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

4

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#### 20 ABSTRACT

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

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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 deep-seated décollements during the inception of shortening. Progressive orogenic involvement of 36 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 eventually links to the main subduction interface, favoring underplating and the upward motion of 39 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 of paramount importance in the Wilson cycle (e.g., Wilson, 1966; Jackson, 1980; Cohen, 1982). 45 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 located either within the lower plate sedimentary cover or within its middle-lower crust (e.g., 57 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 earthquake, Tavani et al., 2018a), which requires a basal décollement re-localization. Such process 62 apparently contradicts the simple notion that hinterland basement thrusts climb up-section foreland-63 ward (Fig. 1A), but points out that the inherited crustal rheology across the previous rifted margin 64 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) (Fig. 1B). Consistently, in several mountain belts involving former rifted margins, the distal 67 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 (Schmid et al., 2004), the Taiwan belt (Mouthereau and Lacombe, 2006; Brown et al, 2012; 72 73 Lacombe and Bellahsen, 2016), and the Zagros belt in the Lurestan region (Vergés et al., 2011) are some well-known examples of this structural evolution. The architecture of these orogenic systems 74 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.

Here we review information from several collisional systems, with an emphasis on the role of inherited rift-related structures, to establish with unprecedented resolution a direct correlation between the structure of collisional orogens and the detailed architecture of precursor rifted 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**

This work is organized into two main sections: case studies (section 3) and discussion (section 4). 85 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 the inferred architecture of the inherited rifted margin. Each case study section is further divided 88 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 hanging wall is mostly composed of the intensely folded and shortened deep-water sedimentary cover of the distal portions of the Arabian rifted margin. The Simply Folded Belt (SFB) in the footwall of the HZF extends to the Mountain Front Flexure (MFF) (Fig. 2A), a major topographic feature controlled by the underlying basement-involved Mountain Front Fault (e.g., Berberian and 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 and Miocene-to-recent age (Homke et al. 2009; Saura et al., 2015). The older stage involved the 115 116 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 (iii) the Kermanshah-Qulqula nappe (e.g., Vergés et al., 2011). The Bisotun-Avalon nappe is made 119 120 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, marls and limestones deposited in the Kermanshah-Oulqula basin (Gharib and De Wever, 2010); 125 126 this basin was originally interposed between the Arabian margin and the Bisotun-Avalon carbonate platform (Fig. 2D), and partly developed on a exhumed mantle domain (Wrobel-Daveau et al., 127 128 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 along sub-horizontal décollements located within the Triassic (Bisotun-Avalon nappe) and the 133 Lower Jurassic (Kermanshah-Oulgula nappe) successions. Further south, footwall splays of the 134

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

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

Apart from the splays of the HZF (Fig. 3A), the proximal domain of the Arabian margin was 143 deformed in a piggy-back sequence since the Miocene (Barber et al., 2018). The basement-cover 144 145 interface of the SFB is recognized as a regional mechanical weakness, which promoted partial decoupling between the sedimentary cover and the crystalline basement (Vergés et al., 2011; Tavani 146 147 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). Steeply-dipping basement thrusts are rooted into a mid-crustal décollement at about 20-25 km depth 149 150 (e.g., Vergés et al., 2011), and many of them are interpreted as inverted extensional faults (Tavani et al., 2018a). The occurrence of the mid-crustal ductile décollement level is also evidenced by a 151 seismic gap occurring between 20 and 30 km depth (e.g., Talebian and Jackson, 2004; Tavani et al., 152 153 2020). The Mountain Front Fault is a moderate- to gently-dipping active thrust at the tip of the midcrustal décollement, with a displacement in the order of 10 km (Vergés et al., 2011; Tavani et al., 154 155 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 conditions persisted until the Late Cretaceous onset of convergence. Middle to Upper Jurassic rocks 168 169 are poorly exposed in the SFB, hence a clearer marker of the drowned Jurassic tracts is provided by the extensively exposed Lower Cretaceous facies, which are here used to infer the limits between 170 171 the different domains of the rifted margin.

In detail, three domains can be recognized (Fig. 2B). To the SW, shallow-water carbonate 172 173 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 Lower Cretaceous Garau Fm., which is made up of limestones, marls, and shales (Jassim and Goff, 177 2006). Radiolarian-rich beds in its lower portion are well known (Fig. 3B) and testify for a 178 179 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-Daveau et al., 2010). The width of the Kermanshah-Qulqula radiolarite basin and of the Bisotun-186 Avalon extensional ribbon are from Verges et al. (2011). 187

#### 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 rifted margin, i.e. the Adria promontory of Africa. Remnants of the overriding plate are extensively 195 196 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 horst and graben/half-graben structures formed upon rifting, in which carbonate platforms and 205 206 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).

During the Eurasia-Adria Cenozoic convergence, deep-water sediments and slices of ophiolites were scraped off from the downgoing Tethys oceanic plate and accreted to form the Liguride complex (Ogniben, 1969; Knott, 1987). During the middle Miocene, this accretionary complex was thrust eastward for tens of kilometers and placed on top of the Apennine Platform and 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-Mesozoic basement and placed on top of the Apulian Platform (Roure et al., 1991; Menardi 222 223 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 Noguera and Rea, 2000). This slope domain is currently in the footwall to the Lagonegro basin 231 232 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 and Rea, 2000; Di Bucci et al., 2006; Mazzoli et al., 2014). It is noteworthy that thick-skinned 234 235 thrusting has not developed in the Apennine Platform, which was originally interposed between two areas of hyper-extended crust. Deeply rooted thrusts linked to a deep crustal décollement have 236 237 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 of the Apennine carbonate platform is estimated to be nearly 60 km (Menardi Noguera and Rea, 248 249 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 of the coastline (e.g., Capri Island, ISPRA 2014), the maximum original width for the Apennine 252 253 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 (Menardi Noguera and Rea, 2000). However, we propose a different interpretation coherent with 257 the main message of this work and contractional slip transfer considerations. The step is here 258 259 interpreted as a NE-dipping lower crust extensional shear zone related to Mesozoic crustal thinning. Our interpretation is strongly supported by the two following observations: (i) the Moho step is 260 261 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 plus the middle and lower crust); such a crustal thinning requires significant extension in the area. 266 Within this framework, the Moho step would be placed within or immediately to the NE of the 267

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

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#### 273 **3.3 Oman**

#### 274 3.3.1 Geological setting

The Oman mountain belt (also known as Al-Hajar mountains) developed following the Late 275 Cretaceous closure of the Neo-Tethys ocean between the Arabian and Eurasian plates, and by 276 subsequent shortening during the Cenozoic (Glennie et al., 1973; Tarapoanca et al., 2010; Hansman 277 278 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 95-93 Ma (Hacker, 1994), when the Semail ophiolites were thrust over the various domains of the 284 distal portion of the former Arabian rifted margin. The highly deformed and partly dismembered 285 286 remnants of the Arabian margin are presently sandwiched between the Semail ophiolites and the Arabian shelf, which includes the stable continent and the proximal domain of the rift. The 287 288 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 distal domains of the Arabian margin was progressively detached from its basement, incorporated 291 292 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 conformable over the foredeep sediments to the SW. 302

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 by thermochronological data (Mount et al., 1998; Saddiqi et al., 2006; Tarapoanca et al., 2010; 307 308 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; Jacobs et al., 2015; Hansman et al., 2017), and a later reactivation at 20-15 Ma (Jacobs et al., 2015; 312 313 Corradetti et al., 2020). Eocene uplift and unroofing relate to a major shortening event that affected the whole nappe edifice, triggering flexuring of the lower plate (e.g., Homewood et al., 1986) and 314 315 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., 1990; Tarapoanca et al., 2010; Hansman et al., 2017). Notably, in the area crossed by our section no 318 319 evidence of thick-skinned tectonics during the Cretaceous stage has been documented so far for the 320 Arabian shelf.

