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**The syn-rift tectono-stratigraphic record of rifted margins (Part II):  
A new model to break through the proximal/distal interpretation frontier**

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Keywords: hyperextended margin, rifting, tectono-stratigraphic record

Highlights

- A new tectono-stratigraphic model to help break through the proximal/distal interpretation frontier
- Each rift domain is associated with specific structures and sedimentary architectures
- Each rift domain displays a specific ratio of vertical vs horizontal accommodation space creation
- Rift deformation localizes sequentially, inducing a younging of the base of passive infill oceanward
- The early syn-rift stratigraphic record is largely discontinuous/broken up in the distal margin

## 24 Abstract

25 One fundamental lesson from the last decades research into rifted margins is that  
26 stratigraphic horizons cannot be simply correlated from the proximal into distal domain, which  
27 makes the interpretation of poorly calibrated and largely inaccessible distal margins difficult.  
28 In this contribution, we propose a new tectono-stratigraphic model to help break through the  
29 proximal/distal interpretation frontier. After reviewing the scientific advances on rifted margins  
30 achieved during the last century, we describe the primary spatio-temporal evolution of active  
31 deformation during rifting. We show that different rift domains are associated with specific  
32 (1) periods of active deformation; (2) tectonic structures and related stratigraphic  
33 architecture; (3) amounts of accommodation creation owing to specific ranges of horizontal  
34 widening and vertical deepening; and (4) depositional environments. We demonstrate that  
35 the sequential localization of extension during rifting results in a younging of the base of  
36 passive infill (i.e., sediment deposited in a non-tectonic setting) oceanward. We propose a  
37 panel of few to few tens of My time intervals related to specific tectonic events – our *Tectonic*  
38 *Sequences* – that may be correlated more or less easily and continuously across rifted  
39 margins. This study underlines the value of an iterative approach between fieldwork on fossil  
40 rifted margins and offshore geophysical studies, where field studies offer easy-to-access and  
41 high-resolution calibration of dismembered rifted margin remnants, while geophysical studies  
42 provide comparatively low-resolution imaging of entire, intact but largely inaccessible, rifted  
43 margins. It also highlights the necessity of jettisoning outdated dogma on rifted margins and  
44 changing our routines when interpreting seismic sections and outcrops.

45

## 46 1. Introduction

47 Rifting is a process involving extension of the crust and upper mantle, during which  
48 accommodation is created by both horizontal widening and vertical deepening. The  
49 involved sedimentary systems record the timing, location and type of tectonic

50 deformation, the distribution of depocenters, the reorganization of sedimentary sources,  
51 and the evolution of water depths. Yet, distal parts of present-day rifted margins have  
52 remained largely inaccessible to date, including to drilling, which makes the interpretation  
53 of their syn-rift stratigraphic architecture, the determination of their sedimentary facies,  
54 and their correlation with more proximal parts of the margin difficult.

55 Over the last decades, the scientific community has made significant progress in  
56 understanding the processes at play during rifting. The iterative workflow between  
57 onshore and offshore studies, together with the improvement of imaging, dating and  
58 modelling techniques, enabled us to identify the characteristics common to rift systems,  
59 and to recognize a systematic deformation sequence during their formation. The ensuing  
60 question is: does a comparable formation history among rift systems imply a similar  
61 tectono-stratigraphic record, despite the differences in their local sedimentary sources  
62 and depositional environments?

63 In this contribution, we propose a new conceptual tectono-stratigraphic model  
64 synthesizing the evolution of deformation over space and time during rifting. We review  
65 *when, where* and *how* accommodation is created, and show that specific domains of  
66 rifted margins are associated with specific (1) periods of tectonic activity; (2) style of  
67 deformation and tectono-stratigraphic architectures; (3) ratios of horizontal vs vertical  
68 accommodation creation; and (4) depositional environments. After briefly summarizing  
69 the historical evolution of models for rifted margins, we describe our tectono-stratigraphic  
70 model and compare it with observations from both present-day and fossil rifted margins.  
71 Then, we discuss how this template can help interpreting seismic sections in poorly  
72 calibrated rift systems and to correlate remnants of dismembered margins embedded in  
73 orogens (e.g., Alps, Pyrenees, Oman, Timor, Zagros). Finally, we examine the limitations  
74 of our model, in particular the correlation of timelines/time intervals both along dip and  
75 strike of rifted margins, the effect of sedimentation rates with respect to the rates of

76 subsidence, the impact of magma, and the challenge of including a three-dimensional  
77 perspective at segmented and multiphase rift systems.

78

## 79 2. Historical overview

80 Rifting has always been regarded as a process creating accommodation space that is  
81 filled with sediments and/or volcanic products and/or water, both during and after the  
82 termination of deformation (McKenzie, 1978). However, our understanding of *when*,  
83 *where* and *how* accommodation is created and filled has been significantly evolving over  
84 the last decades (e.g., Bond and Kominz, 1988; De Graciansky *et al.*, 2011; Gawthorpe  
85 *et al.*, 1994; Gawthorpe and Leeder, 2000; Olsen and Morgan, 2006; Schlische, 1995;  
86 Withjack *et al.*, 2002).

87

### 88 2.1. The Geosyncline Theory (from the late 1850's to the 1960's)

89 The premise of tectonics emerged from the work of Hall (1859), who first recognized  
90 the existence of vertical movements (subsidence and uplift) of the Earth surface and  
91 the folding of sedimentary series (Figure 1a). Hall suggested that folding occurred at  
92 the *axis* of *synclinal* structures that resulted from Earth surface deformation. Dana  
93 (1873) suggested that the subsidence of *Geosynclines* (which corresponded to Hall's  
94 synclinal axis) was balanced by the rising of adjacent *Geanticlines*, both resulting  
95 from Earth contraction-driven lateral compression. A few decades later, Haug (1900)  
96 introduced the distinction between shallow- and deep-water facies, and emphasized  
97 their lateral change between the margins and the axial part of a Geosyncline. He also  
98 suggested that Geosynclines were mobile zones located at the boundary between  
99 two stable continental masses. These basic concepts were further developed during  
100 subsequent decades with the work of Argand (1916), Kay (1951), Kober (1923) and

101           Stille (1913, 1924, 1936), among others. For further details on the birth and evolution  
102           of the Geosyncline Theory see the review by Aubouin (1965).

103

104           2.2. The early stages of Plate Tectonic Theory: Pure-Shear vs Simple-Shear models and  
105           the birth of seismic stratigraphy (from the 1970's to the early 1990's)

106           Based on the first reflection seismic sections at rifted margins (e.g., Montadert *et al.*  
107           1979), rift systems and rifted margins were envisioned as essentially comprising  
108           oceanward-dipping, high-angle (ca. 60 °) normal faults separating a series of tilted  
109           fault blocks and associated half-graben/graben whose size decreased oceanward  
110           (Figure 1b). The oceanic crust was supposed to start directly where thinned  
111           continental crust terminated, which implied a localized and instantaneous breakup  
112           related to the onset of steady-state magma emplacement. McKenzie (1978) and  
113           Jarvis and McKenzie (1980) envisioned rifting as Pure-Shear (depth-uniform) thinning  
114           of the lithosphere, resulting in both rapid (“instantaneous”) subsidence associated  
115           with crustal thinning, and subsequent protracted, progressively decreasing  
116           subsidence related to thermal relaxation of the thinned lithospheric mantle.

117           A few years later, Lemoine *et al.* (1987) suggested that Wernicke's (1985) alternative  
118           Simple-Shear Model would better explain the ancient Alpine Tethys rift system  
119           (Figure 1c). Indeed, observations from the Alpine Tethys margins exposed in the Alps  
120           diverged from the Pure-Shear Model predictions in that: (1) the topographic and  
121           thermal evolution of the former conjugate margins appeared to be largely asymmetric  
122           (Lemoine, 1985), contrasting with the predictions of the symmetrical Pure-Shear  
123           Model; (2) at least one of the two margins was uplifted during rifting (e.g., the  
124           Briançonnais Block; Lemoine *et al.*, 1986), while the Pure-Shear model only predicted  
125           subsidence; and (3) no evidence for a Penrose-type oceanic crust could be found;  
126           instead, Alpine geologists had evidenced exhumation of subcontinental mantle to the  
127           seafloor (Decandia and Elter, 1972, 1969).

128 Contemporaneously, Falvey (1974) laid the ground for modern rifted margin  
129 stratigraphy. He stated that the onset of rifting was marked by an unconformity  
130 generated by a phase of thermal doming, which he named the *Rift Onset*  
131 *Unconformity (ROU)*. He also mentioned the existence of a so-called *Breakup*  
132 *Unconformity (BU)*, associated with the onset of rapid subsidence of the entire  
133 margin. This phase of rapid subsidence, modelled a few years later as an isostatic  
134 adjustment following the mechanical breakup of the lithosphere by Braun and  
135 Beaumont (1989), was thought to be expressed by the progradation of sedimentary  
136 deposits over the shelf and continental slope.

137 Hubbard (1988) introduced the term “megasequence”, which he defined as a “major  
138 sedimentary package” bounded by “the most prominent regional unconformities”. He  
139 suggested that these prominent regional unconformities “record the principal periods  
140 of change in basin geometry and polarity”. Hubbard (1988), and then Prosser (1993),  
141 recognized three types of megasequences, namely pre-, syn- and post-rift  
142 megasequences, deposited respectively prior to the onset of rifting, during rifting and  
143 after the termination of rifting.

144 In the 1980’s–early 1990’s, the architecture of the pre-, syn and post-rift  
145 megasequences were conceptualized with the help of analytical (Barr, 1991, 1987),  
146 analogue (Buchanan and McClay, 1991), and numerical modelling (Hardy, 1993).  
147 The pre-rift megasequence was supposed to be coupled to the underlying basement,  
148 except in the presence of low frictional, efficiently decoupling materials like salt or  
149 clays. In the absence of pre-rift tectonic events, pre-rift deposits are essentially  
150 comprised of parallel strata (i.e., without growth structures; violet sediments in  
151 Figures 1b and 1c).

152 The syn-rift megasequence was regarded as comprised of wedge-shape sedimentary  
153 deposits that thicken toward the adjacent normal fault (blue sediments in Figures 1b  
154 and 1c). However, the nature of the isostatic compensation (none, local or regional),

155 the geometry of the normal fault involved (planar or listric), sedimentation rates and  
156 climate may result in various stratal geometries of these syn-rift deposits (e.g., Barr,  
157 1991; Buchanan and McClay, 1991; Prosser, 1993).

158 The post-rift megasequence was interpreted to be made of un-extended sedimentary  
159 packages that drape or onlap against preexisting topography, or downlap onto  
160 oceanic crust (green sediments in Figures 1b and 1c). In isolated basins, when the  
161 sedimentation rate is very low compared to the rate of subsidence, these deposits are  
162 parallel and horizontal. Otherwise, they display clinofold shapes (e.g., significant  
163 thickening of strata toward the basin center; Fig. 5 of Waltham *et al.*, 1993).

