (Gourley et al., 2007; Wu et al., 2007; Kuo-Chen et al., 2012; Brown et al., 2017),
401 indicate that basement is uplifted along steeply dipping faults that penetrate into the middle crust
402 and formed as a result of the inversion of pre-existing extensional faults (Mouthereau and Lacombe,

403 2006; Brown et al., 2012; 2017; Lacombe and Bellahsen 2016) (Fig. 6D). For example, the
404 development of the Hsuehshan Range is associated with the inversion of the Hsuehshan basin due
405 to the reactivation of its bounding faults (i.e., the Shuilikeng and the Lishan faults; Brown et al.,
406 2017; Fig. 6C,D).

407 Importantly, the Eurasian rifted margin underwent at least two major phases of deformation
408 during convergence that led to the development of the Taiwan mountain belt (Brown et al., 2012).
409 With the arrival of the leading edge (i.e., distal part) of the margin at the subduction zone, a first
410 phase of thin-skinned deformation was responsible for detaching (and westward translation) of the
411 syn- to post-rift sediments from their pre-rift basement (i.e., the current Lushan belt; Brown et al,
412 2012); the Tili thrust represented the leading thrust during this initial thin-skinned deformation
413 phase. Remnants of this first thrust system can be currently found in the Taiwan mountain belt as a
414 folded and faulted thrust fault cropping out in the Hsuehshan Range and in the Lushan slate belt,
415 and inferred also in the eroded portion of the Tananao complex (Fig. 6D). Once the relatively
416 poorly extended proximal domain of the rifted margin reached the subduction zone as convergence
417 proceeded, shortening was mostly accommodated by thick-skinned inversion of pre-existing
418 extensional faults and extensive basement uplift as described above (Mouthereau and Lacombe,
419 2006; Brown et al., 2012; 2017) (Fig. 6D).

420

421 ***3.4.2 Reconstruction of the rifted margin***

422 Prior to collision in Eocene times, the Eurasian margin had the structure of a
423 hyperextended margin (Brown et al., 2012; McIntosh et al., 2013) (Fig. 6D). In the cross-section of
424 Figure 6C, the thinned crust of the hyperextended domain is not present as resulting from the
425 subduction of the entire distal domain (e.g., McIntosh et al., 2005). However, the obliquity between
426 the rifted margin and the thrust belt, and the preservation of the former in the foreland of the
427 southern portion of the Taiwan belt, allows setting the approximate boundaries between the
428 different rift domains. In detail, in figure 6D we show a slightly modified version of the
429 reconstruction by Brown et al. (2012). In particular, albeit with some uncertainty, the necking

430 domain in the cross-section of Figure 6C can be reasonably placed some tens of km to the east of
431 the suture, coherently with the eastward thinning of the crust which occurs in the subducted plate. In
432 fact, despite the remarkable deformation, the crustal thickness underneath the Hsuehshan and
433 Central ranges (particularly the occurrence of a thick basement overlying the ductile crust),
434 indicates that these areas were forming part of the proximal domain.

435

436 **4 DISCUSSION**

437 This discussion is organized in three sub-sections. Firstly, we recall the structure of the four
438 illustrated rifted margins, stressing their architecture and the recognition of different structural
439 domains. Then, we discuss the process of décollement re-localization and, finally, we discuss our
440 findings in the context of the role of structural inheritance in mountain building processes.

441

442 **4.1 Recognition of inherited rifted margins**

443 *4.1.1 Lurestan region of the Zagros*

444 The architecture of the former Arabian rifted margin in the Lurestan is nowadays well
445 understood (e.g., Wrobel-Daveau et al., 2010; Vergés, et al., 2011; Saura et al., 2015; Fig. 2D), and
446 archetypal domains of hyperextended magma-poor rifted margins (e.g., Whitmarsh et a., 2001;
447 Lavier and Manatschal, 2006; Péron-Pinvidic and Manatschal 2009) have recently been recognized
448 in this area. To the NE, the Bisotun-Avalon carbonate platform marks the occurrence of a former
449 extensional ribbon of continental crust (*sensu* Lister et al. 1986). In addition, the presence of
450 radiolarites, indicating paleo-water depths below the CCD (De Wever et al., 1994), and the
451 occurrence of serpentinitized peridotites underlying them (Wrobel-Daveau et al., 2010) reveal that
452 the Kermanshah-Qulqula basin represents a hyperextended (i.e. a domain in which the continental
453 crust was thinned down to less than 10 km; Péron-Pinvidic and Manatschal, 2009) to exhumed
454 mantle domain. The necking domain of the Jurassic margin, i.e. the thinned and faulted part
455 separating the weakly extended crust of the proximal domain from the hyperextended domain, may
456 be placed in the NE portion of the Arabian margin (Fig. 2), coherently with the fact that the post-rift

457 Garau Fm. there includes meter-thick radiolarites (Figs. 2,3) pointing to the proximity of the CCD.
458 To the SW, the area labeled as “Arabian Plate” is the region devoid of any significant crustal
459 stretching.