321

323 The Arabian rifted margin in the Oman region formed during the Permo-Triassic magmarich rifting (e.g., Robertson and Searle, 1990). The different paleogeographic domains of the 324 margins are the Arabian Shelf to the SW, and the Hawasina basin to the NE, the latter including the 325 Sumeini slope, and the Hamrat Duru and Umar deep-water basins separated by remnants of 326 327 platform carbonates such as the Kawr Group (Fig. 5D) (Béchennec et al., 1988; Rabu et al., 1993). Accordingly, we here interpret these paleogeographic domains in terms of structures of a rifted 328 margin. In detail, the reconstruction of the Arabian margin includes, from SW to NE: (i) the 329 330 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 radiolarities 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 domain; (iii) the distal portion of the margin, mostly made by the deep-water sediments resting on 338 top of Permian-Triassic syn-rift volcanic rocks. In detail, the latter domain is divided in two basins: 339 340 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 (Béchennec 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,

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

354

#### 355 **3.4 The Taiwan mountain belt**

#### 356 3.4.1 Geological setting

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

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 the north (Clark et al., 1993; Brown et al., 2012) and against the Western Foothills across the 383 Chaochou Fault in the south (Mouthereau et al., 2002; Tang et al., 2011, Fig. 6A). The Hsuehshan 384 Range comprises variably metamorphosed Eocene and Oligocene clastics (Beyssac et al., 2007; 385 386 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 (Beyssac 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 deformed Pliocene to Holocene syn-orogenic sediments of the foreland basin. 393

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 velocities taking place along deeply rooted clusters of earthquake hypocenters in the interior of the 399 400 mountain belt (Gourley et al., 2007; Wu et al., 2007; Kuo-Chen et al., 2012; Brown et al., 2017), indicate that basement is uplifted along steeply dipping faults that penetrate into the middle crust 401 402 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

Prior to collision in Eocene times, the Eurasian margin had the structure of a 422 hyperextended margin (Brown et al., 2012; McIntosh et al., 2013) (Fig. 6D). In the cross-section of 423 424 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 different rift domains. In detail, in figure 6D we show a slightly modified version of the 428 429 reconstruction by Brown et al. (2012). In particular, albeit with some uncertainty, the necking

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

435

#### 436 4 DISCUSSION

This discussion is organized in three sub-sections. Firstly, we recall the structure of the four illustrated rifted margins, stressing their architecture and the recognition of different structural domains. Then, we discuss the process of décollement re-localization and, finally, we discuss our 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; Lavier and Manatschal, 2006; Péron-Pinvidic and Manatschal 2009) have recently been recognized 447 448 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 serpentinized 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

Garau Fm. there includes meter-thick radiolarites (Figs. 2,3) pointing to the proximity of the CCD.
To the SW, the area labeled as "Arabian Plate" is the region devoid of any significant crustal
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 Jurassic magma-poor rifted margin (e.g., Mohn et al., 2012). Deep water radiolarites occur in the 465 central portion of the Lagonegro basin (e.g., Mazzoli et al., 2001) and the observed thinning of the 466 467 crust down to ~20 km at the transition with the Apulian Platform (Fig. 4C) allows interpreting at least the central part of the Lagonegro basin as a hyperextended domain. The Apennine Platform, in 468 469 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 framework the Apulian Platform would represent the proximal rift domain. 473

474

#### 475 *4.1.3 Oman*

The former Arabian rifted margin in the Oman Mts. slightly differs from that of the 476 Lurestan, in terms of both timing and magma supply. The timing of rifting is Permian-Triassic, as 477 478 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 Duru and Umar domains, were most likely also deposited on an extensional ribbon. Radiolarites 482 483 belonging to the Hamrat Duru and Umar domains suggest a paleodepth close to the CCD and a remarkable crustal thinning for these domains. Consistently, the crustal-scale cross-section in Figure 5C shows that the pre-orogenic crust at the Sumeini slope to Arabian shelf transition, i.e. in the thicker portion of the Hawasina basin, should have been nearly 20 km thick. Accordingly, we infer hyper-extension in the Hamrat Duru domain, which was placed in a more distal position with respect to the 20 km thick crust of the Sumeini slope. The substratum of the Umar sub-basin is 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 hyperextended domains; however, neither the basement nor the respective sedimentary covers of 495 496 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**

After rifted margin development and during the subsequent plate convergence, the distal 503 504 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-Oulgula 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.

As collision evolves, slightly different evolutions can be defined based on the examples 515 above. In the Lurestan, the thin-skinned basal décollement propagated from the hyperextended 516 517 domain into the necking domain across weak stratigraphic levels. Further shortening and convergence (re-)activated the inherited basement faults of the necking and proximal domains. The 518 519 first stage of this thick-skinned deformation is characterized by limited shortening accommodated by inversion tectonics (Granado and Ruh, 2019; Tavani et al., 2020). In the latest stage of 520 521 deformation, the seismically active low-angle Lurestan Mountain Front Fault developed, with a cumulative displacement of nearly 10 km. The activity of this fault testifies the presence of an 522 523 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 includes both the stable Arabian plate and the proximal domain of its margin, and no evidence for 527 any remarkable Cretaceous thick-skinned tectonics can be found in the area crossed by our section. 528 529 In this portion of the tectonic pile, the development of major basement thrusts, and related development of the Jabal Akhdar structural culmination are attributed to the second (i.e. Cenozoic) 530 531 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 Apulian platform. Indeed, after the emplacement of the thin-skinned tectonic wedge (i.e. including 536 the Liguride, the Apennine Platform, and the Lagonegro basin units) onto the Apulian Platform (i.e. 537

the proximal domain of the margin), the basal décollement directly re-localized into the middlelower crust. Further shortening has been transferred to the proximal domain by the development of 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 in shortening style is therefore mainly controlled by the crustal thickness and rheology inherited 545 from the rifted margin. In detail, the inherited rheology corresponds to the transition from thinned 546 crust in the necking domain to normal thickness crust in the proximal domain (Fig. 1B). During 547 548 shortening of the hyperextended domain, the coupled crust and upper mantle are characterized by a dominating brittle-frictional rheology. During this stage, the main reactivation-prone mechanical 549 550 weakness is the basement-cover interface, often represented by an exhumation fault soled along 551 hydrated crust or serpentinized mantle (e.g., Sutra et al., 2013; Lescoutre and Manatschal, 2020). With the arrival of the necking domain in the subduction zone (Fig. 7), the weak mid crustal level, 552 553 absent in the hyperextended crust as a result of rifting and crustal thinning (Fig. 7A-C), may localize forelandward-directed simple shear (Fig. 7D-E) and constitute a large interconnected 554 555 décollement linked with the main subduction interface.