164 In the “rigid domino fault blocks” models for rifted margins (Figures 1b and 1c), all  
165 faults and intervening tilted blocks are required to move and stop moving  
166 simultaneously, and thus the syn-rift megasequence is synchronous across the entire  
167 margin (blue sediments in Figures 1b and 1c; Waltham *et al.*, 1993). Both the base  
168 and top of the syn-rift megasequence are therefore ideally marked by an isochronous  
169 change in stratal dip across the entire margin, either between parallel pre-rift deposits  
170 and wedge-shape syn-rift deposits in the former case, or between wedge-shape syn-  
171 rift deposits and parallel post-rift deposits in the latter.

172

173

### 174 2.3. Toward a Polyphase Rifting Model (from the 1990's to the early 2010's)

175 Stratigraphic studies from both the Alps (Froitzheim and Eberli, 1990) and the Iberia–  
176 Newfoundland margins (Murillas *et al.*, 1990; Tucholke *et al.*, 2007; Wilson *et al.*,  
177 2001) highlighted that Alpine Tethys and southern North-Atlantic rifting were achieved  
178 via two main, distinct and successive extensional phases. During the first phase,  
179 numerous and broadly distributed half-graben depocenters, bounded by high-angle  
180 normal faults, formed over a wide region (dark blue sediments in Figure 1d). At this



181 stage, crustal and lithospheric thinning were minor ( $1/\beta \ll 1.5$ , where  $1/\beta$   
182 corresponds to the ratio *crust thickness prior to extension / crust thickness at the end*  
183 *of extension*; McKenzie, 1978). During the second phase, tectonic activity was  
184 localized only in the future distal margins. There, significant crustal thinning  
185 ( $1.5 < 1/\beta < \infty$ ) was achieved via both normal and extensional detachment faulting, as  
186 well as via the attenuation of ductile middle- and/or lower crustal layers (Manatschal  
187 *et al.* 2001; light blue sediments in Figure 1d). Crustal thinning was accompanied by  
188 significant lithospheric thinning.

189 Using numerical models, Lavier and Manatschal (2006) suggested that rifting was  
190 achieved via successive and distinct deformation modes, which we call here  
191 *Polyphase Rifting*. In their model, deformation evolved from initially distributed  
192 extension with limited crustal and lithospheric thinning (the so-called *stretching*  
193 *phase*) to increasingly localized deformation with significant crustal and lithospheric  
194 thinning (the so-called *thinning* and (*mantle*) *exhumation* phases). At the shift  
195 between the *stretching* and *thinning* phases, deformation stopped in the future  
196 proximal domain while rifting was still active in the future distal domain. This  
197 numerical model was consistent with the observation of passive infill sedimentation in  
198 the proximal domain at a time when rifting was not achieved yet, as in the Alpine  
199 Tethys (Froitzheim and Eberli, 1990; Masini *et al.*, 2013) and in Iberia–Newfoundland  
200 (Manatschal and Bernoulli, 1999; Murillas *et al.*, 1990; Péron-Pinvidic *et al.*, 2007;  
201 Tucholke *et al.*, 2007).

202 The observation of passive infill deposits emplaced during rifting required the  
203 refinement of the term *syn-rift*. Following Masini *et al.* (2013), we here restrict the use  
204 of *syn-rift* to its temporal dimension with respect to the overall rifting event, and use  
205 the terms *pre-tectonic*, *syn-tectonic* and *post-tectonic* to characterize the architecture  
206 of sedimentary deposits with respect to local tectonic activity (Figures 2a–2d). In this  
207 perspective, pre-rift sediments are necessarily pre-tectonic and post-rift sediments

208 are necessarily post-tectonic. However, syn-rift sediments may be either pre-, syn- or  
209 post-tectonic (Figure 2e).

210 In the absence of pre-rift tectonic events and inherited topography, pre-tectonic  
211 sediments display horizontal and parallel strata of constant thickness (Figure 2a).  
212 Syn-tectonic sediments are usually wedge-shaped, but their architecture varies  
213 depending on the type of fault they are associated with (Figure 2f). Indeed, while  
214 normal faults generate only minor horizontal extension (< 5 km), extensional  
215 detachment faults may exhume basement over more than 15 km (e.g., Epin and  
216 Manatschal, 2018). We qualify deposits as *half-graben wedge-shaped* when  
217 associated with a high-angle normal fault (Figure 2b) and *supra-detachment wedge-*  
218 *shaped* when related to an extensional detachment fault (Figure 2c). Note that the  
219 development of such growth structures is conditional on a sufficient syn-tectonic  
220 sedimentation rate. Extensional detachment faults are generally associated with  
221 cataclastic breccias and black gouges (Chester and Logan, 1986; Froitzheim and  
222 Eberli, 1990; Manatschal *et al.*, 2000), and, when exhumed to the seafloor, their  
223 surface is typically overlain by tectono-sedimentary breccias (e.g., Collettini, 2011,  
224 and references therein; Masini *et al.*, 2011). Post-tectonic deposits (Figure 2d) are  
225 either made of strata onlapping against, downlapping onto, or draping residual rift-  
226 related topography. Away from rift-related faults and/or in the absence of rift-related  
227 topography or intervening syn-tectonic deposits, pre- and post-tectonic deposits may  
228 practically be indistinguishable and only a paraconformity might capture the  
229 intervening syn-tectonic stage.

230

### 231 3. Building a stratigraphic model for idealized rifted margins

232 Over the last decade, the intense iterative workflow between field studies at remnants  
233 of rifted margins exposed on land (e.g., Andersen *et al.*, 2012; Masini *et al.*, 2013;

234 Mohn *et al.*, 2010; Epin *et al.*, 2017; Gawthorpe *et al.*, 2018; Ribes *et al.*, 2019b;  
235 Serck *et al.*, 2021; Stockli and Bosworth, 2018), interpretation of rifted margin seismic  
236 sections with ever higher resolutions (e.g., Causer *et al.*, 2020; Clerc *et al.*, 2018;  
237 Gillard *et al.*, 2016; Hauptert *et al.*, 2016; Lymer *et al.*, 2019; Osmundsen *et al.*, 2016;  
238 Whiting *et al.*, 2020; Yang and Welford, 2021; Zhang *et al.*, 2019) and numerical  
239 modelling (e.g., Huismans and Beaumont, 2011; Brune *et al.*, 2014; Chenin *et al.*,  
240 2020; Duretz *et al.*, 2020; Jourdon *et al.*, 2020; Naliboff *et al.*, 2017; Pérez-Gussinyé  
241 *et al.*, 2020; Petri *et al.*, 2019) have allowed to considerably improve our  
242 understanding of the architecture and evolution of rifted margins.

### 243 3.1. Benefits and pitfalls of using a template to guide interpretation

244 In this contribution, we propose a tectono-stratigraphic model to guide seismic  
245 and field interpretations in correlating observations from proximal margins into  
246 their distal continuation. Although model-driven approaches have historically  
247 caused significant misinterpretations and remain problematic when blindly  
248 applied (e.g., review by Manatschal *et al.*, 2021), having a first-order template  
249 may help when attempting to unravel the tectono-stratigraphic-record of  
250 uncalibrated distal domains. Indeed, sedimentary facies of given age may vary  
251 between the shallow-water proximal domain and deep-water hyperextended  
252 domain, and thus one cannot rely on the seismic signature/sedimentary field  
253 observations to correlate sedimentary units. Furthermore, as new real estate  
254 (i.e., new depositional surface) is created via the exhumation of basement along  
255 extensional detachment faults, the stratigraphic record may be largely  
256 discontinuous (e.g., sediments older than the age of exhumation are missing  
257 over the corresponding new real estate). This implies that seismic/field  
258 interpretations should not attempt to find continuous pre-rift or early syn-rift  
259 deposits in distal margins, where extensional detachment faults are likely.

260 The tectono-stratigraphic model we propose aims to describe an interpretation  
261 philosophy rather than providing a template to be copy-pasted, and we stress  
262 that careful seismic observations, rigorous application of the first principles of  
263 stratigraphy, and mindful consideration of the local geological history remain  
264 crucial to reliable interpretations.

265

## 266 3.2. Preliminary concepts and definitions

267

### 268 3.2.1. A spatial framework for rifted margins: definition of rift domains

269 Based on the geometry of top basement and Moho, which are two of the most  
270 consensual interfaces observed on seismic sections (red lines on Figure 3a),  
271 Sutra *et al.* (2013) identified four systematic rift domains. From the continent to  
272 the oceans, these domains are: (1) the *proximal domain*, where top basement  
273 and Moho are sub-horizontal and parallel, and where the  $30 \pm 5$  km-thick crust  
274 displays only minor thinning. When the lithosphere was equilibrated prior to  
275 rifting (i.e. sub-horizontal top basement, Moho and base of the lithosphere),  
276 crustal thickness in the proximal domain may be used as a proxy for pre-rift  
277 crustal thickness; (2) the *necking domain*, where the crust is thinned from  $30 \pm 5$   
278 km to ca. 10 km; (3) the *hyperextended domain*, where the crust is thinned from  
279 ca. 10 km to 0 km. Compared with the necking domain, the angle of aperture  
280 between top basement and Moho is lower and more consistent in the  
281 hyperextended domain (Nirrengarten *et al.*, 2016; Sutra and Manatschal, 2012;  
282 Figure 3a); and (4) the *exhumed mantle domain*, where subcontinental  
283 lithospheric mantle is exhumed to the seafloor (Figure 3a).

284 Based on a worldwide compilation of seismic sections across magma-poor rifted  
285 margins and rift systems (i.e.,  $< 6$  km thickness of magmatic addition prior to  
286 lithospheric breakup; Tugend *et al.*, 2020), Chenin *et al.* (2017) proposed that

287 each domain displays a typical range of width, specifically: necking domains  
288 range between 10 and 100 km, with an average of 55 km; hyperextended  
289 domains between 20 and 70 km, with an average of 50 km; and exhumed  
290 mantle domains between 20 and 110 km with an average of 60 km (Figure 3b).  
291 They highlighted the absence of correlation between the width of the different  
292 domains for a given margin (Figure 3c).

293

### 294 3.2.2. Linking domains to rheological architecture and deformation processes

295 The variability in the angle of aperture between top basement and Moho among  
296 the different domains of rifted margins (red lines in Figure 3a) is interpreted to  
297 reflect along-dip changes in the rheology of the continental crust and underlying  
298 mantle, which may arise from composition, strain-rate, structural and/or  
299 temperature evolution during rifting (Chenin *et al.*, 2017; Nirrengarten *et al.*,  
300 2016; Sutra and Manatschal, 2012; Brune *et al.*, 2014; Ros *et al.*, 2017; Figures  
301 3d and 3e). In the proximal domain, crustal thinning is minor, broadly distributed,  
302 and largely accommodated by ductile deformation within the middle- to lower  
303 crust, which accounts for the horizontal and parallel position of both top  
304 basement and Moho (Sutra *et al.*, 2013). In the necking domain, the  
305 convergence between top basement and Moho is interpreted to reflect the  
306 progressive removal of ductile layers (Pérez-Gussinyé *et al.*, 2003; Sutra *et al.*,  
307 2013). In the hyperextended domain, the lower dip angle of top basement is  
308 thought to express the frictional thinning of a fully brittle continental crust over a  
309 weak, serpentinized mantle (Nirrengarten *et al.*, 2016).