460

461 *4.1.2 Southern Apennines*

462 The telescoped Adria rifted margin presently incorporated in the southern Apennines is also
463 well constrained (e.g., D’Argenio et al., 1975; Patacca and Scandone, 2007; Cosentino et al, 2010).
464 In this area, the southern margin of the Alpine Tethys has been defined as a Triassic to Early
465 Jurassic magma-poor rifted margin (e.g., Mohn et al., 2012). Deep water radiolarites occur in the
466 central portion of the Lagonegro basin (e.g., Mazzoli et al., 2001) and the observed thinning of the
467 crust down to ~20 km at the transition with the Apulian Platform (Fig. 4C) allows interpreting at
468 least the central part of the Lagonegro basin as a hyperextended domain. The Apennine Platform, in
469 between the Lagonegro basin to the east and the Alpine Tethys to the west, can be regarded as an
470 extensional ribbon, similar to the Bisotun-Avalon carbonate platform of the Lurestan area. The
471 former necking domain of the western rifted margin of Adria can be located along the eastern
472 margin of the Lagonegro basin, immediately to the SW of the Moho step (Fig. 4C); within this
473 framework the Apulian Platform would represent the proximal rift domain.

474

475 *4.1.3 Oman*

476 The former Arabian rifted margin in the Oman Mts. slightly differs from that of the
477 Lurestan, in terms of both timing and magma supply. The timing of rifting is Permian-Triassic, as
478 recorded by the Triassic deep-water sedimentary record (e.g., Rabu et al., 1993). Furthermore, the
479 abundant Permian volcanites included in various sedimentary successions of the Hawasina basin
480 (e.g., Robertson and Searle, 1990) point to significant magmatic input. Apart from these
481 differences, the shallow-water carbonates of the Kawr domain, between the deep-water Hamrat
482 Duru and Umar domains, were most likely also deposited on an extensional ribbon. Radiolarites
483 belonging to the Hamrat Duru and Umar domains suggest a paleodepth close to the CCD and a

484 remarkable crustal thinning for these domains. Consistently, the crustal-scale cross-section in
485 Figure 5C shows that the pre-orogenic crust at the Sumeini slope to Arabian shelf transition, i.e. in
486 the thicker portion of the Hawasina basin, should have been nearly 20 km thick. Accordingly, we
487 infer hyper-extension in the Hamrat Duru domain, which was placed in a more distal position with
488 respect to the 20 km thick crust of the Sumeini slope. The substratum of the Umar sub-basin is
489 instead interpreted as made of both hyperextended continental crust and oceanic crust.

490

491 *4.1.4 Taiwan*

492 The rifted margin deformed in the Taiwan belt is the least constrained one (Fig. 6). Rifting is
493 Eocene in age (Brown et al., 2012; McIntosh et al., 2013) and no record for extensional ribbon
494 occurs there. The margin consisted of three sectors matching the proximal, necking and
495 hyperextended domains; however, neither the basement nor the respective sedimentary covers of
496 the latter two domains are exposed. On the other hand, the geometry of these extensional domains
497 as well as the approximate location of their boundaries, can be proposed by taking advantage of the
498 obliquity between the rifted margin and the thrust belt (i.e. the former proximal domain of the
499 margin is preserved on the active foreland at the southern portion of the Taiwan belt; e.g., Brown et
500 al., 2012).

501

502 **4.2 Décollement re-localization at the transition from subduction to collision**

503 After rifted margin development and during the subsequent plate convergence, the distal
504 parts of the rifted margin arrives to the subduction zone once the attached oceanic crust has been
505 consumed during subduction. At this stage, the basal décollement propagates from the base of the
506 ophiolitic sequence into the base of (or within) the rifted margin sedimentary cover. The
507 involvement of continental extensional ribbons seems not to produce any remarkable effect on the
508 thin-skinned thrust system: examples illustrated above indicate that the basal décollement remains
509 confined to the sedimentary cover by shearing off the structural highs. When the hyperextended
510 domain of the lower plate (i.e. the Lagonegro basin in the Apennines; the Kermanshah-Qulqula

511 basin in Lurestan; and the Hamrat Duru sub-basin in Oman) arrives at the trench, shortening is still
512 accommodated by thin-skinned thrusting. This is shown by the Lagonegro basal thrust, the High
513 Zagros Fault, and the Hawasina basal thrust, which detached and telescoped the sedimentary
514 succession of the distal portion of the rifted margin on top of its proximal domain.