556 Once activated, the mid-crustal décollement propagates forelandward across the preserved middle crust ductile level of the proximal domain (as seen in Oman, Lurestan, and Apennine belts), 557 558 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-E). At this stage, the subduction interface re-localizes at a deeper structural level, allowing for basal 561 562 accretion and crustal stacking (as seen in Taiwan), which in turn allows for the internal reorganization of the orogenic wedge by the exhumation of hinterland metamorphic units (Fig. 7D). 563

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. Therefore, although we have inferred a pre-orogenic ductile middle crust for these extensional 567 ribbons, its activation as a basal décollement - if any - must have been limited (e.g. some low-568 569 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 belts). But why the ductile crust within these extensional ribbons does not significantly activate 571 572 with their arrival at the subduction zone? We can just speculate about two hypotheses: (1) The cross-sectional width of the Bisotun-Avalon and Apennine Platform extensional ribbons, excluding 573 574 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 ductile décollement level. (2) The ductile crust underneath extensional ribbons could be actually 576 577 thinner and less interconnected with respect to a "normal" proximal domain. In the end, the extensional ribbons illustrated in this work do not show evidence of thick-skinned thrusting and 578 579 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**

Since the 80's, normal faults reactivated with a reverse kinematics have been documented 583 worldwide (e.g., Williams et al., 1989; Nemčok et al., 1995; Marshak et al., 2000; Zanchi et al., 584 585 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., Butler and Mazzoli, 2006, and references therein). It is largely demonstrated that the map outline of 587 588 orogens partly retraces the shape of the inherited rift systems (e.g., Macedo and Marshak, 1999; Lescoutre and Manatschal, 2020). This overlap is due to various reasons: (i) the reactivation of 589 590 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 the last concept of rift-related crustal rheological inheritances. The interaction between evolving 596 597 rheological properties of the lithosphere and orogen growth has been quantitatively demonstrated by numerical models (e.g., Jamieson and Beaumont, 2013), which have also improved our 598 understanding of how inherited crustal composition, and the resultant variation of crustal strength, 599 may control orogenic processes (e.g., Jammes et al., 2014). In line with these numerical results, we 600 601 have demonstrated the correspondence between the evolution of collisional systems, from thinskinned to thick-skinned and eventually to underplating and crustal stacking, and the rheology of 602 603 the extensional domain subducted at each stage of plate convergence.

604

#### 605 5 CONCLUSIONS

Our review of geological data from four orogenic systems reactivating their former rifted 606 margins shows a close link between the evolution of the orogenic system and the inherited crustal 607 architecture and related rheology. In all the reviewed systems, the first stage of collision 608 609 corresponds to the activation of a thin-skinned décollement level, along which the sedimentary cover of the distal portions of the former rifted margin has been detached from its basement and has 610 611 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, crustal619 stacking, and exhumation of (metamorphic) units from the interior of the belt.

620

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- 625
- 626

#### 627 FIGURE CAPTIONS

628

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

632

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

639

**Figure 3.** (A) Geological cross-section across the inner part of the simply folded Lurestan belt (see Fig. 2A for location) with stratigraphic succession exposed in the area. (B) Thin-section of the base of the Garau Fm., showing a radiolarian packstone (35.04912°N; 46.16426°E). (C) Field photo illustrating exposure of the Garau Fm in the northeasternmost sector of the study area, with thick intervals of radiolarites (35.10778°N; 46.20355°E). **Figure 4**. Geology of the southern Apennines. (A) Simplified geological map of the study area (modified after Ascione et al., 2012), with inset showing the structural framework of Italy. (B) Schematic stratigraphy of the different paleogeographic domains of Adria. (C) Crustal cross-section across the southern Apennines (modified after Mazzoli et al., 2014; location in A). The depth and thickness of the middle crust is speculative. (D) Schematic representation of the Adria rifted margin during Late Cretaceous.

652

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

660

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

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

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

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#### 682 **REFERENCES CITED**

- Ascione, A., Ciarcia, S., Di Donato, V., Mazzoli, S., and Vitale, S., 2012, The Pliocene-Quaternary
   wedge-top basins of southern Italy: an expression of propagating lateral slab tear beneath the
- 685 Apennines: Basin Research, v. 24, p. 456–474, doi:10.1111/j.1365-2117.2011.00534.x.
- 686 Ballato, P., Uba, C.E., Landgraf, A., Strecker, M.R., Sudo, M., Stockli, D.F., Friedrich, A., and
- 687 Tabatabaei, S.H., 2011, Arabia-Eurasia continental collision: Insights from late Tertiary
- 688 foreland-basin evolution in the Alborz Mountains, northern Iran: Geological Society of
- 689 America Bulletin, v. 123, p. 106–131, doi:10.1130/B30091.1.
- 690 Barber, D.E., Stockli, D.F., Horton, B.K., and Koshnaw, R.I., 2018, Cenozoic Exhumation and
- 691 Foreland Basin Evolution of the Zagros Orogen During the Arabia-Eurasia Collision, Western
- 692 Iran: Tectonics, v. 37, p. 4396–4420, doi:10.1029/2018TC005328.
- Barrier, E., and Vrielynck, B., 2008, Palaeotectonic map of the Middle East, Atlas of 14 maps,
- tectonosedimentary-palinspastic maps from Late Norian to Pliocene: Commission for theGeologic Map of the World (CCMW, CCGM), Paris, France,.
- Béchennec, F., Le Métour, J., Rabu, D., Villey, M., and Beurrier, M., 1988, The Hawasina Basin: A
  fragment of a starved passive continental margin, thrust over the Arabian Platform during
- obduction of the Sumail Nappe: Tectonophysics, v. 151, p. 323-343. doi:10.1016/0040-
- 699 1951(88)90251-X.
- 700 Bellahsen, N., Jolivet, L., Lacombe, O., Bellanger, M., Boutoux, A., Garcia, S., Mouthereau, F., Le
- 701 Pourhiet, L., and Gumiaux, C., 2012, Mechanisms of margin inversion in the external Western
- Alps: Implications for crustal rheology: Tectonophysics, v. 560–561, p. 62–83,
- 703 doi:10.1016/j.tecto.2012.06.022.
- Berberian, M., and King, G.C.P., 1981, Towards a paleogeography and tectonic evolution of Iran:
  Canadian Journal of Earth Sciences, v. 18, p. 210–265, doi:10.1139/e81-019.
- 706 Beyssac, O., Simoes, M., Avouac, J.P., Farley, K.A., Chen, Y.-G., Chan, Y.-C., and Goffé, B.,
- 707 2007, Late Cenozoic metamorphic evolution and exhumation of Taiwan: Tectonics, v. 26,
- 708 doi:10.1029/2006TC002064.

- 709 Boote, D.R.D., Mou, D., and Waite, R.I., 1990, Structural evolution of the Suneinah Foreland,
- 710 Central Oman Mountains: Geological Society, London, Special Publications, v. 49, p. 397–

711 418, doi:10.1144/GSL.SP.1992.049.01.25.

- 712 Breton, J.P., Béchennec, F., Le Métour, J., Moen-Maurel, L., and Razin, P., 2004, Eoalpine
- 713 (Cretaceous) evolution of the Oman Tethyan continental margin: Insights from a structural
- field study in Jabal Akhdar (Oman Mountains): GeoArabia, v. 9, p. 41–58.
- 715 Brown, D., Alvarez-Marron, J., Schimmel, M., Wu, Y.-M., and Camanni, G., 2012, The structure
- and kinematics of the central Taiwan mountain belt derived from geological and seismicity

717 data: Tectonics, v. 31, doi:10.1029/2012TC003156.