310 Following Péron-Pinvidic *et al.* (2013), we propose that each domain is linked to  
311 a specific deformation mode at play during rifting (Figure 3f): (1) the proximal  
312 domain displays numerous and broadly distributed half-graben depocenters  
313 formed during the *stretching phase* of Lavier and Manatschal (2006). The

314 tectono-thermal evolution of this domain is essentially ruled by the McKenzie  
315 (1978) model; (2) the necking domain is largely shaped by viscoplastic necking  
316 of the lithosphere (Chenin *et al.*, 2018). This *necking phase* corresponds to the  
317 early thinning phase of Lavier and Manatschal (2006). It is characterized by  
318 focusing of deformation into a narrower region compared to the early stretching  
319 phase (20–100 km according to the analytical study by Fletcher and Hallet,  
320 1983; the seismic observations by Chenin *et al.*, 2017; and the numerical  
321 models by Chenin *et al.*, 2018). There, significant crustal thinning is achieved via  
322 detachment faulting (Mohn *et al.*, 2012; Osmundsen and Péron-Pinvidic, 2018;  
323 Ribes *et al.*, 2020); (3) the hyperextended domain is essentially shaped by  
324 Coulomb Wedge-type frictional thinning of fully brittle continental crust  
325 (Nirrengarten *et al.*, 2016). Here, crustal thinning is achieved via in-sequence  
326 detachment faulting in highly damaged continental crust (Epin and Manatschal,  
327 2018; Lymer *et al.*, 2019; Manatschal, 1999). The *hyperextension phase*  
328 corresponds to the late thinning phase of Lavier and Manatschal (2006); and (4)  
329 the exhumed mantle domain records the exhumation of subcontinental mantle  
330 (i.e., the *mantle exhumation phase* of Lavier and Manatschal, 2006) via first in-  
331 sequence detachment faulting and then out-of-sequence or “flip-flop”  
332 detachment faulting (Epin *et al.*, 2019; Gillard *et al.*, 2016a; Reston and  
333 McDermott, 2011). Flip-flop mantle exhumation generates relatively symmetrical  
334 mantle exhumation (Gillard *et al.*, 2016b).

335

336

337 3.2.3. Constraining the spatio-temporal evolution of rifting using the stratigraphic  
338 record

339 Recently, Ribes *et al.* (2019b) studied the migration of deformation across the  
340 Adriatic margin during Alpine Tethys rifting. In each rift domain they identified

341 syn-tectonic sedimentary packages and placed them into a chronostratigraphic  
342 time framework by comparing their position with respect to established timelines  
343 (i.e., sedimentary markers tied to dated global events; Figure 4). From this  
344 analysis, they concluded that: (1) Alpine Tethys rifting was associated with strain  
345 localization, namely the region of active deformation (pink color in Figure 4)  
346 became narrower as rifting advanced, and finally focused into the region of  
347 future breakup; (2) the focusing of deformation was achieved through discrete  
348 steps (i.e., *sequential*) rather than progressively (see the staircase architecture  
349 of the roof of the syn-tectonic sequence, which climbs up between the proximal,  
350 necking, hyperextended and exhumed mantle domain in Figure 4); and (3)  
351 significant amounts of *new real estate* were locally created due to basement  
352 (crust or mantle) exhumation via extensional detachment faults (Figure 4). This  
353 newly exhumed material could only collect sediments that are younger than its  
354 exhumation age, which implies a discontinuous stratigraphic record along dip of  
355 the margin.

356 Comparable conclusions were drawn from the study of high-resolution reflection  
357 seismic sections in the South China Sea by Chao *et al.* (2021) and Luo *et al.*  
358 (2021), in the Porcupine Basin by Whiting *et al.* (2020), in the North Atlantic by  
359 Sharp *et al.* (2017), and from numerical models by Pérez-Gussinyé *et al.* (2020):  
360 all recognized a sequential localization of deformation into the distal domain as  
361 rifting progresses, and/or increasing fragmentation of the pre-rift top basement  
362 with ever wider domains of new real estate created between older fragments  
363 oceanward.

364 Considering that each rift domain (*proximal, necking, hyperextended* and  
365 *exhumed mantle domain*) is dominantly controlled by one specific deformation  
366 mode (*stretching, necking, hyperextension* and *mantle exhumation*, respectively;  
367 Figure 3f), and assuming that these deformation modes are active during distinct

368 periods of time, Ribes *et al.* (2019b) defined four *system tracts*, namely the  
369 *stretching, necking, hyperextension* and *mantle exhumation system tracts*. In the  
370 following, we will refer to these periods as *Tectonic Sequences (TS)*; see next  
371 section for a discussion of terminology-related issues). A *Tectonic Sequence*  
372 encompasses all the sediments deposited during the period of activity of the  
373 corresponding deformation mode.

374

375 3.2.4. A short discussion on the complexity of finding accurate and illustrative  
376 terminology in an already jargon-rich field

377 In seismic interpretation calibrated by well data, it is common to map seismic  
378 *sequences* along rifted margins and to assign them a non-interpretive name  
379 linked to their relative age (e.g., J10, J20, J30, K19, K20, etc.; for instance Alves  
380 *et al.*, 2003). In contrast, in field studies or in interpretation of seismic images  
381 where chronostratigraphic constraints are lacking, it is common to map *system*  
382 *tracts/depositional sequences* or seismic *packages* according to the  
383 characteristic structures they display (e.g., Posamentier and Vail, 1988; Prosser,  
384 1993; Ribes *et al.*, 2019b). Both these "non-interpretative" and "sequence  
385 stratigraphy" approaches are equivalent in 2D, however differences arise when it  
386 comes to analyze the stratigraphy of a propagating or segmented rift system in  
387 3D: In the system tract approach, one correlates deformation phases without  
388 regard on their timing (i.e., the age of their upper and lower boundary may  
389 change ages across the rift system). In contrast, in the non-interpretative  
390 approach, one correlates timelines/isochronous seismic packages without  
391 regard to the tectonic processes they record (i.e., one package of given age may  
392 display the characteristic deformation patterns of the stretching phase in one  
393 segment, and of the hyperextension phase in the adjacent segment).



394 While the term *sequence* often has a time connotation, the term *system tract*  
395 rather relies on sedimentary architectures (e.g., stratal terminations). The term  
396 *package* is probably the most neutral term that can be used to present  
397 observations. Here we chose to use the term *Tectonic Sequence (TS)* to refer to  
398 the sediment body deposited during a given deformation phase to underline that  
399 (1) they are related to and coeval with a specific deformation mode; and (2) the  
400 upper- and lower limit of these packages are isochrons along dip of a given  
401 rifted margin transect (2D). We discuss in section 6.4.3 the difficulties raised by  
402 studying rifted margins in 3D.

403

### 404 3.3. Characteristic Tectonic Sequences (TSs) of an idealized 2D along-dip transect 405 across a rifted margin

406 In the present section, we describe the characteristics of the TSs of our idealized  
407 model (Figure 5a–5g) without going through the details of construction. The  
408 observations on which this model was built derive essentially from the Alps and are  
409 presented in Part I (companion paper by Manatschal *et al.*, this volume). The present  
410 contribution aims to present our idealized model and show its application. Of course,  
411 because it is idealized, our model is simplified (see discussion in section 6.1). It aims  
412 to capture the first-order structures and patterns that can generally be identified at  
413 rifted margins; however, when applying this model, one must always keep in mind  
414 that each rift system is unique and builds on its own inheritance.

415 Based on the field and seismic observations, conceptual models and numerical  
416 modelling results presented in section 3.2, we present the Wheeler diagram of  
417 idealized two-dimensional rifted margins. By *idealized* we mean that the seven  
418 following hypotheses are fulfilled: (1) rifting is achieved via **four distinct and**  
419 **successive extension phases** that are the stretching, necking, hyperextension and  
420 mantle exhumation phases; (2) these extension phases **do not overlap in time,**

421 therefore **each** one is **associated with one single Tectonic Sequence**, namely  
422 stretching, necking, hyperextension and mantle exhumation TS; (3) each extension  
423 phase is **controlled by a single deformation mode**, namely stretching, necking,  
424 hyperextension and mantle exhumation; (4) the **structures** formed during a  
425 deformation phase are **characteristic and specific to the corresponding**  
426 **deformation mode**; (5) each **deformation mode is spatially restricted to the rift**  
427 **domain(s) it affects** (a deformation mode may affect several rift domains); (6) when  
428 a **rift domain** is affected by a deformation mode, its **entire width is affected**; and  
429 (7) rifting is a localizing process during which the **successive deformation modes**  
430 **affect an ever narrower region**.

431

#### 432 3.3.1. The stretching TS

433 The stretching TS records the stretching phase, which is characterized by a low-  
434  $\beta$  ( $\beta \ll 1.5$ , where  $\beta$  corresponds to the ratio *length prior to extension / length at*  
435 *the end of extension*; McKenzie, 1978) pure shear extension of the lithosphere  
436 accommodated by broadly distributed, high-angle normal faulting of brittle layers  
437 (usually the upper crust and upper mantle) and plastic deformation of ductile  
438 ones (usually the middle- and lower crust). The corresponding half-graben  
439 depocenters are distributed over the entire rift domain (Figures 5b and 5g1). The  
440 **stretching TS** will thus display **comparable depositional characteristics in**  
441 **the future proximal, necking and hyperextended domains** (fifth line of Figure  
442 6a), although it will be increasingly overprinted and dismembered by subsequent  
443 extensional events in the distal domains (fifth line of Figure 6b). In contrast, the  
444 stretching TS will be absent in the exhumed mantle domain since no mantle is  
445 exhumed to the seafloor during the stretching phase.

446 The stretching TS is characterized by **wedge-shaped sedimentary deposits**  
447 that thicken toward the main controlling normal faults (Figure 5g1 and fifth line of

448 Figure 6a). Related deposits often include breccias derived from the adjacent  
449 footwall crest and fault plane. Indeed, the activity of a normal fault is usually  
450 associated with footwall uplift and erosion due to the rotation of the fault block  
451 (Yielding, 1990) and/or to the flexural response of the lithosphere (Kusznir and  
452 Ziegler, 1992). Most of the infill is however linked to axial sedimentation  
453 (Gawthorpe and Leeder, 2000; Leeder and Gawthorpe, 1987).

454 Accommodation creation during the stretching phase is predominantly created  
455 by fault-controlled subsidence within rift basins. No significant lithospheric  
456 thinning occurs during this stage, so that only limited accommodation is created  
457 outside the basins, and only minor subsequent thermal subsidence is recorded  
458 across the future proximal domain. Moreover, the usually high ratio between the  
459 sedimentation- and accommodation creation rates hampers the formation of  
460 significant bathymetry during the stretching phase (fifth line of Figure 6c).