515 As collision evolves, slightly different evolutions can be defined based on the examples
516 above. In the Lurestan, the thin-skinned basal décollement propagated from the hyperextended
517 domain into the necking domain across weak stratigraphic levels. Further shortening and
518 convergence (re-)activated the inherited basement faults of the necking and proximal domains. The
519 first stage of this thick-skinned deformation is characterized by limited shortening accommodated
520 by inversion tectonics (Granado and Ruh, 2019; Tavani et al., 2020). In the latest stage of
521 deformation, the seismically active low-angle Lurestan Mountain Front Fault developed, with a
522 cumulative displacement of nearly 10 km. The activity of this fault testifies the presence of an
523 efficient and laterally continuous mid-crustal décollement along which large displacement is
524 accumulated. In the central portion of the Oman Mountains, Late Cretaceous convergence ended
525 with the emplacement of the obducted Semail supra-subduction ophiolites and the distal margin
526 sedimentary cover onto the Arabian shelf, besides locking of major thrusts. The Arabian shelf
527 includes both the stable Arabian plate and the proximal domain of its margin, and no evidence for
528 any remarkable Cretaceous thick-skinned tectonics can be found in the area crossed by our section.
529 In this portion of the tectonic pile, the development of major basement thrusts, and related
530 development of the Jabal Akhdar structural culmination are attributed to the second (i.e. Cenozoic)
531 shortening pulse (Hansman et al., 2017). In the frontal portion of the Oman belt, secondary
532 décollement levels developed within the sedimentary cover of the Arabian shelf (Corradetti et al.,
533 2020); however, these décollement levels probably emanate from deeply rooted thrusts and do not
534 connect with the Hawasina basal thrust. Such situation resembles the structural style observed in the
535 southern Apennines, where no remarkable décollement levels occur in the sedimentary cover of the
536 Apulian platform. Indeed, after the emplacement of the thin-skinned tectonic wedge (i.e. including
537 the Liguride, the Apennine Platform, and the Lagonegro basin units) onto the Apulian Platform (i.e.

538 the proximal domain of the margin), the basal décollement directly re-localized into the middle-
539 lower crust. Further shortening has been transferred to the proximal domain by the development of
540 deep-seated, basement-involving thrusts.

541 Apart from the different propagation of the thin-skinned basal décollement in the proximal
542 domain of the Lurestan, Apennine and Oman belts, their common history shows that the
543 involvement of the mid-crustal décollement initiates in the proximal domain once the necking
544 domain enters the subduction channel, marking the transition from soft to hard collision. The switch
545 in shortening style is therefore mainly controlled by the crustal thickness and rheology inherited
546 from the rifted margin. In detail, the inherited rheology corresponds to the transition from thinned
547 crust in the necking domain to normal thickness crust in the proximal domain (Fig. 1B). During
548 shortening of the hyperextended domain, the coupled crust and upper mantle are characterized by a
549 dominating brittle-frictional rheology. During this stage, the main reactivation-prone mechanical
550 weakness is the basement-cover interface, often represented by an exhumation fault soled along
551 hydrated crust or serpentized mantle (e.g., Sutra et al., 2013; Lescoutre and Manatschal, 2020).
552 With the arrival of the necking domain in the subduction zone (Fig. 7), the weak mid crustal level,
553 absent in the hyperextended crust as a result of rifting and crustal thinning (Fig. 7A-C), may
554 localize forelandward-directed simple shear (Fig. 7D-E) and constitute a large interconnected
555 décollement linked with the main subduction interface.

556 Once activated, the mid-crustal décollement propagates forelandward across the preserved
557 middle crust ductile level of the proximal domain (as seen in Oman, Lurestan, and Apennine belts),
558 but also hinterland-ward. The hinterland-ward propagation of the mid-crustal ductile level is also
559 ensured by the increased metamorphic conditions within the subducting thinned continental crust,
560 which allows ductile deformation to occur in the former hyperextended domain of the rift (Fig. 7D-
561 E). At this stage, the subduction interface re-localizes at a deeper structural level, allowing for basal
562 accretion and crustal stacking (as seen in Taiwan), which in turn allows for the internal
563 reorganization of the orogenic wedge by the exhumation of hinterland metamorphic units (Fig. 7D).

564 Notably, thin-skinned thrusting observed in the Kawr (Oman), Bisotun-Avalon (Lurestan),
565 and Apennine Platform (Apennines) extensional ribbons, indicates that their sedimentary cover has
566 been detached from the basement and this basement has not been later involved in thrusting.
567 Therefore, although we have inferred a pre-orogenic ductile middle crust for these extensional
568 ribbons, its activation as a basal décollement - if any - must have been limited (e.g. some low-
569 displacement upper crustal duplexes linking upward with the basement-cover interface and soling
570 down into the middle crust; however, these structures have not been found in any of those three
571 belts). But why the ductile crust within these extensional ribbons does not significantly activate
572 with their arrival at the subduction zone? We can just speculate about two hypotheses: (1) The
573 cross-sectional width of the Bisotun-Avalon and Apennine Platform extensional ribbons, excluding
574 the necking domains bounding them, is largely below 100 km (Figs. 2D, 4D). Accordingly, a few
575 tens of kilometers tract of thick ductile middle crust could be not enough to nucleate a regional
576 ductile décollement level. (2) The ductile crust underneath extensional ribbons could be actually
577 thinner and less interconnected with respect to a “normal” proximal domain. In the end, the
578 extensional ribbons illustrated in this work do not show evidence of thick-skinned thrusting and
579 mid-crustal décollement level activation, because either these blocks are not large enough or their
580 ductile crust is not thick and interconnected enough, or a combination of both.