- 718 Brown, D., Alvarez-Marron, J., Biete, C., Kuo-Chen, H, Camanni, G., Ho, C.-W., 2017, How the
- structural architecture of the Eurasian continental margin affects the structure, seismicity, and
- topography of the south-central Taiwan fold and thrust belt. Tectonics,
- 721 doi:10.1002/2017TC004475.
- Butler, R.W.H., and Mazzoli, S., 2006, Styles of continental contraction: A review and introduction,
   *in* Styles of Continental Contraction, Geological Society of America, v. 80301, p. 1–10,
- 724 doi:10.1130/2006.2414(01).
- 725 Byrne, T., Chan, Y.C., Rau, R.J., Lu, C.Y., Lee, Y.H., and Wang, Y.J., 2011, The Arc–Continent
- 726 Collision in Taiwan, *in* Brown, D. and Ryan, P.D. eds., Arc-Continent Collision, Berlin,
- Heidelberg, Springer Berlin Heidelberg, p. 213–245.
- Carrera, N., Muñoz, J.A., Sàbat, F., Mon, R., and Roca, E., 2006, The role of inversion tectonics in
- the structure of the Cordillera Oriental (NW Argentinean Andes): Journal of Structural
- 730 Geology, v. 28, p. 1921–1932, doi:10.1016/j.jsg.2006.07.006.
- 731 Chenin, P., Manatschal, G., Picazo, S., Müntener, O., Karner, G., Johnson, C., & Ulrich, M., 2017,
- 732 Influence of the architecture of magma-poor hyperextended rifted margins on orogens
- produced by the closure of narrow versus wide oceans. Geosphere, v. 13, p. 559-576,
- 734 doi:10.1130/GES01363.1
- 735 Chemenda, A.I., Mattauer, M., and Bokun, A.N., 1996, Continental subduction and a mechanism

/36 for exhumation of high-pressure metamorphic rocks: new modelling and field dat
--

737 Oman: Earth and Planetary Science Letters, v. 143, p. 173–182, doi:10.1016/0012-

738 821X(96)00123-9.

- 739 Chen, W.-S., Chung, S.-L., Chou, H.-Y., Zugeerbai, Z., Shao, W.-Y., and Lee, Y.-H., 2017, A
- reinterpretation of the metamorphic Yuli belt: Evidence for a middle-late Miocene accretionary
- 741 prism in eastern Taiwan: Tectonics, v. 36, p. 188–206, doi:10.1002/2016TC004383.
- 742 Clark, M.B., Fisher, D.M., Lu, C.-Y., and Chen, C.-H., 1993, Kinematic analyses of the Hsüehshan
- Range, Taiwan: A large-scale pop-up structure: Tectonics, v. 12, p. 205–217,
- 744 doi:10.1029/92TC01711.
- 745 Cohen, C.R., 1982, Model for a passive to active continental margin transition: implications for
- hydrocarbon exploration.: American Association of Petroleum Geologists Bulletin, v. 66, p.
- 747 708–718, doi:10.1306/03b5b6f5-16d1-11d7-8645000102c1865d.
- Cooper, D.J.W., 1988, Structure and sequence of thrusting in deep-water sediments during ophiolite
  emplacement in the south-central Oman Mountains: Journal of Structural Geology, v. 10, p.
- 750 473–485, doi:10.1016/0191-8141(88)90035-1.
- 751 Cooper, D.J.W., Ali, M.Y., and Searle, M.P., 2014, Structure of the northern Oman Mountains from
- the Semail Ophiolite to the Foreland Basin: Geological Society, London, Special Publications,
- v. 392, p. 129–153, doi:10.1144/SP392.7.
- 754 Cooper, M.A., Williams, G.D., de Graciansky, P.C., Murphy, R.W., Needham, T., de Paor, D.,
- 55 Stoneley, R., Todd, S.P., Turner, J.P., and Ziegler, P.A., 1989, Inversion tectonics a
- discussion: Geological Society, London, Special Publications, v. 44, p. 335–347,
- 757 doi:10.1144/GSL.SP.1989.044.01.18.
- 758 Corradetti, A., Spina, V., Tavani, S., Ringenbach, J.C., Sabbatino, M., Razin, P., Laurent, O.,
- 759 Brichau, S., and Mazzoli, S., 2020, Late-stage tectonic evolution of the Al-Hajar Mountains,
- 760 Oman: new constraints from Palaeogene sedimentary units and low-temperature
- thermochronometry: Geological Magazine, v. 157, p. 1031-1044,
- 762 doi:10.1017/S0016756819001250.

- Cosentino, D., Cipollari, P., Marsili, P., and Scrocca, D., 2010, Geology of the central Apennines: a
  regional review: Journal of the Virtual Explorer, v. 36.
- 765 Curzi, M., Billi, A., Carminati, E., Rossetti, F., Albert, R., Aldega, L., Cardello, L., Conti, A.,
- 766 Gerdes, A., Smeraglia, L., Van der Lelij, R., Vignaroli, G., & Viola, G. (2020). Disproving the
- 767 presence of Paleozoic-Triassic metamorphic rocks on the Island of Zannone (central Italy):
- Implications for the early stages of the Tyrrhenian-Apennines tectonic evolution. Tectonics,
   e2020TC006296.
- D'Argenio, B., Pescatore, T., and Scandone, P., 1975, Structural pattern of the Campania-Lucania
  Apennines, *in* Structural Model of Italy, Quaderni Ricerca Scientifica CNR, v. 90, p. 313–327.
- 772 De Wever, P., Azema, J., and Fourcade, E., 1994, Radiolaires et radiolarites : production primaire,
- diagenèse et paléogéographie; Radiolarians and radiolarites : primary production, diagenesis
- and paleogeography: Bulletin des centres de recherches exploration Production Elf-
- 775 Aquitaine, v. 18, p. 315–379.
- Dewey, J.F., Helman, M.L., Knott, S.D., Turco, E., and Hutton, D.H.W., 1989, Kinematics of the
  western Mediterranean: Geological Society, London, Special Publications, v. 45, p. 265–283,
  doi:10.1144/GSL.SP.1989.045.01.15.
- 779 Di Bucci, D., Ravaglia, A., Seno, S., Toscani, G., Fracassi, U., and Valensise, G., 2006,
- 780 Seismotectonics of the southern Apennines and Adriatic foreland: Insights on active regional
- E-W shear zones from analogue modeling: Tectonics, v. 25, doi:10.1029/2005TC001898.
- 782 Doglioni, C., 1991, A proposal for the kinematic modelling of W-dipping subductions possible
- applications to the Tyrrhenian-Apennines system: Terra Nova, v. 3, p. 423–434,
- 784 doi:10.1111/j.1365-3121.1991.tb00172.x.
- 785 Dorsey, R.J., 1988, Provenance Evolution and Unroofing History of a Modern Arc-Continent
- 786 Collision: Evidence from Petrography of Plio-pleistocene Sandstones, Eastern Taiwan: SEPM
- Journal of Sedimentary Research, v. Vol. 58, doi:10.1306/212F8D5A-2B24-11D7-
- 788 8648000102C1865D.
- 789 Faccenna, C. et al., 2014, Mantle dynamics in the Mediterranean: Reviews of Geophysics, v. 52, p.

790 283–332, doi:10.1002/2013RG000444.