461

### 462 3.3.2. The necking TS

463 The necking TS records the major phase of crustal thinning, from ca. 30 to  
464 10 km ( $1.5 < 1/\beta < 3$ ). During the early stages of the necking phase, a few large-  
465 offset (up to few tens of km; e.g., Osmundsen and Péron-Pinvidic, 2018; Ribes  
466 *et al.*, 2020; Sutra *et al.*, 2013) detachment faults form on either side of a largely  
467 undeformed crustal **keystone** (*H-block* in Lavier and Manatschal, 2006; Figure  
468 5c). These major **crustal-scale detachment faults**, which may **exhume mid-**  
469 **crustal levels to the seafloor** (Ribes *et al.*, 2020), are specific and  
470 characteristic structures of the future necking domains. The isostatic uplift  
471 resulting from detachment-related tectonic denudation (Spencer, 1984) may  
472 result in **erosion** of the exhumed basement and nearby deposition of  
473 **siliciclastic sediments/sandstones** (e.g., the *Grès Singuliers* in the Mont

474 Blanc region; Ribes *et al.*, 2020, and references therein), ideally with a *supra-*  
475 *detachment wedge-shaped* architecture.

476 At the same time, most of the normal faults formed during the stretching phase  
477 are inactivated outside the necking domain. As a result, **the necking TS**  
478 **outside the necking domain is characterized by the sealing of the older**  
479 **half-graben depocenters by post-tectonic, passively infilling sediments**  
480 (fourth line of Figure 6a; e.g., Masini *et al.*, 2013). As no mantle is exhumed to  
481 the seafloor during the necking phase, the necking TS is absent in the exhumed  
482 mantle domain.

483 The onset of crustal necking is linked to the yield of the strongest lithospheric  
484 layer, which is usually the upper mantle in a thermally equilibrated lithosphere  
485 (Figure 3d). When the upper crust is mechanically decoupled from the mantle,  
486 the mechanical breakdown of the upper mantle is followed by a flexural rebound  
487 that tends to bring the necked system back to isostatic equilibrium (Braun and  
488 Beaumont, 1989). This flexural rebound accounts for the shallow-water depth, or  
489 even the **uplift and local erosion of the keystone** (i.e., the future necking and  
490 hyperextended domains) during the necking phase (see erosional unconformity  
491 in Figures 5c, 5d and 5g5; Chenin *et al.*, 2019, 2020). Uplift of the keystone may  
492 be further increased by the thermal erosion related to the asthenosphere  
493 upwelling beneath the rift system.

494 The late necking phase (Figure 5d) is linked to significant crustal and  
495 lithospheric thinning, which will eventually counterbalance the flexural rebound  
496 of the lithosphere and cause rapid and significant subsidence (e.g., Ribes *et al.*,  
497 2020). The timing of uplift vs subsidence depends strongly on the rheology of  
498 the lithosphere (Chenin *et al.*, 2020). Subsidence might be further enhanced by  
499 large sediment supply and by subsequent thermal relaxation of the rift system.

500 In summary, **in the necking domain of the final margin, the necking TS** is  
501 expressed by (fourth line of Figures 6a–6c): (1) **creation of new real estate** due  
502 to the exhumation of upper- to middle crustal rocks to the seafloor via  
503 extensional detachment faulting; (2) ***supra-detachment wedge-shaped***  
504 **sedimentary deposits of necking phase age** emplaced **directly onto**  
505 **exhumed basement**; (3) uplift and **erosion of exhumed basement and pre-**  
506 **necking sediments** potentially visible **in both the necking and**  
507 **hyperextended domains**; and (4) **contemporaneous deposition of**  
508 **basement-derived material in adjacent regions**. These sediments may be  
509 “cannibalized” (i.e., remobilized and re-deposited) during the necking phase  
510 owing to the progressive exhumation–uplift–rollover of the detachment fault on  
511 which the sandstones were deposited.

512

### 513 3.3.3. The hyperextension TS

514 The hyperextension TS records the localized dismembering of fully brittle  
515 continental crust from 10 to 0 km thick ( $3 < 1/\beta < \infty$ ). Hyperextension is the first  
516 deformation stage to display a marked asymmetry (Nirrengarten *et al.*, 2016;  
517 Ranero and Pérez-Gussinyé, 2010): on the one hand, the future lower-plate  
518 margin is extended via in-sequence detachment faulting; and on the other hand,  
519 the future conjugate upper-plate margin is delaminated by the slip on the  
520 underlying detachment fault (Figures 5e, 5g2 and 5g4; Hauptert *et al.*, 2016).  
521 Each of these deformation processes results in a particular morphology of the  
522 margin (see below).

523 During hyperextension, the proximal and necking domains are tectonically  
524 inactive, which results in continued passive infill in the proximal domain and the  
525 sealing of syn-tectonic structures/deposits in the necking domain (third line of

526 Figure 6a). Thus, **in the proximal and necking domains of the future margin,**  
527 **the hyperextension TS is characterized by post-tectonic deposits.** Due to  
528 the lack of mantle exhumation at this stage, the hyperextension TS is absent in  
529 the exhumed mantle domain.

530 The hyperextension phase is associated with **rapid subsidence in both the**  
531 **necking and hyperextended domains** (third line of Figure 6c). Subsidence is  
532 related to extreme crustal thinning above a mechanically weak mantle (the  
533 mechanical breakup of the upper mantle was achieved during the previous  
534 necking phase). In the necking domain, subsidence is at this stage essentially  
535 driven by thermal relaxation (i.e., lithosphere cooling).

536 In the lower-plate hyperextended domain, each successive in-sequence  
537 detachment fault crosscuts the former one, resulting in the typical in-sequence  
538 architecture illustrated in Figure 5g2 and 6a (Ranero and Pérez-Gussinyé,  
539 2010). These detachment faults create new real estate via the exhumation of  
540 basement to the seafloor (Figure 5g2). The amount of basement exhumed by a  
541 detachment fault was reckoned to 10–20 km along-dip in the hyperextended  
542 domain of the Adriatic lower-plate margin (Alpine Tethys) by Epin and  
543 Manatschal (2018). The exhumed basement is typically pre-rift upper- to middle  
544 crust. Each detachment fault is associated with a supra-detachment  
545 sedimentary basin floored by tectono-sedimentary breccias. These basins are  
546 filled with supra-detachment wedge-shaped syn-tectonic sediments emplaced  
547 directly onto the detachment surface, getting younger toward the detachment  
548 exhumation point (third line of Figure 6a; Masini *et al.*, 2011). As the successive  
549 in-sequence detachment faults are active one after another, syn-tectonic  
550 sedimentary deposits become younger oceanward (Péron-Pinvidic *et al.*, 2007).  
551 Therefore, sedimentary deposits of a given age that are syn-tectonic in one

552 supra-detachment basin are post-tectonic in supra-detachment basins located  
553 more continent-ward.

554 In summary, **in the hyperextended domain of the final lower-plate margin,**  
555 **the hyperextension TS is expressed by** (third line of Figures 6a– 6c):  
556 (1) creation of **new real estate** due to the exhumation of upper- to middle crustal  
557 rocks to the seafloor via **in-sequence detachment faults that dip exclusively**  
558 **oceanward;** (2) **tectono-sedimentary breccias and overlying supra-**  
559 **detachment wedge-shaped sedimentary deposits of hyperextension phase**  
560 **age emplaced directly onto exhumed basement;** and (3) a progressive  
561 **younging of supra-detachment basins fill oceanward.**

562 The hyperextended domain of upper-plate margins represents the residual part  
563 of the initial keystone block (Hauptert *et al.*, 2016; Figure 5g4). Compared with its  
564 conjugate, the upper-plate hyperextended domain preserves therefore much  
565 more of the stretching and necking-related structures and deposits, namely high-  
566 angle normal faults that can dip continent- or oceanward. The base of the outer  
567 part of the hyperextended domain is a mantle exhumation fault that may  
568 eventually bring lower crustal rocks to the seafloor.

569 Ribes *et al.* (2019a) suggested that the individualization of the upper-plate  
570 hyperextended domain is often associated with the formation of a *mega-fault*  
571 *scarp* at its boundary with the adjacent necking domain. They defined a *mega-*  
572 *fault scarp* as a "topographic ramp [formed by a] system of high-angle faults and  
573 facing oceanwards a domain of substantial underfilled accommodation space".  
574 They noticed that mega-fault scarps are linked with the emplacement of breccias  
575 at their toe, over a wide part of the hyperextended domain (Figure 5g4).

576 In summary, **in the hyperextended domain of the final upper-plate margin,**  
577 **the hyperextension TS is expressed by:** (1) creation of **a mega-fault scarp**

578 **and** associated **mega-breccias** at the boundary between the necking and future  
579 hyperextended domain; (2) good preservation of earlier sedimentary structures  
580 due to the **lack of significant new real estate creation**; and (3) possible  
581 **exhumation of lower crustal rocks to the seafloor** at the outer edge of the  
582 hyperextended domain.

583

#### 584 3.3.4. The mantle exhumation TS

585 The mantle exhumation TS records the phase of subcontinental mantle  
586 exhumation to the seafloor (Figure 5f). Mantle exhumation occurs while tectonic  
587 activity has ceased in the continental domains of the future margin, **therefore**  
588 **the mantle exhumation TS in the proximal, necking and hyperextended**  
589 **domains is characterized by post-tectonic sedimentation** (second line of  
590 Figure 6a). The entire margin keeps on subsiding during the mantle exhumation  
591 phase.

592 The exhumed mantle domain is entirely comprised of new real estate, except  
593 from few potential allochthonous blocks of continental crust and/or pre to syn-rift  
594 sediments. Subcontinental mantle exhumation is achieved via out-of-sequence  
595 flip-flop detachment faults that develop successively on either side of the future  
596 spreading center, generating a relatively symmetrical pattern (Epin *et al.*, 2019;  
597 Gillard *et al.*, 2016b; Figure 5g and second line of Figure 6a). The activity of a  
598 mantle exhumation fault is associated with the formation of a topographic high at  
599 the detachment fault breakaway (Spencer, 1984), and thus characterized by  
600 over-tilted sedimentary sequences (see to the very right of Figure 5g3; Gillard *et*  
601 *al.*, 2015). As tectonic activity continuously steps toward the locus of future  
602 breakup, the age of the over-tilted sequences decreases (youngs) oceanward.  
603 The first sediments overlying the exhumed mantle are tectono-sedimentary



604 breccias, and if sedimentary sources are available, the next ones are supra-  
605 detachment wedge-shaped syn-tectonic sediments deposited directly onto the  
606 detachment surface.

607 The exhumation phase often displays incipient magmatic activity, which results  
608 in discontinuous magmatic intrusions and lava flows (Coltat *et al.*, 2019; Epin *et*  
609 *al.*, 2019; Gillard *et al.*, 2016b, 2015). Magmatic flows are syn-tectonic, and thus  
610 they are interbedded with sediments of exhumation phase age.