581

582 **4.3 Rift-inheritances and thrust tectonics**

583 Since the 80's, normal faults reactivated with a reverse kinematics have been documented
584 worldwide (e.g., Williams et al., 1989; Nemčok et al., 1995; Marshak et al., 2000; Zanchi et al.,
585 2006; Carrera et al., 2006). This observation has shaped the concepts of inversion tectonics (e.g.,
586 Glennie and Boegner, 1981; Cohen, 1982; Cooper et al., 1989) and structural inheritances (e.g.,
587 Butler and Mazzoli, 2006, and references therein). It is largely demonstrated that the map outline of
588 orogens partly retraces the shape of the inherited rift systems (e.g., Macedo and Marshak, 1999;
589 Lescoutre and Manatschal, 2020). This overlap is due to various reasons: (i) the reactivation of
590 basin-bounding faults (e.g., Carrera et al., 2006); (ii) the rift architecture control on the lateral extent

591 of sedimentary units that can act as decoupling levels (e.g., Bellahsen et al., 2012; Muñoz et al.,
592 2013); and (iii) the variable thickness of the ductile crust in the different inherited extensional
593 domains that allows/prevents the development of a mid-crustal décollement level (e.g., Lacombe
594 and Bellahsen, 2016; Tavani et al., 2020). The four collisional systems presented in this study,
595 despite the differences and simplifications we have introduced, have allowed us to further expand
596 the last concept of rift-related crustal rheological inheritances. The interaction between evolving
597 rheological properties of the lithosphere and orogen growth has been quantitatively demonstrated by
598 numerical models (e.g., Jamieson and Beaumont, 2013), which have also improved our
599 understanding of how inherited crustal composition, and the resultant variation of crustal strength,
600 may control orogenic processes (e.g., Jammes et al., 2014). In line with these numerical results, we
601 have demonstrated the correspondence between the evolution of collisional systems, from thin-
602 skinned to thick-skinned and eventually to underplating and crustal stacking, and the rheology of
603 the extensional domain subducted at each stage of plate convergence.

604

605 **5 CONCLUSIONS**

606 Our review of geological data from four orogenic systems reactivating their former rifted
607 margins shows a close link between the evolution of the orogenic system and the inherited crustal
608 architecture and related rheology. In all the reviewed systems, the first stage of collision
609 corresponds to the activation of a thin-skinned décollement level, along which the sedimentary
610 cover of the distal portions of the former rifted margin has been detached from its basement and has
611 been placed onto the former proximal domain. This level forms in the previously thinned domain of
612 the margin, where syn-rift coupling between the brittle upper crust and the brittle upper mantle
613 prevents the syn-orogenic formation of a ductile crustal layer beneath the sedimentary cover.
614 Arrival of the necking and proximal domains of the rifted margin (in which the ductile crust is
615 preserved) at the subduction zone allows for the reactivation of the ductile middle crust as a
616 décollement level. This triggers the switch to thick-skinned tectonics. Activation of this second and
617 deeper décollement causes the re-deformation of the former thin-skinned décollement. Eventually,

618 the mid-crustal décollement links to the main subduction interface, promoting underplating, crustal
619 stacking, and exhumation of (metamorphic) units from the interior of the belt.

620

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625

626

627 **FIGURE CAPTIONS**

628

629 **Figure 1.** Scheme showing forward décollement propagation for (A) layer-cake lithosphere, and (B)
630 magma-poor rifted margins. Rheological (strength) profiles are after Lescoutre and Manatschal
631 (2020).

632

633 **Figure 2.** Geology of the Lurestan Arc area. (A) Simplified geological map of the study area, with
634 inset showing the broader structural context of Zagros. (B) Schematic stratigraphy of the rocks
635 involved in the Zagros orogeny. (C) Crustal section across the Lurestan Arc area (modified after
636 Tavani et al, 2020, and Vergés et al., 2011; Moho depth after Jiménez-Munt et al., 2012; located in
637 A). (D) Schematic representation of the Arabian rifted margin during Late Cretaceous (after
638 Wrobel-Daveau et al., 2010).

639

640 **Figure 3.** (A) Geological cross-section across the inner part of the simply folded Lurestan belt (see
641 Fig. 2A for location) with stratigraphic succession exposed in the area. (B) Thin-section of the base
642 of the Garau Fm., showing a radiolarian packstone (35.04912°N; 46.16426°E). (C) Field photo
643 illustrating exposure of the Garau Fm in the northeasternmost sector of the study area, with thick
644 intervals of radiolarites (35.10778°N; 46.20355°E).

645

646 **Figure 4.** Geology of the southern Apennines. (A) Simplified geological map of the study area
647 (modified after Ascione et al., 2012), with inset showing the structural framework of Italy. (B)
648 Schematic stratigraphy of the different paleogeographic domains of Adria. (C) Crustal cross-section
649 across the southern Apennines (modified after Mazzoli et al., 2014; location in A). The depth and
650 thickness of the middle crust is speculative. (D) Schematic representation of the Adria rifted margin
651 during Late Cretaceous.

652

653 **Figure 5.** Geological framework of the Oman mountains. (A) Simplified geological map of the Al-
654 Hajar mountains, showing trace of geological section (modified after Corradetti et al., 2020). (B)
655 Schematic stratigraphy of the rocks involved in the Oman orogeny. (C) Crustal cross-section across
656 Jabal Akhdar. The depth and thickness of the middle crust is speculative. (D) Schematic
657 representation of the Arabian rifted margin prior to the onset of convergence during Late
658 Cretaceous times. MU (Musandam), JA (Jabal Akhdar) and SH (Saih Hatat) are the three main
659 tectonic windows exposing the Arabian Platform domain.