791	Gharib, F., and De Wever, P., 2010, Radiolaires mésozoïques de la formation de Kermanshah
792	(Iran): Comptes Rendus Palevol, v. 9, p. 209–219, doi:10.1016/j.crpv.2010.06.003.
793	Ghasemi, A., and Talbot, C.J., 2006, A new tectonic scenario for the Sanandaj–Sirjan Zone (Iran):
794	Journal of Asian Earth Sciences, v. 26, p. 683–693, doi:10.1016/j.jseaes.2005.01.003.
795	Glennie, K.W., and Boegner, P.L., 1981, Sole pit inversion tectonics, in Illing, L. V. and D., H.G.
796	eds., Petroleum Geology of the Continental Shelf of North-West Europe., London, p. 110-120.
797	Glennie, K.W., Boeuf, M.G.A., Clarke, M.W.H., Moody-Stuart, M., Pilaar, W.F.H., and Reinhardt,
798	B.M., 1973, Late Cretaceous nappes in Oman Mountains and their geologic evolution: AAPG
799	Bulletin, v. 57, p. 5–27.
800	Gourley, J.R., Byrne, T., Chan, YC., Wu, F., and Rau, RJ., 2007, Fault geometries illuminated
801	from seismicity in central Taiwan: Implications for crustal scale structural boundaries in the
802	northern Central Range: Tectonophysics, v. 445, p. 168–185, doi:10.1016/j.tecto.2007.08.013.
803	Granado, P., and Ruh, J.B., 2019, Numerical modelling of inversion tectonics in fold-and-thrust
804	belts: Tectonophysics, v. 763, p. 14–29, doi:10.1016/j.tecto.2019.04.033.
805	Grobe, A., Virgo, S., von Hagke, C., Urai, J.L., and Littke, R., 2018, Multiphase Structural
806	Evolution of a Continental Margin During Obduction Orogeny: Insights From the Jebel
807	Akhdar Dome, Oman Mountains: Tectonics, v. 37, p. 888–913, doi:10.1002/2016TC004442.
808	Hacker, B.R., 1994, Rapid emplacement of young oceanic lithosphere: Argon geochronology of the
809	oman ophiolite: Science, v. 265, p. 1563–1565, doi:10.1126/science.265.5178.1563.
810	Handy, M.R., M. Schmid, S., Bousquet, R., Kissling, E., and Bernoulli, D., 2010, Reconciling
811	plate-tectonic reconstructions of Alpine Tethys with the geological-geophysical record of
812	spreading and subduction in the Alps: Earth-Science Reviews, v. 102, p. 121–158,
813	doi:10.1016/j.earscirev.2010.06.002.
814	Hansman, R.J., Albert, R., Gerdes, A., and Ring, U., 2018, Absolute ages of multiple generations of
815	brittle structures by U-Pb dating of calcite: Geology, v. 46, p. 207–210.

816 Hansman, R.J., Ring, U., Thomson, S.N., den Brok, B., and Stübner, K., 2017, Late Eocene Uplift

817 of the Al Hajar Mountains, Oman, Supported by Stratigraphy and Low-Temperature

818 Thermochronology: Tectonics, v. 36, p. 3081–3109, doi:10.1002/2017TC004672.

- Hessami, K., Koyi, H.A., and Talbot, C.J., 2001, The significance of strike-slip faulting in the
  basement of the Zagros fold and thrust belt: Journal of Petroleum Geology, v. 24, p. 5–28,
  doi:10.1111/j.1747-5457.2001.tb00659.x.
- 021 doi.10.1111/j.1717.5157.2001.000059.X.
- 822 Homewood, P., Allen, P.A., and Williams, G.D., 1986, Dynamics of the Molasse Basin of Western
- Switzerland, *in* Foreland Basins, Oxford, UK, Blackwell Publishing Ltd., p. 199–217,
  doi:10.1002/9781444303810.ch10.
- 825 Homke, S., Vergés, J., Serra-Kiel, J., Bernaola, G., Sharp, I., Garcés, M., Montero-Verdú, I.,
- 826 Karpuz, R., and Goodarzi, M.H., 2009, Late Cretaceous–Paleocene formation of the proto–
- 827 Zagros foreland basin, Lurestan Province, SW Iran: GSA Bulletin, v. 121, p. 963–978,
- doi:10.1130/B26035.1.
- 829 Huang, C.-Y., Yuan, P.B., and Tsao, S.-J., 2006, Temporal and spatial records of active arc-
- continent collision in Taiwan: A synthesis: Geological Society of America Bulletin, v. 118, p.
  274–288, doi:10.1130/B25527.1.
- ISPRA, 2011, Carta geologica d'Italia, Isola di Capri (Foglio 484): Servizio Geologico d'Italia,
  scale 1:25,000, sheet 484.
- Jackson, J.A., 1980, Reactivation of basement faults and crustal shortening in orogenic belts:
  Nature, v. 283, p. 343–346, doi:10.1038/283343a0.
- 836 Jacobs, J., Thomas, R.J., Ksienzyk, A.K., and Dunkl, I., 2015, Tracking the Oman Ophiolite to the
- 837 surface New fission track and (U-Th)/He data from the Aswad and Khor Fakkan Blocks,
- 838 United Arab Emirates: Tectonophysics, v. 644, p. 68–80, doi:10.1016/j.tecto.2014.12.018.
- 839 Jamieson, R.A., and Beaumont, C., 2013, On the origin of orogens: Geological Society of America
- 840 Bulletin, v. 125, p. 1671–1702, doi:10.1130/B30855.1.
- Jammes, S., Huismans, R.S., and Muñoz, J.A., 2014, Lateral variation in structural style of
- 842 mountain building: Controls of rheological and rift inheritance: Terra Nova, v. 26, p. 201–207,
- 843 doi:10.1111/ter.12087.

- Jassim, S.Z., and Goff, J.C., 2006, Geology of Iraq (S. Z. Jassim & J. C. Goff, Eds.): Dolin, Prague
- and Moravian Museum, Brno, 341 p.
- 846 Jiménez-Munt, I., Fernàndez, M., Saura, E., Vergés, J., and Garcia-Castellanos, D., 2012, 3-D
- 847 lithospheric structure and regional/residual Bouguer anomalies in the Arabia-Eurasia collision
- 848 (Iran): Geophysical Journal International, v. 190, p. 1311–1324, doi:10.1111/j.1365-
- 849 246X.2012.05580.x.
- 850 Jolivet, L., Faccenna, C., Agard, P., Frizon de Lamotte, D., Menant, A., Sternai, P., Guillocheau, F.,
- and Polat, A., 2016, Neo-Tethys geodynamics and mantle convection: from extension to
- 852 compression in Africa and a conceptual model for obduction 1: Canadian Journal of Earth
- 853 Sciences, v. 53, p. 1190–1204, doi:10.1139/cjes-2015-0118.
- 854 Keyser, W., Tsai, C.-H., Iizuka, Y., Oberhänsli, R., and Ernst, W.G., 2016, High-pressure
- 855 metamorphism in the Chinshuichi area, Yuli belt, eastern Taiwan: Tectonophysics, v. 692, p.
- 856 191–202, doi:10.1016/j.tecto.2015.09.012.
- 857 Knott, S.D., 1987, The Liguride Complex of Southern Italy a Cretaceous to Paleogene
- 858 accretionary wedge: Tectonophysics, v. 142, p. 217–226, doi:10.1016/0040-1951(87)90124-7.
- 859 Koshnaw, R.I., Horton, B.K., Stockli, D.F., Barber, D.E., Tamar-Agha, M.Y., and Kendall, J.J.,
- 860 2017, Neogene shortening and exhumation of the Zagros fold-thrust belt and foreland basin in
- the Kurdistan region of northern Iraq: Tectonophysics, v. 694, p. 332–355,
- 862 doi:10.1016/j.tecto.2016.11.016.
- 863 Kuo-Chen, H., Wu, F.T., and Roecker, S.W., 2012, Three-dimensional P velocity structures of the
- 864 lithosphere beneath Taiwan from the analysis of TAIGER and related seismic data sets:
- Journal of Geophysical Research: Solid Earth, v. 117, doi:10.1029/2011JB009108.
- 866 Lacombe, O., and Bellahsen, N., 2016, Thick-skinned tectonics and basement-involved fold-thrust
- belts: insights from selected Cenozoic orogens: Geological Magazine, v. 153, p. 763–810,
- 868 doi:10.1017/S0016756816000078.
- 869 Lacombe, O., and Mouthereau, F., 2002, Basement-involved shortening and deep detachment
- 870 tectonics in forelands of orogens: Insights from recent collision belts (Taiwan, Western Alps,