611 In summary, **in the exhumed mantle domain of the final margin, the mantle**  
612 **exhumation TS is expressed by** (second line of Figures 6a and 6b):  
613 (1) **exclusively new real estate basement** comprised of **subcontinental**  
614 **lithospheric mantle** exhumed to the seafloor via **out-of-sequence**  
615 **detachment faulting**; (2) **tectono-sedimentary breccias** and overlying **supra-**  
616 **detachment wedge-shaped sedimentary deposits** of mantle exhumation  
617 phase age emplaced directly onto exhumed mantle; (3) **over-tilted**  
618 **sedimentary packages** at detachment faults breakaway; and (4) a progressive  
619 **younging of supra-detachment basins fill oceanward.**

620

#### 621 4. Illustration of our model with calibrated natural examples

622

623 The best calibrated rifted margins worldwide (i.e., constrained by either outcrop  
624 observations or drilling) are by far those of the fossil Alpine Tethys exposed in the Alps.  
625 This example is described in detail in the companion paper by Manatschal *et al.* (this  
626 volume) and only briefly summarized here (Figures 7c and 7d). Another well calibrated  
627 example is the conjugate transect across the Iberia–Newfoundland margins imaged by  
628 the TGS–LG 12 and SCREECH 2 reflection seismic profiles, respectively. In addition to  
629 drill hole data from the proximal domain and inner part of the necking domain (e.g., Alves  
630 *et al.*, 2002, on the Iberian side; Sinclair, 1994 and references therein; Wielens *et al.*,

631 2006, on the Newfoundland one), deep-water drilling was performed in the Iberian  
632 hyperextended domain and in the inner part of the Newfoundland exhumed mantle  
633 domain, as part of the Ocean Drilling Program (ODP) Legs 149, 173 and 210. Besides,  
634 the viability of the tectono-stratigraphic interpretation by Sutra *et al.* (2013) was supported  
635 by its surface-conservative kinematic restoration.

636

#### 637 4.1. Geological setting of Mesozoic rifting events in Western Europe

638 The opening of the Alpine Tethys and North-Atlantic Ocean was part of the  
639 fragmentation of the Pangea supercontinent, which progressed both from east to  
640 west and from south to north via *multistage rifting* (Figure 7a): it initiated with the  
641 opening of the Meliata-Vardar domain at the western end of the Neothethys. It  
642 continued with the opening of the Ligurian, Piemonte and Valais domains between  
643 the Eurasian/Iberian plate and the Adriatic/African plate, ahead of the northward-  
644 propagating Central Atlantic rifting, while the Meliata–Vardar domain was closing. It  
645 continued with the opening of the Bay of Biscay and southern North-Atlantic Ocean,  
646 while extension had ceased in the Alpine Tethys realm (Angrand *et al.*, 2020; Le  
647 Breton *et al.*, 2020).

648 A broadly distributed extensional event affected a large part of Western Europe and  
649 eastern North America from the Permo-Triassic onward (Leleu *et al.*, 2016, and  
650 references therein), and is here regarded as the stretching phase of both the Alpine  
651 Tethys and Iberia–Newfoundland rifting. In the Alpine Tethys rift system, the onset of  
652 localized deformation (necking phase) started at  $185 \pm 5$  Ma (Sinemurian–  
653 Pliensbachian), evolved to hyperextension from  $175 \pm 5$  Ma (Toarcian–Aalenian),  
654 and culminated with mantle exhumation from  $165 \pm 5$  Ma onward (Bajocian–  
655 Bathonian; Manatschal *et al.*, this volume). Rifting stopped in the Alpine Tethys  
656 realm before steady-state seafloor spreading occurred in any branch of the Alpine rift  
657 system (Picazo *et al.*, 2016). Then, focused deformation started between Iberia and

658 Newfoundland with the necking phase from Tithonian to Berriasian time, continued  
659 with hyperextension from latest Berriasian to Valanginian time, and with mantle  
660 exhumation from the Valanginian (Mohn *et al.*, 2015, and references therein) until  
661 the late Aptian–early Albian lithospheric breakup (Bronner *et al.*, 2011). Note that the  
662 Iberia–Newfoundland rift system comprises three main segments that underwent  
663 diachronous extension from south to north (Alves *et al.*, 2009, and references  
664 therein). Here we focus on the central segment (i.e., South Iberia Abyssal Plain–  
665 Northern Newfoundland Basin).

666

## 667 4.2. The fossil Alpine Tethys rift system

### 668 4.2.1. The proximal domains

669 The former proximal domains of the European and Adriatic margins were  
670 associated with numerous half-graben depocenters (Figure 7b), among which  
671 the best-exposed are the Bourg d’Oisans Basin on the European margin  
672 (Chevalier, 2002; Dumont *et al.*, 2008), the Il Motto Basin on the northern  
673 Adriatic margin (Conti *et al.*, 1994; Eberli, 1988; Ribes *et al.*, 2019b), and the  
674 Monte Generoso Basin on the former southern Adriatic margin (Bernoulli, 1964;  
675 Bertotti *et al.*, 1993).

676 The sedimentary record is comparable in the three basins, displaying half-  
677 graben wedge-shaped deposits from latest Triassic to Sinemurian–  
678 Pliensbachian time (Figures 7c and 7d). In the temporal framework of the Alpine  
679 Tethys rifting, this corresponds to the early syn-rift history, namely the *stretching*  
680 *phase*. From Pliensbachian–Toarcian time onward, the three basins record  
681 passive infill (i.e., post-tectonic deposits belonging to the syn-rift  
682 megasequence).

683

684 4.2.2. The necking domains

685 The best described necking domain in the Alps used to belong to the former  
686 European margin and now lies in the Mont Blanc region (Ribes *et al.*, 2020;  
687 Figure 7b). Necking domains of the former northern- and southern Adriatic  
688 margin were respectively identified in the Campo–Grosina region (no  
689 sedimentary record available) by Mohn *et al.* (2010), and in the Cusio–Biellese  
690 region by Beltrando *et al.* (2015). Besides, the sedimentary record of the  
691 intervening keystone block is preserved in the Briançonnais region (Claudel and  
692 Dumont, 1999; Hauptert *et al.*, 2016).

693 Observations from the Mont Blanc (Ribes *et al.*, 2020, and references therein)  
694 and Cusio–Biellese region (Beltrando *et al.*, 2015, and references therein) show  
695 that (Figures 7c and 7d): (1) the base of the necking TS is a tectonic  
696 unconformity owing to detachment-related basement exhumation (new real  
697 estate) and/or local erosion; (2) the necking domain is a local region of  
698 siliciclastic deposits (sandstones) due to local basement exhumation and  
699 erosion/deposition; (3) water-depth remained relatively shallow during the  
700 necking phase and deposition of related sandstones. Observations from the  
701 Briançonnais domain indicate that the keystone block underwent a phase of  
702 transient emergence during the necking phase.

703

704 4.2.3. The hyperextended domains

705 The outer edge of a European upper plate is exceptionally well-preserved in the  
706 Tasna Nappe of southeastern Switzerland (Manatschal *et al.*, 2006; Ribes *et al.*,  
707 2019c; Figure 7b). The distal edge of the northern Adria lower plate is exposed  
708 in the Err and Upper Platta nappes (Figure 7b) in Southeast Switzerland (Epin  
709 and Manatschal, 2018; Manatschal and Nievergelt, 1997; Masini *et al.*, 2012).

710 The hyperextended domain of South Adria is recognized in the Canavese region  
711 of the Southern Alps. It was recently interpreted as an upper plate by Decarlis *et*  
712 *al.* (2017).

713 The tectono-sedimentary record of the Tasna, Err and Canavese nappes is  
714 comparable in that the three of them display (Figures 7c and 7d):  
715 (1) discontinuous pre-rift sediments only present as/over allochthonous blocks;  
716 (2) basement exhumation via detachment faults overlain by tectono-sedimentary  
717 breccias; and (3) deep-marine syn-tectonic sedimentary deposits. They differ in  
718 that no lower crust seems to be exhumed at former lower plates (Err Nappe), in  
719 contrast to former upper plates (Tasna Nappe and Canavese Zone).

720

#### 721 4.2.4. The exhumed mantle domains

722 The best-preserved outcrops of an exhumed mantle domain and embryonic  
723 seafloor spreading system in the Alps are exposed in the Lower Platta Nappe  
724 (Figure 7b) in southeastern Switzerland (Desmurs *et al.*, 2002; Epin *et al.*, 2019;  
725 Ribes *et al.*, 2019b) and at the Chenaillet Ophiolite in the French-Italian Alps  
726 (Manatschal *et al.*, 2011).

727 The tectono-sedimentary record of the Platta Nappe and Chenaillet Ophiolite  
728 shows that (Figures 7c and 7d): (1) mantle exhumation was achieved via  
729 successive in-sequence detachment faults; (2) topography was created during  
730 mantle exhumation, which resulted in gravitational sliding and/or onlapping of  
731 syn-exhumation magmatic products and sediments; (3) the exhumed mantle  
732 was affected by normal faults, which served as feeder channels for the overlying  
733 volcanic sequences; (4) no pre- and early syn-rift deposits existed in the  
734 exhumed mantle domain; and (5) the first sedimentary deposits (radiolarites)  
735 were characteristic of abyssal deposition environments.

736

### 737 4.3. The Iberia–Newfoundland conjugate margins

#### 738 4.3.1. The proximal domains

739 The proximal domains of the Iberia and Newfoundland margins (Figure 8a) are  
740 characterized by minor stretching factors ( $\beta < 2$ ; Mohn *et al.*, 2015), except in  
741 the northern Jeanne d’Arc Basin that was affected by later extension related to  
742 the formation of the Orphan Basin further north (Skogseid, 2010). Moho and top  
743 basement are sub-horizontal at the scale of both proximal domains (Sutra *et al.*,  
744 2013; Figures 8b and 8b1). Extension formed half-graben depocenters such as  
745 the (South) Jeanne d’Arc, Whale and Horseshoe basins on the Newfoundland  
746 side, and the Porto and Lusitanian basins on the Iberian side (Tucholke *et al.*,  
747 2007).

748 The northern Lusitanian- and Jeanne d’Arc basins (Figure 8a) display a  
749 comparable evolution and sedimentary fill indicating discontinuous tectonic  
750 activity from Late Triassic to Late Jurassic (Alves *et al.*, 2002; Sinclair, 1993).  
751 The first rift-related deposits in both basins are Late Triassic to Pliensbachian  
752 continental and/or lacustrine sediments and evaporites, representative of  
753 terrestrial to epicontinental sea environments. They show *half-graben wedge-*  
754 *shape* sedimentary architectures, indicative of their syn-tectonic setting  
755 (*stretching TS*). Subsequent marine sandstones and carbonates alternating with  
756 shales were passively deposited from the Pliensbachian–Callovian in the  
757 northern Lusitanian Basin (Iberian margin) and from the Tithonian in the central  
758 Lusitanian Basin and at the Newfoundland margin. They were interpreted to  
759 record post-tectonic thermal relaxation at both margins (Alves *et al.*, 2002; Alves  
760 *et al.*, 2003; Enachescu, 1987; Hubbard, 1988; McAlpine, 1990; Sinclair, 1988;  
761 Ravnås *et al.*, 1997).