660

661 **Figure 6.** Geology of the Taiwan Belt. (A) Simplified geological map of the Taiwan belt. (B)
662 Schematic stratigraphy of the area. (C) Crustal cross-section (modified after Brown et al., 2012;
663 location in A). The depth and thickness of the middle crust and mid-crustal stacking in the interior
664 of the belt are speculative. (D) Schematic representation of the Eurasian rifted margin during
665 Miocene times.

666

667 **Figure 7.** Schematic drawings of the evolutionary scenario by which a rifted continental margin
668 becomes progressively involved in the deformation above a foreland-propagating, branching basal
669 décollement. (A) Onset of shortening and localization of the deformation along a basal décollement
670 confined within the sedimentary cover of the margin's hyperextend domain (i.e., deep-water facies).
671 (B) Bifurcation of the basal décollement, and involvement in the deformation of the sedimentary

672 cover of the margin's necking domain (i.e., shallow water to slope sediments) at multiple structural
673 levels. (C) Onset of inversion of pre-existing extensional faults located at the margin's necking
674 domain and basement-involved deformation. (D) Foreland- and deep-ward propagation of the basal
675 décollement into mid-crustal structural levels of the margin's necking to proximal domains, coeval
676 with inception of high-pressure metamorphism and re-thickening of the ductile layer in the interior
677 of the belt. (E) Progressive shortening above a mid-crustal décollement, with development of the
678 basement ramps in the foreland part of the developing mountain belt and junction of the mid-crustal
679 décollement with the subduction interface. (F) Final stage of the belt evolution, with basal
680 accretion, nappe stacking and exhumation of the metamorphic units in the interior of the belt.
681