- 871 Pyrenees): Tectonics, v. 21, p. 12-1-12–22, doi:10.1029/2001TC901018.
- 872 Lacombe, O., Mouthereau, F., Deffontaines, B., Angelier, J., Chu, H.T., and Lee, C.T., 1999,
- 873 Geometry and Quaternary kinematics of fold-and-thrust units of southwestern Taiwan:
- 874 Tectonics, v. 18, p. 1198–1223, doi:10.1029/1999TC900036.
- Lavier, L.L., and Manatschal, G., 2006, A mechanism to thin the continental lithosphere at magmapoor margins: Nature, v. 440, p. 324–328, doi:10.1038/nature04608.
- 877 Lee, Y.-H., Chen, C.-C., Liu, T.-K., Ho, H.-C., Lu, H.-Y., and Lo, W., 2006, Mountain building
- 878 mechanisms in the Southern Central Range of the Taiwan Orogenic Belt From accretionary
- wedge deformation to arc–continental collision: Earth and Planetary Science Letters, v. 252, p.
- 880 413–422, doi:10.1016/j.epsl.2006.09.047.
- 881 Lescoutre, R., and Manatschal, G., 2020, Role of rift-inheritance and segmentation for orogenic
- evolution: example from the Pyrenean-Cantabrian system: Bulletin de la Société Géologique
  de France, v. In Press.
- Lin, A.T., Watts, A.B., and Hesselbo, S.P., 2003, Cenozoic stratigraphy and subsidence history of
- the South China Sea margin in the Taiwan region: Basin Research, v. 15, p. 453–478,
- 886 doi:10.1046/j.1365-2117.2003.00215.x.
- Lister, G.S., Etheridge, M.A., and Symonds, P.A., 1986, Detachment faulting and the evolution of
- passive continental margins: Geology, v. 14, p. 246, doi:10.1130/0091-
- 889 7613(1986)14<246:DFATEO>2.0.CO;2.
- 890 Macedo, J., and Marshak, S., 1999, Controls on the geometry of fold-thrust belt salients: Geological
- 891 Society of America Bulletin, v. 111, p. 1808–1822, doi:10.1130/0016-
- 892 7606(1999)111<1808:COTGOF>2.3.CO;2.
- 893 MacKay, M.E., 1995, Structural variation and landward vergence at the toe of the Oregon
- 894 accretionary prism: Tectonics, v. 14, p. 1309–1320, doi:10.1029/95TC02320.
- 895 Marshak, S., Karlstom, K., and Timmons, J.M., 2000, Inversion of proterozoic extensional faults:
- 896 An explanation for the pattern of Laramide and Ancestral rockies intracratonic deformation,
- 897 United States: Geology, v. 28, p. 735–738, doi:10.1130/0091-

- 898 7613(2000)28<735:IOPEFA>2.0.CO;2.
- Mazzoli, S., Ascione, A., Buscher, J.T., Pignalosa, A., Valente, E., and Zattin, M., 2014, Low-angle
  normal faulting and focused exhumation associated with late Pliocene change in tectonic style
- 901 in the southern Apennines (Italy): Tectonics, v. 33, p. 1802–1818,
- 902 doi:10.1002/2014TC003608.
- 903 Mazzoli, S., Barkham, S., Cello, G., Gambini, R., Mattioni, L., Shiner, P., and Tondi, E., 2001,
- 904 Reconstruction of continental margin architecture deformed by the contraction of the
- Lagonegro Basin, southern Apennines, Italy: Journal of the Geological Society, v. 158, p. 309–
  319.
- 907 McIntosh, K., van Avendonk, H., Lavier, L., Lester, W.R., Eakin, D., Wu, F., Liu, C.-S., and Lee,
- 908 C.-S., 2013, Inversion of a hyper-extended rifted margin in the southern Central Range of
- 909 Taiwan: Geology, v. 41, p. 871–874, doi:10.1130/G34402.1.
- 910 McIntosh, K., Nakamura, Y., Wang, T.-K., Shih, R.-C., Chen, A., and Liu, C.-S., 2005, Crustal-
- scale seismic profiles across Taiwan and the western Philippine Sea: Tectonophysics, v. 401,
- 912 p. 23–54, doi:10.1016/j.tecto.2005.02.015.
- 913 Menardi Noguera, A., and Rea, G., 2000, Deep structure of the Campanian–Lucanian Arc (Southern
- 914 Apennine, Italy): Tectonophysics, v. 324, p. 239–265, doi:10.1016/S0040-1951(00)00137-2.
- 915 Mohn, G., Manatschal, G., Beltrando, M., Masini, E., and Kusznir, N., 2012, Necking of
- 916 continental crust in magma-poor rifted margins: Evidence from the fossil Alpine Tethys
  917 margins: Tectonics, v. 31, doi:10.1029/2011TC002961.
- 918 Molli, G., and Malavieille, J., 2011, Orogenic processes and the Corsica/Apennines geodynamic
- 919 evolution: insights from Taiwan: International Journal of Earth Sciences, v. 100, p. 1207–
- 920 1224, doi:10.1007/s00531-010-0598-y.
- 921 Moore, G.F., Shipley, T.H., Stoffa, P.L., Karig, D.E., Taira, A., Kuramoto, S., Tokuyama, H., and
- 922 Suyehiro, K., 1990, Structure of the Nankai Trough Accretionary Zone from multichannel
- 923 seismic reflection data: Journal of Geophysical Research, v. 95, p. 8753,
- 924 doi:10.1029/JB095iB06p08753.