762 On both the Iberian and Newfoundland margins, Callovian–Oxfordian deposits  
763 were affected by an erosional unconformity (Ian Sharp personal  
764 communication). This unconformity preceded renewed extension, during which  
765 marine sandstones were deposited in the northern Lusitanian Basin (Alves *et al.*,  
766 2003, 2002; Ravnås *et al.*, 1997). Renewed extension in the Jeanne d’Arc Basin  
767 during the Tithonian was preceded or accompanied by erosion (Sinclair, 1993,  
768 and references therein) with alternating carbonates and shales being  
769 subsequently deposited.

770 In the Lusitanian Basin, the onset of post-tectonic sedimentation occurred in the  
771 late Tithonian–early Berriasian (Ravnås *et al.*, 1997). Sediments younger than  
772 Kimmeridgian are comprised of shallow-marine to continental units that display  
773 post-tectonic architectures and testify to limited thermal subsidence (Alves *et al.*,  
774 2003). In the Jeanne d’Arc Basin, the onset of post-tectonic sedimentation was  
775 recognized during the Valanginian (CNLOPB, 2010; Enachescu, 1987; Sinclair,  
776 1988). Sediments include sandstones and shales. Subsequent tectono-  
777 stratigraphic evolution in the Jeanne d’Arc Basin was affected by extensional  
778 events in the adjacent Orphan Basin (Skogseid, 2010) and is therefore not  
779 discussed here.

780

#### 781 4.3.2. The necking domains

782 Alves *et al.* (2006) interpreted the necking zone of the Iberian margin to be  
783 structured by only high-angle normal faults but noticed the lack of evidence for  
784 pre-rift sediments in its outer part. Sutra *et al.* (2013) and Mohn *et al.* (2015) also  
785 interpreted the inner Iberian necking domain to contain a large normal fault  
786 (Figure 8b) but suggested that top basement in the outer part of the necking

787 domain is made of new real estate basement exhumed by a detachment fault  
788 (Figure 8b2).

789 The necking domain of the Newfoundland margin has traditionally been  
790 regarded as structured by only high-angle normal faults, like the rest of the  
791 margin (Causer *et al.*, 2020; Hopper *et al.*, 2006; Welford *et al.*, 2010). However,  
792 Péron-Pinvidic and Manatschal (2009), Sutra *et al.* (2013) and Mohn *et al.*  
793 (2015) interpreted the inner necking fault of the Newfoundland margin as a high-  
794 angle normal fault that started to evolve as a detachment fault (minor amount of  
795 basement exhumed at low angle; Figure 8b). Nobody has recognized new real  
796 estate creation in the rest of the Newfoundland necking domain so far.

797 Drill holes in the northern part of Carson Basin (Newfoundland margin; Figure  
798 8a) indicate a Tithonian unconformity overlain by sandstone deposits, like that  
799 reported in the Jeanne d'Arc Basin (Ian Sharp personal communication). A later  
800 Albian unconformity in the northern Carson Basin is interpreted to mark the rift to  
801 drift transition in the southern segment (Wielens *et al.*, 2006).

802 In the Iberian necking domain (Peniche Basin; Figure 8a), Alves *et al.* (2006)  
803 interpreted Triassic–Late Jurassic syn-tectonic deposits in the depocenter  
804 adjoining the main necking fault. These would correspond to the sedimentary fill  
805 of a stretching phase half-graben depocenter. Such deposits are lacking in the  
806 outer part of the necking domain. Latest Berriasian to Aptian sediments cover  
807 the entire necking domain and were interpreted as post-tectonic (Alves *et al.*,  
808 2006; Mohn *et al.*, 2015). Alves *et al.* (2006) reported toplaps of Albian age,  
809 contemporaneous to the erosional unconformity reported in the Carson Basin.

810

811 4.3.3. The hyperextended domains



812 The hyperextended domain of the Iberian margin is consensually interpreted as  
813 structured by detachment faults that exhumed upper- to mid-crustal basement to  
814 the seafloor (Krawczyk *et al.*, 1996; Manatschal *et al.*, 2001; Reston and  
815 McDermott, 2011; Figure 8b3). In contrast, the hyperextended domain of the  
816 Newfoundland margin is regarded as essentially controlled by high-angle normal  
817 faulting that crosscut the entire crust (Mohn *et al.*, 2015; Péron-Pinvidic and  
818 Manatschal, 2009; Sutra *et al.*, 2013). This distinction prompted Manatschal  
819 (2004) and Hauptert *et al.* (2016) to interpret the Iberian margin as a lower plate  
820 and its Newfoundland conjugate as an upper plate.

821 Drill holes in the hyperextended domain of the Iberian margin indicate upper  
822 crustal basement affinities (ODP Sites 639, 901, 1065 and 1069). Exhumed  
823 middle- or lower crust has been evidenced only at ODP Sites 900 and 1067  
824 (outer edge of the hyperextended domain). No drill hole data from the  
825 hyperextended domain of the Newfoundland margin is available.

826 None of the drill holes in the Iberian hyperextended domain penetrated Late  
827 Triassic–Early Jurassic sediments. However, since all but one of the deep-water  
828 drill holes (ODP Site 639 in the northern segment) stopped before reaching the  
829 pre-rift top basement, we cannot exclude that such sequences exist, or existed  
830 and were eroded prior to hyperextension.

831 In the Iberian hyperextended domain, pre-rift top basement is overlain by  
832 concordant Mid- to Late Jurassic sandstones and conglomerates interpreted as  
833 emplaced in an offshore shelf environment (ODP Sites 1065 and 1069; Figure  
834 8b2). We consider that these syn-rift deposits, which display pre-tectonic  
835 architectures, represent the stretching TS in the hyperextended domain. The  
836 subsequent Tithonian–Berriasian deposits are still pre-tectonic in the central  
837 segment. They comprise clay, silty clay, coarse sands and conglomerates  
838 interpreted to reflect a deepening sequence (up to 500–1000 m; ODP Sites 901,

839 1065 and 1069). We consider that they represent the necking TS in the  
840 hyperextended domain. Whether the sands/conglomerates are derived from  
841 basement erosion is not resolved to date.

842 The sedimentary fill of supra-detachment basins was only drilled in the  
843 hyperextended domain of the northern Iberian segment (ODP Sites 638–641).  
844 There, it consists of debris flow, turbidites and slumps alternating with  
845 background sedimentation. The overall sedimentary sequence displays a  
846 wedge-shaped architecture (Boillot and Winterer, 1988). In the hyperextended  
847 domain of the central Iberian segment, Péron-Pinvidic *et al.* (2007) suggested  
848 that the lack of obvious growth structures may be related to syn-basement  
849 exhumation sedimentation in a largely sediment-starved rift system (i.e., the  
850 sediment-starved version of our *supra-detachment wedge-shaped* sediment  
851 architecture; see Figure 2f). They also noticed an oceanward younging of in the  
852 sedimentary fill of supra-detachment basins, where coeval sedimentary  
853 sequences display post-tectonic architectures in the inner part of the  
854 hyperextended domain, and syn-tectonic architectures in the outer part  
855 (Figure 8b3).

856 At ODP Site 1069, nanofossils (chalk) of late Berriasian to early Valanginian  
857 age overlie the tilted (pre-tectonic) Tithonian to older sedimentary sequence.  
858 These hemipelagic to pelagic sediments were interpreted as being deposited in  
859 open-marine outer-shelf to upper-slope environments (> 1500 m; Urquhart,  
860 2001). They highlight the shutdown of the terrigenous input in the outermost part  
861 of the hyperextended domain during late hyperextension, possibly due to  
862 sediment capture by inner basins (Wilson *et al.*, 2001). Mohn *et al.* (2015)  
863 attributed the essentially post-tectonic infill of the supra-detachment basins to  
864 the short (5–10 My) duration of the hyperextension phase.

865

866 4.3.4. The exhumed mantle domains

867 Wilson *et al.* (2001) suggested that exhumation of already serpentized mantle  
868 started during Valanginian–Hauterivian time. While Manatschal (2004) initially  
869 envisioned mantle exhumation via a single detachment fault, Gillard *et al.* (2016)  
870 identified four successive in- and out-of-sequence detachment faults on the  
871 SCREECH 1 seismic profile further north, across the exhumed mantle domain of  
872 the Newfoundland margin.

873 At ODP Site 1068 (mantle exhumation point; Figure 8b2), drilling reported  
874 tectono-sedimentary breccias supported by a calcareous chalk matrix of  
875 Valanginian–Barremian age deposited onto exhumed mantle (Figure 8b4). The  
876 next sediments drilled are much younger Campanian turbidites, which onlap  
877 against the tilted tectono-sedimentary breccias. This relationship indicates either  
878 that ODP Site 1068 was a topographic high already at the onset of mantle  
879 exhumation, or that the tectono-sedimentary breccias were tilted and uplifted  
880 after their deposition and prior to Campanian time, and that the adjacent basin  
881 was progressively filled by turbidites and pelagic sediments. Comparable  
882 relationships were documented by Péron-Pinvidic *et al.* (2007) at the conjugate  
883 Newfoundland margin (ODP Site 1277; Figure 8b5): breccias of presumably  
884 Aptian age and overlying exhumed mantle are over-tilted compared to the Albian  
885 to Cenomanian onlapping, sub-horizontal, passive infill.

886 Magmatic activity contemporaneous with lithospheric breakup is evidenced by  
887 diabase sills that were emplaced at the Aptian–Albian boundary (Jagoutz *et al.*,  
888 2007; Karner and Shillington, 2005).

889

890 4.3.5. Summary

891 The primary spatio-temporal evolution of deformation in the Iberia–  
892 Newfoundland rift system can be deduced from the architecture of the main  
893 sedimentary package interfaces (Figure 8b2): in the inner necking domain, the  
894 blue interface marks the top of the syn-tectonic succession, while in the  
895 hyperextended domain, it marks the top of the pre-tectonic succession. This  
896 implies that deformation was active in the inner necking domain when no  
897 deformation had occurred in the hyperextended domain yet.

898 In the hyperextended domain, a comparable evolution can be observed  
899 (Figure 8b3): in the inner part, the red horizon lies within a post-tectonic  
900 succession, while it marks the top of syn-tectonic sequence in the outer part.  
901 The overlying green horizon lies within a post-tectonic succession along the  
902 entire hyperextended domain, while it marks the top of a syn-tectonic  
903 succession in the adjacent exhumed mantle domain. This indicates that active  
904 deformation was progressively advancing oceanward.

905 The tectono-stratigraphic history of the Iberia–Newfoundland rift system is  
906 summarized in Figure 8c. It is consistent with our idealized model in that: (1) it  
907 indicates a sequential localization of deformation oceanward; (2) the proximal  
908 domain is only affected by high-angle normal faults whose tectonic activity has  
909 largely ceased by the end of the stretching phase; (3) syn-stretching and syn-  
910 necking deposits in the outer necking and hyperextended domain of the Iberian  
911 margin display shallow-water facies comparable to those of the proximal  
912 domain; however, they display pre-tectonic stratigraphic architectures in the  
913 distal margin, while they are syn-tectonic in the proximal domain; (4) a syn-  
914 necking (Tithonian) unconformity overlain by sandstones exists at least in the  
915 proximal and necking domain of the Newfoundland margin; (5) the  
916 hyperextended Iberian lower plate is structured by in-sequence detachment  
917 faults associated with supra-detachment wedge-shaped deposits that become

918 younger oceanward; and (6) the exhumation domain displays over-tilted syn-rift  
919 pre- to syn-tectonic sequences against which post-rift sediments onlap.