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Figure 1

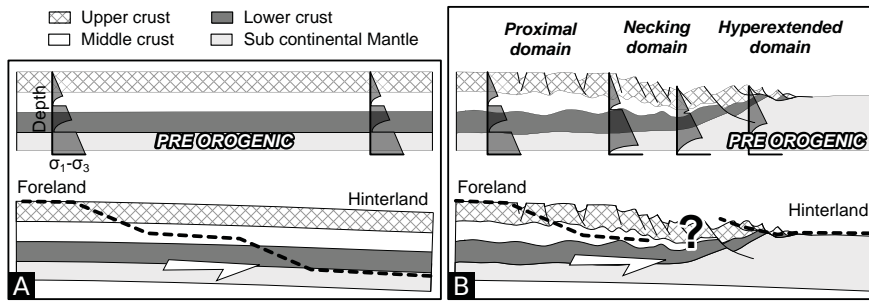


Figure 1
2 columns in 3 column layout

Figure 2

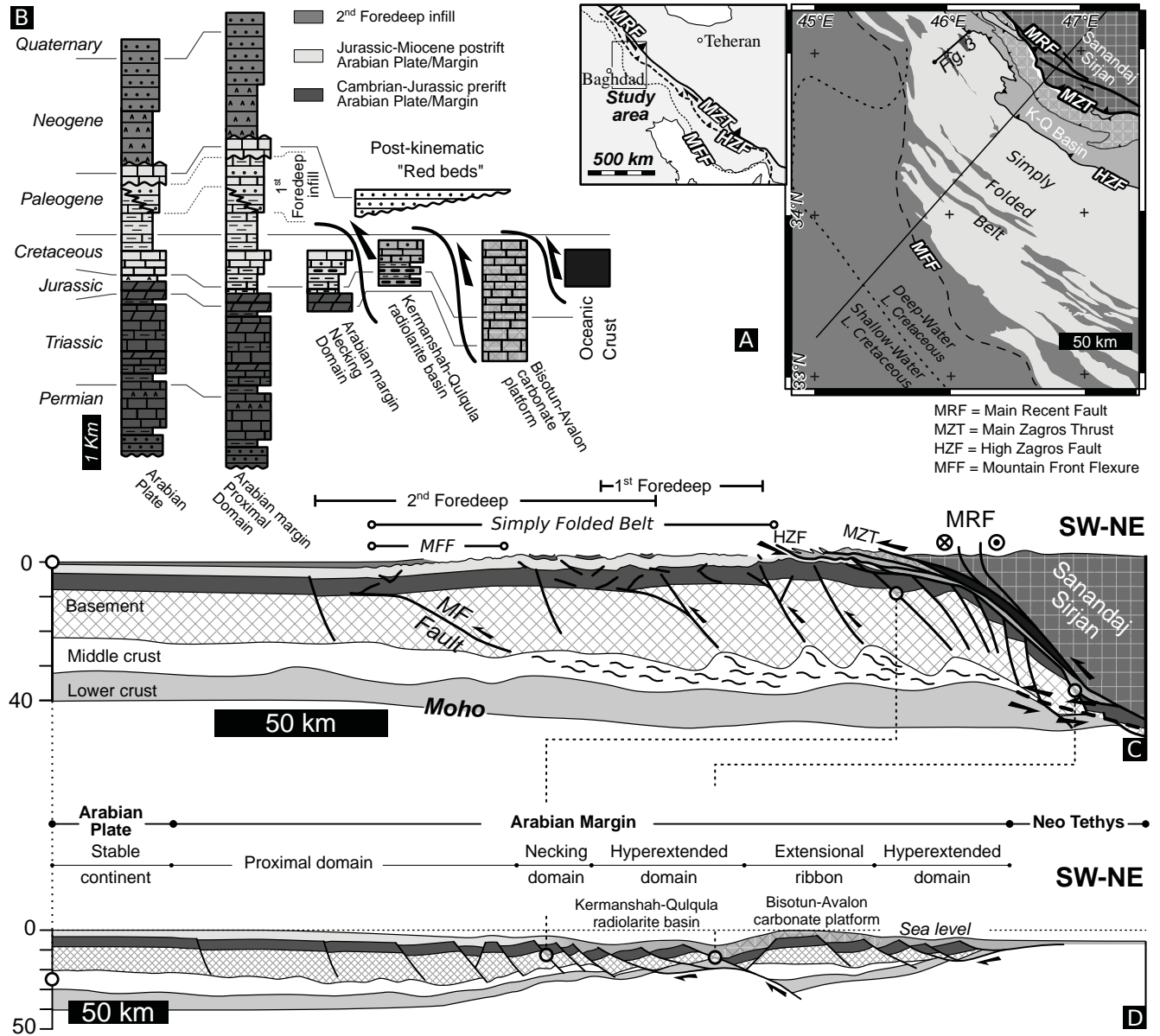


Figure 2
3 columns in 3 column layout