- 925 Mount, V.S., Crawford, R.I.S., and Bergman, S.C., 1998, Regional structural style of the Central
- 926 and Southern Oman Mountains: Jebel Akhdar, Saih Hatat, and the Northern Ghaba Basin:

927 GeoArabia, v. 3, p. 475–490.

- 928 Mouthereau, F., Deffontaines, B., Lacombe, O., and Angelier, J., 2002, Variations along the strike
- 929 of the Taiwan thrust belt: Basement control on structural style, wedge geometry, and
- kinematics, *in* Byrne, T.B. and Liu, C.-S. eds., Geology and Geophysics of an Arc-continent
- 931 Collision, Taiwan, Republic of China, Colorado, Boulder, p. 35–58.
- 932 Mouthereau, F., and Lacombe, O., 2006, Inversion of the Paleogene Chinese continental margin and
- 933 thick-skinned deformation in the Western Foreland of Taiwan: Journal of Structural Geology,
- 934 v. 28, p. 1977–1993, doi:10.1016/j.jsg.2006.08.007.
- 935 Mouthereau, F., Lacombe, O., and Vergés, J., 2012, Building the Zagros collisional orogen:
- 936 Timing, strain distribution and the dynamics of Arabia/Eurasia plate convergence:
- 937 Tectonophysics, v. 532–535, p. 27–60, doi:10.1016/j.tecto.2012.01.022.
- 938 Muñoz, J.A., Beamud, E., Fernández, O., Arbués, P., Dinarès-Turell, J., and Poblet, J., 2013, The
- Ainsa Fold and thrust oblique zone of the central Pyrenees: Kinematics of a curved
- 940 contractional system from paleomagnetic and structural data: Tectonics, v. 32, p. 1–34,
- 941 doi:10.1002/tect.20070.
- 942 Nemcok, M., Gayer, R., and Miliorizos, M., 1995, Structural analysis of the inverted Bristol
- 943 Channel Basin: implications for the geometry and timing of fracture porosity: Geological
- 944 Society, London, Special Publications, v. 88, p. 355–392,
- 945 doi:10.1144/GSL.SP.1995.088.01.20.
- 946 Nemčok, M., Mora, A., and Cosgrove, J., 2013, Thick-skin-dominated orogens; from initial
- 947 inversion to full accretion: An introduction: Geological Society Special Publication, v. 377, p.
  948 1–17, doi:10.1144/SP377.17.
- Ogniben, L., 1969, Schema introduttivo alla geologia del confine calabro-lucano: Mem. Soc. Geol.
  It., v. 8, p. 453–763.
- 951 Oncken, O., von Winterfeld, C., and Dittmar, U., 1999, Accretion of a rifted passive margin: The

- 952 Late Paleozoic Rhenohercynian fold and thrust belt (Middle European Variscides): Tectonics,
- 953 v. 18, p. 75–91, doi:10.1029/98TC02763.
- Patacca, E., and Scandone, P., 2007, Geology of the southern Apennines: Bollettino della Società
  Geologica Italiana, v. 7, p. 75–119.
- 956 Péron-Pinvidic, G., and Manatschal, G., 2009, The final rifting evolution at deep magma-poor
- passive margins from Iberia-Newfoundland: a new point of view: International Journal of
- 958 Earth Sciences, v. 98, p. 1581–1597, doi:10.1007/s00531-008-0337-9.
- 959 Péron-Pinvidic, G., & Manatschal, G., 2010, From microcontinents to extensional allochthons:
  960 witnesses of how continents rift and break apart?. Petroleum Geoscience, v. 16, p. 189-197.
- 961 doi:10.1144/1354-079309-903
- 962 Pfiffner, O.A., 2017, Thick-Skinned and Thin-Skinned Tectonics: A Global Perspective:
- 963 Geosciences, v. 7, p. 71, doi:10.3390/geosciences7030071.
- Rabu, D. et al., 1993, Stratigraphy and structure of the Oman mountains: Documents BRGM, 300
  p.
- Rioux, M., Bowring, S., Kelemen, P., Gordon, S., Dudás, F., and Miller, R., 2012, Rapid crustal
- 967 accretion and magma assimilation in the Oman-U.A.E. ophiolite: High precision U-Pb zircon
- geochronology of the gabbroic crust: Journal of Geophysical Research: Solid Earth, v. 117, p.
- 969 1–12, doi:10.1029/2012JB009273.
- 970 Robertson, A.H.F., and Searle, M.P., 1990, The northern Oman Tethyan continental margin:
- 971 stratigraphy, structure, concepts and controversies: Geological Society, London, Special

972 Publications, v. 49, p. 3–25, doi:10.1144/GSL.SP.1992.049.01.02.

- 873 Roure, F., Casero, P., and Vially, R., 1991, Growth processes and melange formation in the
- southern Apennines accretionary wedge: Earth and Planetary Science Letters, v. 102, p. 395–
  412.
- 976 Saddiqi, O., Michard, A., Goffe, B., Poupeau, G., and Oberhänsli, R., 2006, Fission-track
- 977 thermochronology of the Oman Mountains continental windows, and current problems of
- 978 tectonic interpretation: Bulletin de la Societe Geologique de France, v. 177, p. 127–143,

979 doi:10.2113/gssgfbull.177.3.127.

- 980 Santantonio, M., and Carminati, E., 2011, Jurassic rifting evolution of the Apennines and Southern
- Alps (Italy): Parallels and differences: Geological Society of America Bulletin, v. 123, p. 468–
  484, doi:10.1130/B30104.1.
- 983 Saura, E., Garcia-Castellanos, D., Casciello, E., Parravano, V., Urruela, A., and Vergés, J., 2015,
- Modeling the flexural evolution of the Amiran and Mesopotamian foreland basins of NW
  Zagros (Iran-Iraq): Tectonics, v. 34, p. 377–395, doi:10.1002/2014TC003660.
- 986 Schmid, S.M., Fügenschuh, B., Kissling, E., and Schuster, R., 2004, Tectonic map and overall
- 987 architecture of the Alpine orogen: Eclogae Geologicae Helvetiae, v. 97, p. 93–117,

988 doi:10.1007/s00015-004-1113-x.

- 989 Schmid, S.M., and Kissling, E., 2000, The arc of the western Alps in the light of geophysical data
- 990 on deep crustal structure: Tectonics, v. 19, p. 62–85, doi:10.1029/1999TC900057.
- Scrocca, D., Carminati, E., and Doglioni, C., 2005. Deep structure of the southern Apennines, Italy:
  Thin-skinned or thick-skinned?. Tectonics, v. 24, TC3005, doi:10.1029/2004TC001634
- Scrocca, D., 2010, Southern Apennines: structural setting and tectonic evolution: Journal of the
  Virtual Explorer, v. 36, doi:10.3809/jvirtex.2010.00225.
- 995 Searle, M.P., and Cox, J., 2002, Subduction zone metamorphism during formation and
- 996 emplacement of the Semail ophiolite in the Oman Mountains: Geological Magazine, v. 139, p.
- 997 241–255, doi:10.1017/S0016756802006532.
- 998 Shyu, J.B.H., Sieh, K., Chen, Y.-G., Chuang, R.Y., Wang, Y., and Chung, L.-H., 2008,
- 999 Geomorphology of the southernmost Longitudinal Valley fault: Implications for evolution of
- 1000 the active suture of eastern Taiwan: Tectonics, v. 27, doi:10.1029/2006TC002060.
- Sibuet, J.-C., and Hsu, S.-K., 2004, How was Taiwan created? Tectonophysics, v. 379, p. 159–181,
  doi:10.1016/j.tecto.2003.10.022.
- 1003 Simoes, M., Avouac, J.P., Beyssac, O., Goffé, B., Farley, K.A., and Chen, Y.-G., 2007, Mountain
- 1004 building in Taiwan: A thermokinematic model: Journal of Geophysical Research, v. 112, p.
- 1005 B11405, doi:10.1029/2006JB004824.

- 1006 Simoes, M., Beyssac, O., and Chen, Y.-G., 2012, Late Cenozoic metamorphism and mountain
- 1007 building in Taiwan: A review: Journal of Asian Earth Sciences, v. 46, p. 92–119,

1008 doi:10.1016/j.jseaes.2011.11.009.