920 Debated questions concern (1) the existence of extensional detachment faults in  
921 the necking domain of both the Iberian and Newfoundland margins; and (2) the  
922 existence of (dismembered) half-graben wedge-shape deposits in the stretching  
923 TS of the hyperextended domain.

924

## 925 5. Application of our model to incompletely calibrated natural examples

926 In the following, we rely on our model to interpret the tectono-stratigraphic record of an  
927 uncalibrated transect across conjugate NW-South China Sea margins; and of transects  
928 across the distal domain of the Mid-Norwegian and Socotra margins with support of  
929 observations and data from their adjacent, better-calibrated proximal domains.

### 930 5.1. The northwest South China Sea conjugate margins

#### 931 5.1.1. Geological setting

932 The NW South China Sea is an aborted, V-shaped oceanic basin that benefits  
933 from extensive seismic coverage and some drill holes in the most distal part of  
934 the margin, ca. 200 km east of the transect presented below (International  
935 Ocean Discovery Program (IODP) Legs 349, 367 and 368; Figure 9a). The  
936 South China Sea differs largely from Atlantic-type rifts in that the rheology of the  
937 pre-rift lithosphere was significantly weaker due to the high geothermal gradient  
938 related to its back-arc position and orogenic and post-orogenic history (Li *et al.*,  
939 2018; Sun *et al.*, 2019).

940 Rifting of the NW South China Sea is considered to have begun at ca. 65 Ma  
941 (Xie *et al.*, 2019), although this age is poorly constrained. Besides, because of  
942 the propagating nature of extension, rifting was largely diachronous along strike.

943 The onset of seafloor spreading is dated to 30–28 Ma based on the presence of  
944 magnetic anomalies in the oceanic crust (Gee and Kent, 2007).

945 The NW South China Sea has been widely studied, for instance by Cameselle,  
946 *et al.* (2017); Gao *et al.*, (2016); Larsen *et al.* (2018); Nirrengarten *et al.* (2020);  
947 Ren *et al.* (2018); Zhang *et al.* (2021). Here we focus on the CGN-1  
948 multichannel high-resolution 2D seismic section, for which Chao *et al.* (2021)  
949 recently proposed a detailed tectono-sedimentary interpretation and balanced  
950 kinematic restoration. To study the spatio-temporal evolution of deformation  
951 during rifting, they grouped seismic packages of same interpreted age along dip  
952 into one *Sequence (S)*. In total, they identified five syn-rift *Sequences* (packages  
953 of same interpreted age along dip) with distinct evolutions across the conjugate  
954 margins.

955

#### 956 5.1.2. The proximal domain of the northern margin

957 The proximal domain of the northern margin is characterized by sub-horizontal  
958 top basement and Moho separated by a ca. 22 km-thick continental crust  
959 (Figure 9b). Top basement and pre-rift sediments are affected by only minor  
960 normal faulting (fault throw < 0.2 sTWT). The faulted top basement is directly  
961 overlain by post-rift deposits, which prevents any accurate dating of crustal  
962 faulting. The proximal domain of the southern margin is not imaged and thus is  
963 not discussed here.

964

#### 965 5.1.3. The necking domains

966 On the northern margin, the inner limit of the necking domain is a crustal-scale  
967 high-angle normal fault with a fault throw of 1.5 sTWT. 10–15 km further  
968 outboard, a major detachment fault exhumed upper- to mid-crustal basement to

969 the seafloor over more than 15 km along dip (Figures 9b and 9b1; Chao *et al.*,  
970 2021). In the outer part of the necking domain, the basement is affected by a  
971 series of smaller high-angle normal faults, which created tilted blocks with minor  
972 basement exhumation ( $< 0.5$  sTWT fault throw).

973 The oldest sediments reported in the northern margin necking domain (S1) are  
974 interpreted as pre-necking, half-graben wedge-shaped deposits that were  
975 dismembered during the necking phase (Figure 9b1). They include faulted  
976 wedge-shaped remnants capping pre-rift top basement and an allochthonous  
977 package overlying the detachment surface itself (Chao *et al.* 2021; see their  
978 Fig. 6). The first sedimentary package whose base is not disrupted by tectonic  
979 activity is S5, which is thus interpreted as the first post-tectonic package in the  
980 necking domain.

981 In-between S1 and S5, sediments of Ss 2–4 are interpreted as syn-tectonic of  
982 the necking phase. They overlie the entire necking domain and display various  
983 sedimentary architectures: sediments of S2 display a progradational architecture  
984 onto the detachment surface and wedge-shaped architectures between the tilted  
985 blocks of the outer necking domain. Sediments of S3 and S4 passively overlie  
986 those of S2 in the inner part of the necking zone, whereas they are faulted (syn-  
987 tectonic) in the outer part.

988 Chao *et al.* (2021) recognized that the tops of Ss 3, 4 and 5 are truncated,  
989 suggesting that the necking domain was subaerial or above the storm-wave  
990 base during relatively advanced stages of rifting (Figure 9b1). If this assumption  
991 is correct, it implies that the necking domain remained at shallow water depths  
992 during a large part of the rifting process.

993 On the southern margin, the inner part of necking domain is not imaged. Its  
994 outer part is affected high-angle normal faults whose steepness decreases

995 oceanward, ending with a sub-horizontal detachment fault that exhumed upper-  
996 to mid-crustal basement to the seafloor over ca. 15 km along dip (Figures 9b  
997 and 9b2). The detailed tectono-sedimentary record at this margin is poorly  
998 constrained compared to its conjugate because it is not possible to distinguish  
999 between deposits related to Ss 1–4 (Chao *et al.* 2021). However, the boundary  
1000 between S1–4 and S5 is interpreted as erosional, like along the northern margin.

1001 The careful tectono-stratigraphic restoration by Chao *et al.* (2021) allowed them  
1002 to reconstruct the distribution of active faulting through time along the NW-South  
1003 China Sea rifted margins: Figure 9c shows that, during the early stages of rifting  
1004 (necking phase), faulting was localized into two fault corridors on either side of a  
1005 largely undeformed region (active faults shown by red color in Figure 9c). This  
1006 event is interpreted as the individualization of a *keystone* block, as predicted by  
1007 numerical and analogue models (e.g., McClay 1996; Lavier and Manatschal  
1008 2006; Huisman and Beaumont 2011). It ended abruptly with the localization of  
1009 deformation and onset of massive sedimentation within the keystone area,  
1010 contemporaneous with the abandonment of the former fault corridors on either  
1011 side. This rapid switch (< 1 My) was interpreted to mark the transition between  
1012 the necking and hyperextension phase (Chao *et al.* 2021).

1013

#### 1014 5.1.4. The hyperextended domains

1015 In contrast to the necking domains, the hyperextended domains of the northern  
1016 and southern margins are markedly asymmetrical: on the northern margin, the  
1017 hyperextended domain is structured by successive high-angle faults that dip  
1018 oceanward and cut through the entire crust (Figures 9b and 9b1). Faults throw  
1019 decreases oceanward from 1.5 sTWT to < 0.5 sTWT, together with inter-fault  
1020 blocks size. At the southern margin, the hyperextended domain is affected by in-



1021 sequence, oceanward-dipping, low-angle detachment faults that root on one and  
1022 the same décollement level (Figures 9b and 9b2). Based on these observations,  
1023 Chao *et al.* (2021) suggested that the southern margin is a lower plate. In  
1024 contrast, the alternance of ramps and terraces shaping the northern margin tip  
1025 the scales in favor of an upper plate.

1026 On the northern margin, the hyperextended domain can be subdivided into three  
1027 sub-domains with distinct characteristics: in the inner sub-domain (He1 in  
1028 Figure 9b1), the first sediments observed on pre-rift top basement are  
1029 concordant (pre-tectonic) sedimentary deposits interpreted to belong to S3  
1030 (Figure 9b1). The next sediments to be deposited after S3 belong to S4 and  
1031 display a wedge-shaped architecture. These are overlain by passive deposits  
1032 interpreted to belong to S5. In the central sub-domain (He2 in Figure 9b1), tilted  
1033 blocks are overlain by wedge-shaped magmatic products of unknown age.  
1034 Magmatism is however younger than S5 because corresponding sediments  
1035 passively overlie this segment. In the outermost sub-domain (He3 in  
1036 Figure 9b1), only wedge-shaped magmatic products are recognized over the  
1037 pre-rift basement. They are directly overlain by post-rift deposits, suggesting that  
1038 magmatism becomes younger oceanward. This hypothesis is consistent with the  
1039 conclusion drawn by Zhang *et al.* (2021) further east.

1040 The sedimentary record in the hyperextended domain of the southern margin is  
1041 poorly constrained and does not allow us to distinguish between syn-rift Ss 1–4  
1042 (Chao *et al.* 2021). S5 is interpreted as syn-tectonic in the entire hyperextended  
1043 domain (Figure 9b2).

1044 Looking at the spatio-temporal distribution of active faulting (Figure 9c)  
1045 highlights a sudden switch from almost no deformation and no sediment  
1046 deposition onto the keystone block (i.e., the future hyperextended domain), to  
1047 increasingly localized deformation and significant syn-rift sedimentation limited

1048 to the keystone area until the onset of proto-oceanic seafloor spreading.  
1049 Keystone individualization (necking phase) is thus associated with extremely  
1050 shallow water depth and/or emergence of the future hyperextended domain,  
1051 which causes a lack of sediment deposition (uncovered basement shown by  
1052 orange color in Figure 9c).

1053

#### 1054 5.1.5. The proto-oceanic domain:

1055 No exhumed mantle has been evidenced in northwest South China Sea,  
1056 instead, a proto-oceanic domain is described by Larsen *et al.* (2018) and Chao  
1057 *et al.* (2021). This domain is strongly deformed by high-angle normal faults and  
1058 overlain by post-rift sediments (Figure 9b).

1059

#### 1060 5.1.6. Summary and predictions:

1061 When looking at the architecture of the main sedimentary package interfaces,  
1062 we can see that the top of S4 is the base of post-tectonic sedimentation in the  
1063 necking domains, while it lies within the syn-tectonic sequence in the  
1064 hyperextended domains. Similarly, the top of S5 is post-tectonic in the necking  
1065 and inner hyperextended domains, while it belongs to the syn-tectonic sequence  
1066 further outboard. This indicates a migration of active deformation oceanward as  
1067 extension progresses. The same trend can be observed in the study of Ren *et*  
1068 *al.* (2018; their Fig. 9), where their T80 horizon is post-tectonic in the inner part  
1069 of the margin and syn-tectonic in more distal domains.