Figure 3

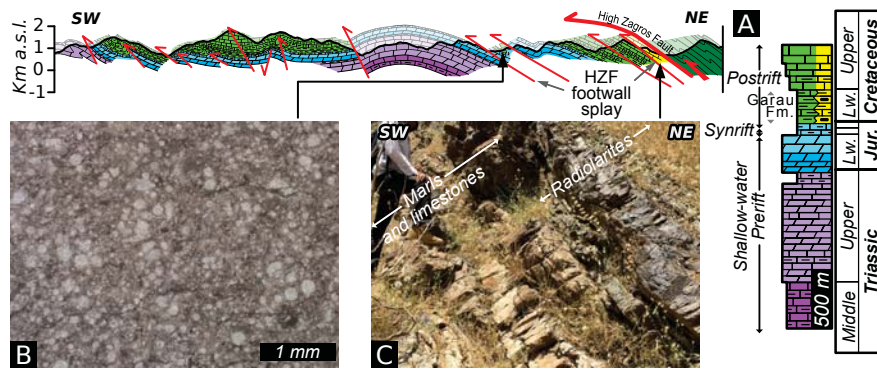


Figure 3
2 columns in 3 column layout

Figure 4

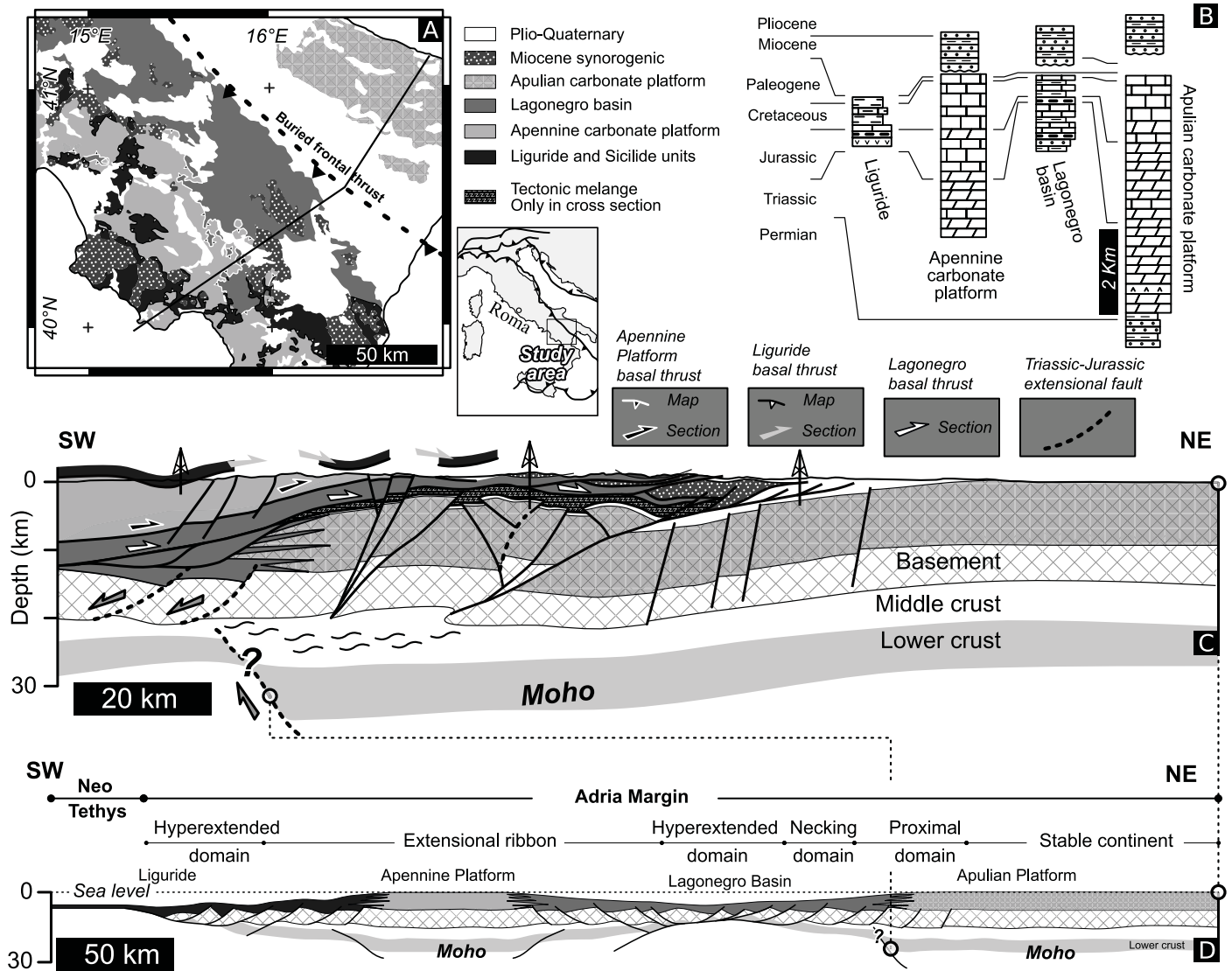


Figure 4
3 columns in 3 column layout

Figure 5

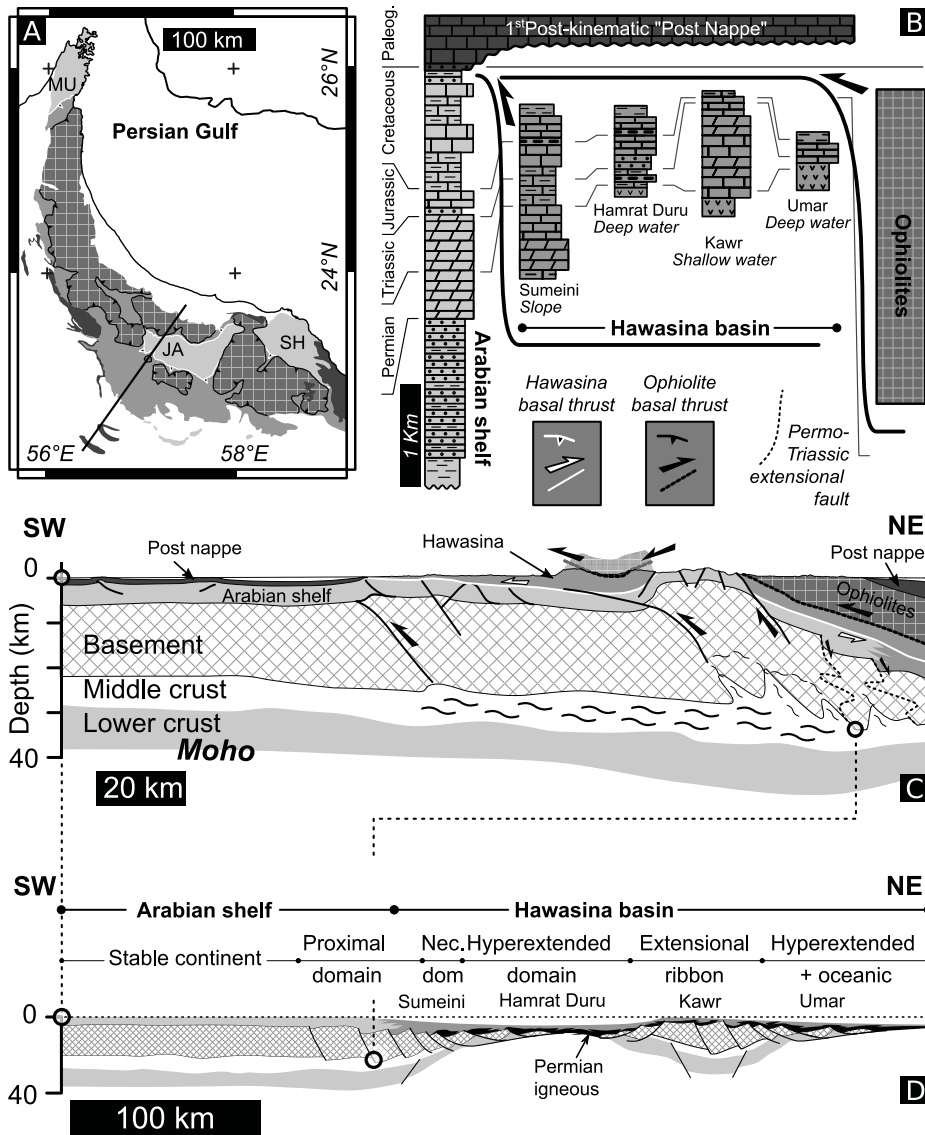