- 1009 Somma, R., Messina, A., and Mazzoli, S., 2005, Syn-orogenic extension in the Peloritani Alpine
- 1010 Thrust Belt (NE Sicily, Italy): Evidence from the Alì Unit: Comptes Rendus Geoscience, v.
- 1011 337, p. 861–871, doi:10.1016/j.crte.2005.03.004.
- 1012 Sutra, E., Manatschal, G., Mohn, G., and Unternehr, P., 2013, Quantification and restoration of
- 1013 extensional deformation along the Western Iberia and Newfoundland rifted margins:
- 1014 Geochemistry, Geophysics, Geosystems, v. 14, p. 2575–2597, doi:10.1002/ggge.20135.
- 1015 Talebian, M., and Jackson, J., 2004, A reappraisal of earthquake focal mechanisms and active
- 1016 shortening in the Zagros mountains of Iran: Geophysical Journal International, v. 156, p. 506–
- 1017 526, doi:10.1111/j.1365-246X.2004.02092.x.
- 1018 Tang, C.-C., Zhu, L., Chen, C.-H., and Teng, T.-L., 2011, Significant crustal structural variation
- 1019 across the Chaochou Fault, southern Taiwan: New tectonic implications for convergent plate
- 1020 boundary: Journal of Asian Earth Sciences, v. 41, p. 564–570,
- 1021 doi:10.1016/j.jseaes.2010.12.003.
- 1022 Tarapoanca, M., Andriessen, P., Broto, K., Chérel, L., Ellouz-Zimmermann, N., Faure, J.-L., Jardin,
- 1023 A., Naville, C., and Roure, F., 2010, Forward kinematic modelling of a regional transect in the
- 1024 Northern Emirates using geological and apatite fission track age constraints on paleo-burial
- 1025 history: Arabian Journal of Geosciences, v. 3, p. 395–411, doi:10.1007/s12517-010-0213-3.
- 1026 Tavani, S., Camanni, G., Nappo, M., Snidero, M., Ascione, A., Valente, E., Gharabeigli, G.,
- 1027 Morsalnejad, D., and Mazzoli, S., 2020, The Mountain Front Flexure in the Lurestan region of
- 1028 the Zagros belt: Crustal architecture and role of structural inheritances: Journal of Structural
- 1029 Geology, v. 135, p. 104022, doi:10.1016/j.jsg.2020.104022.
- 1030 Tavani, S., Parente, M., Puzone, F., Corradetti, A., Gharabeigli, G., Valinejad, M., Morsalnejad, D.,
- and Mazzoli, S., 2018a, The seismogenic fault system of the 2017 M w 7.3 Iran–Iraq
- 1032 earthquake: constraints from surface and subsurface data, cross-section balancing, and

- 1033 restoration: Solid Earth, v. 9, p. 821–831, doi:10.5194/se-9-821-2018.
- 1034 Tavani, S., Parente, M., Vitale, S., Iannace, A., Corradetti, A., Bottini, C., Morsalnejad, D., and
- 1035 Mazzoli, S., 2018b, Early Jurassic Rifting of the Arabian Passive Continental Margin of the
- 1036 Neo-Tethys. Field Evidence From the Lurestan Region of the Zagros Fold-and-Thrust Belt,
- 1037 Iran: Tectonics, v. 37, p. 2586–2607, doi:10.1029/2018TC005192.
- 1038 Teng, L.S., 1990, Geotectonic evolution of late Cenozoic arc-continent collision in Taiwan:
- 1039 Tectonophysics, v. 183, p. 57–76, doi:10.1016/0040-1951(90)90188-E.
- 1040 Teng, L.S., and Lin, A.T., 2004, Cenozoic tectonics of the China continental margin: insights from
- 1041 Taiwan, in Malpas, J., Fletcher, C.J.N., Ali, J.R., and Aitchison, J.C. eds., Aspects of the
- 1042 Tectonic Evolution of China, Geological Society, London, Special Publications, 226, v. 226, p.
- 1043 313–332, doi:10.1144/GSL.SP.2004.226.01.17.
- 1044 Vergés, J., Saura, E., Casciello, E., Fernàndez, M., Villaseñor, A., Jiménez-Munt, I., and García-
- 1045 Castellanos, D., 2011, Crustal-scale cross-sections across the NW Zagros belt: implications for 1046 the Arabian margin reconstruction: Geological Magazine, v. 148, p. 739–761,
- 1047 doi:10.1017/S0016756811000331.
- 1048 Warburton, J., Burnhill, T.J., Graham, R.H., and Isaac, K.P., 1990, The evolution of the Oman
- 1049 Mountains Foreland Basin, in Robertson, A.F.H., Searle, M.P., and Ries, A.C. eds., The
- 1050 Geology and Tectonics of the Oman Region. Geological Society of London, Special
- 1051 Publications, v. 49, p. 419–427, doi:10.1144/GSL.SP.1992.049.01.26.
- 1052 Warren, C.J., Parrish, R.R., Searle, M.P., and Waters, D.J., 2003, Dating the subduction of the
- Arabian continental margin beneath the Semail ophiolite, Oman: Geology, v. 31, p. 889–892,
  doi:10.1130/G19666.1.
- 1055 Whitmarsh, R.B., Manatschal, G., and Minshull, T.A., 2001, Evolution of magma-poor continental
- 1056 margins from rifting to seafloor spreading: Nature, v. 413, p. 150–154, doi:10.1038/35093085.
- 1057 Williams, G.D., Powell, C.M., and Cooper, M.A., 1989, Geometry and kinematics of inversion
- 1058 tectonics: Geological Society, London, Special Publications, v. 44, p. 3–15,
- 1059 doi:10.1144/GSL.SP.1989.044.01.02.

- 1060 Wilson, J.T., 1966, Did the Atlantic Close and then Re-Open? Nature, v. 211, p. 676–681,
- 1061 doi:10.1038/211676a0.
- 1062 Wrobel-Daveau, J.-C., Ringenbach, J.-C., Tavakoli, S., Ruiz, G.M.H., Masse, P., and Frizon de
- 1063 Lamotte, D., 2010, Evidence for mantle exhumation along the Arabian margin in the Zagros
- 1064 (Kermanshah area, Iran): Arabian Journal of Geosciences, v. 3, p. 499–513,
- 1065 doi:10.1007/s12517-010-0209-z.
- 1066 Wu, Y.-M., Chang, C.-H., Zhao, L., Shyu, J.B.H., Chen, Y.-G., Sieh, K., and Avouac, J.-P., 2007,
- 1067 Seismic tomography of Taiwan: Improved constraints from a dense network of strong motion
- stations: Journal of Geophysical Research, v. 112, p. B08312, doi:10.1029/2007JB004983.
- 1069 Yue, L.-F., Suppe, J., and Hung, J.-H., 2005, Structural geology of a classic thrust belt earthquake:
- 1070 the 1999 Chi-Chi earthquake Taiwan (Mw=7.6): Journal of Structural Geology, v. 27, p. 2058–
- 1071 2083, doi:10.1016/j.jsg.2005.05.020.
- Zanchi, A., Berra, F., Mattei, M., R. Ghassemi, M., and Sabouri, J., 2006, Inversion tectonics in
  central Alborz, Iran: Journal of Structural Geology, v. 28, p. 2023–2037,
- 1074 doi:10.1016/j.jsg.2006.06.020.



Figure 1 2 columns in 3 column layout



Figure 2 3 columns in 3 column layout



Figure 3 2 columns in 3 column layout



Figure 4 3 columns in 3 column layout







Figure 6 2 columns in 3 column layout