Figure 5
2 columns in 3 column layout

Figure 6

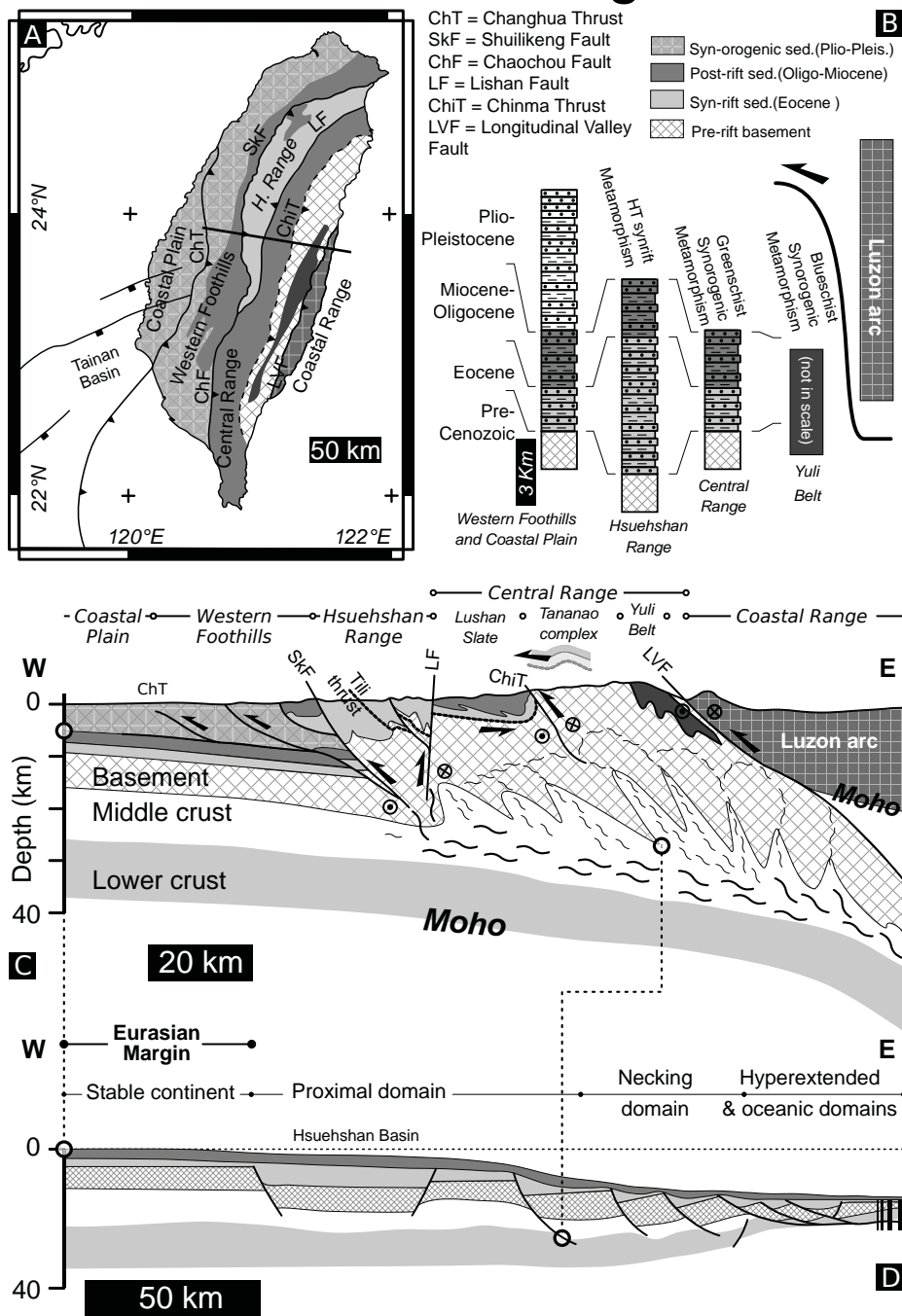


Figure 6
 2 columns in 3 column layout

Figure 7

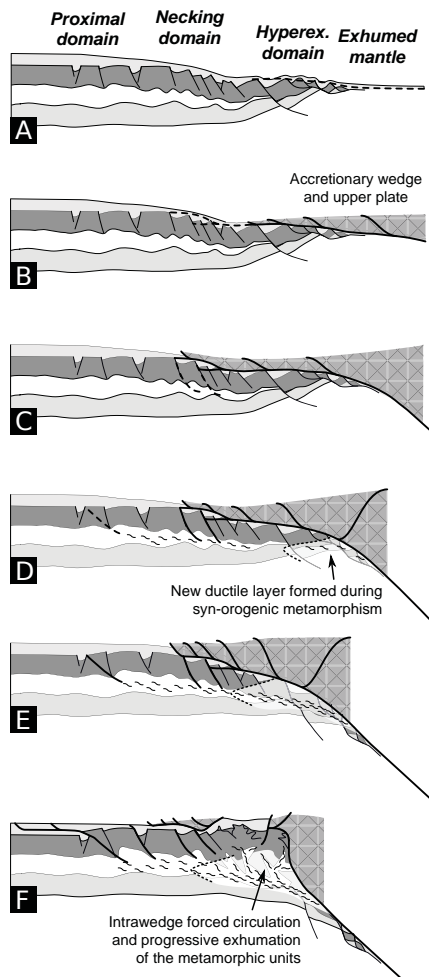


Figure 7
1 column in 3 column layout