1070 Figure 9c highlights an abrupt switch in the spatio-temporal distribution of active  
1071 deformation during rifting: deformation is first focused into two fault corridors on  
1072 either side of an undeformed keystone block that remained at shallow water

1073 depth, or even emerged. Then, deformation shifted “instantaneously” (< 1My) to  
1074 an opposite situation where deformation was limited to the keystone and the  
1075 fault corridors became inactive.

1076 The evolution of the NW-South China Sea section described above is  
1077 summarized in Figure 9d. It is consistent with our idealized model in that: (1)  
1078 successive discrete deformation phases advanced oceanward as extension  
1079 progressed; (2) the proximal domain is only structured by high-angle normal  
1080 faults; (3) the necking domain of both margins display detachment faults that  
1081 exhumed upper- to mid-crustal basement to the seafloor; (4) pre-tectonic  
1082 deposits of same age as the necking phase are observed in the hyperextended  
1083 domain of the northern margin; (5) an erosional unconformity of necking phase  
1084 age is observed throughout the hyperextended domain of both margins; and (6)  
1085 extension was achieved via detachment faults in the hyperextended domain of  
1086 the lower-plate (southern) margin.

1087 Relying on our model, we expect sediments of the necking TS (S2–S4 in Chao  
1088 et al., 2021) in the hyperextended domain to display shallow water facies and  
1089 include reworked pre-rift sediments. Deposits on either side of the keystone  
1090 (necking domains) should display deeper sedimentary facies and be sourced  
1091 from both the continent (siliciclastics) and from the eroded keystone (reworked  
1092 pre- to syn-rift sediments).

1093 We expect sediments of the hyperextension TS in both the necking and  
1094 hyperextended domains to display rapidly deepening sedimentary facies,  
1095 associated with both tectonic and thermal subsidence related to intense crustal  
1096 thinning and thermal relaxation beneath the two domains. This hypothesis is  
1097 supported by drill hole U1501 published by Larsen *et al.* (2018) and the tectono-  
1098 stratigraphic analysis of the Liwan Basin further east by Zhang *et al.* (2019),  
1099 both of which indicated a rapid deepening of sedimentary facies from shallow

1100 marine to bathyal depths during the early Oligocene. This time corresponds to  
1101 the early stages of hyperextension in Chao *et al.* (2021).

1102

## 1103 5.2. The Mid-Norwegian margin

### 1104 5.2.1. Geological setting

1105 Rifting between Norway and Greenland was achieved through complex and  
1106 multistage extension that started during the late Paleozoic and resulted in  
1107 lithospheric breakup during the Eocene (Péron-Pinvidic and Osmundsen, 2018,  
1108 and references therein). At the Mid-Norwegian margin, a broadly distributed  
1109 stretching phase took place discontinuously from Permian to Triassic time,  
1110 following the Devonian–Carboniferous orogenic collapse of the Caledonian belt  
1111 (Péron-Pinvidic and Osmundsen, 2018, and references therein). A first phase of  
1112 localized deformation (the necking phase) started during the Late Jurassic  
1113 (Tithonian) and ended in the Early Cretaceous (e.g., Brekke, 2000; Doré *et al.*,  
1114 1999; Planke *et al.*, 2000). Then, localized deformation stepped further  
1115 outboard, creating detachment faults that possibly reached mantle exhumation  
1116 in the Møre and Vøring basins during late Early- to Late Cretaceous  
1117 (Osmundsen *et al.*, 2016). From the Late Cretaceous to final breakup,  
1118 increasing magmatic activity occurred along the outermost margin and formed  
1119 the so-called *outer highs* (Péron-Pinvidic and Osmundsen, 2018, and references  
1120 therein).

1121 The rift domains of the Mid-Norwegian margin were mapped by Péron-Pinvidic  
1122 *et al.* (2013), who placed the termination of the proximal domain at the outer  
1123 edge of the Trøndelag Platform (Figure 10a). In their view, the necking domain  
1124 corresponds largely to the Halten Terrace and continues further south as a  
1125 narrower domain. The Møre and Vøring basins (Figure 10a) belong to the

1126 hyperextended (and possibly exhumed mantle) domain of the mid-Norwegian  
1127 margin because they display a crustal thickness of less than 10 km (Funk *et al.*  
1128 2017). The nature of the seismic basement beneath these basins is the matter  
1129 of much debate, especially about the proportion of pre-rift continental crust with  
1130 respect to magmatic underplate, inherited Caledonian basement, altered  
1131 continental crust and serpentinized mantle (e.g., Osmundsen *et al.*, 2016;  
1132 Péron-Pinvidic and Osmundsen, 2018; Zastrozhnov *et al.*, 2020). The resolution  
1133 is especially low in the outer part of the Møre and Vøring basins due to  
1134 significant breakup-related magmatic additions.

1135 The Norwegian margin benefited recently from a careful three-dimensional  
1136 mapping of major tectonic structures by Osmundsen and Péron-Pinvidic (2018).  
1137 These authors suggested that each rift domain is framed by laterally persistent  
1138 tectonic structures that allow for significant changes in the margin architecture,  
1139 which they termed *breakaway fault complexes*.

1140

#### 1141 5.2.2. The calibrated proximal and necking domains

1142 The tectono-stratigraphy of the Trøndelag Platform was recently reviewed by  
1143 Péron-Pinvidic *et al.* (2020). These authors highlighted a complex extensional  
1144 history following the post-orogenic collapse of the Caledonian belt. Extension  
1145 was accommodated via both normal faulting and extensional reactivation of core  
1146 complex structures inherited from the post-orogenic collapse. The Trøndelag  
1147 Platform displays post-tectonic sedimentation from Late Triassic onward.

1148 The necking domain width ranges between 20 and 80 km (Figure 10a). It is  
1149 framed by two crustal-scale, *breakaway fault complexes*, namely the *inner-* and  
1150 *outer necking breakaway complex* at its inner and outer edge, respectively  
1151 (Figure 10b; Osmundsen and Péron-Pinvidic, 2018). Both necking breakaway

1152 complexes are often composite structures in that they comprise both high- $\beta$   
1153 (large fault heave) detachment faults and low- $\beta$  (minor fault heave) normal  
1154 faults.

1155 The main fault of the inner necking breakaway complex roots within the ductile  
1156 middle or lower crust and is related to an abrupt but moderate increase in  
1157 accommodation (> 20 km of basement exhumation along dip overlain with ca.  
1158 2.5 km of Jurassic and Lower Cretaceous sediments in Figure 10b). From south  
1159 to north along the Norwegian margin, the inner necking breakaway complex is  
1160 represented by the Møre–Trøndelag, Vingleia, Bremstein, and Revfallet Fault  
1161 Complexes (Figure 10a). The main fault of the outer necking breakaway  
1162 complex cuts through the entire crust, down to the upper lithospheric mantle. It  
1163 is usually associated with significant accommodation creation (> 20 km of  
1164 basement exhumation along dip overlain with > 4 km of Lower Cretaceous  
1165 sediments, in addition to the unknown thickness of Jurassic sediments in Figure  
1166 10b). From south to north along the Norwegian margin, it is represented by the  
1167 Main Møre Boundary Fault and the Ytreholmen Fault Complex (Figure 10a).  
1168 However, the three-dimensional architecture of the inner- and outer necking  
1169 breakaway complexes is complicated since the two merge in places, so that the  
1170 necking domain locally coincides with a unique breakaway fault complex. This is  
1171 the case south of the Halten Terrace (i.e., the Klakk Fault Complex; Figure 10a).

1172 The stratigraphy of the Norwegian necking domain often displays erosional  
1173 truncations. For instance, at the outer edge of southern Halten Terrace, the  
1174 Jurassic sequence on the inner side of both the Klakk (outer) and Vingleia  
1175 (inner) Fault Complexes are deeply eroded before the deposition of deep-marine  
1176 Lower Cretaceous sediments (e.g., Bell *et al.*, 2014, and references therein;  
1177 Figure 10c). At the northern Halten Terrace, Breivik *et al.* (2010) and  
1178 Osmundsen and Péron-Pinvidic (2018) interpreted erosion both at the

1179 Ytreholmen (outer) and Revfallet (inner) Fault Complex. In the inner part of the  
1180 Halten Terrace, Elliott *et al.* (2012) highlighted erosion at the Bremstein (inner)  
1181 Fault Complex. South of the Halten Terrace, Osmundsen and Ebbing (2008)  
1182 reported erosion along the Main Møre Fault (inner breakaway fault complex). In  
1183 all cases, erosion is reckoned to be of latest Jurassic–earliest Cretaceous age,  
1184 which corresponds to the end of the necking phase. The resulting unconformity  
1185 is sealed by a regionally correlatable shale event of early Barremian age (Ian  
1186 Sharp personal communication). The latest Jurassic–earliest Cretaceous  
1187 unconformity can be linked to the so-called *Base Cretaceous Unconformity* or  
1188 *BCU* that is widely recognized across the Norwegian margin (e.g., Brekke,  
1189 2000).

1190 Early Jurassic deposits in the necking domain are syn-tectonic sediments  
1191 dominated by shallow-marine sandstones (Faleide *et al.*, 2010). Late Jurassic  
1192 sedimentary facies indicate rapid deepening, as shown by prevailing first  
1193 shallow- and then deep-marine shales. The deepening sequence is interrupted  
1194 by an episode of major uplift and erosion that affected the necking domain but  
1195 not the inner part of the shallow-water proximal domain, dismissing the  
1196 hypothesis of a thermally driven dynamic topography. Indeed, as the latter  
1197 process involves uplift over several hundreds of kilometers (Campbell, 2005),  
1198 dynamic topography in the necking domain would have necessarily generated  
1199 erosion in the adjacent thick-crustal proximal domain, and thus shallow-water  
1200 proximal domain.

1201

### 1202 5.2.3. The poorly calibrated hyperextended domain

1203 In the hyperextended domain offshore the mid-Norwegian margin, the thickness  
1204 of sedimentary deposits may be over 7 sTWT in the axial part of the Møre and

1205 Vøring basins. Drill holes penetrate only as deep as the Late Cretaceous (e.g.,  
1206 Zastrozhnov *et al.*, 2020, and references therein). The base of the Cretaceous  
1207 sequence is interpreted to be marked by a strong reflection, supposed to  
1208 correspond to both an erosional hiatus and to the base of passive onlap in the  
1209 proximal and necking domains (i.e., the *BCU* mentioned above). Underlying  
1210 deposits are usually interpreted as undifferentiated pre-Cretaceous sediments.  
1211 Basement is consensually interpreted as affected by a series of oceanward-  
1212 dipping, in-sequence detachment faults (e.g., Abdelmalak *et al.*, 2017;  
1213 Osmundsen *et al.*, 2002; Zastrozhnov *et al.*, 2020), which suggests that the mid-  
1214 Norwegian margin is a lower-plate margin (e.g., Mosar *et al.*, 2002). This is  
1215 further supported by the shorter along-dip width of the eastern-Greenland  
1216 conjugate margin (e.g., Péron-Pinvidic *et al.*, 2013).

1217

#### 1218 5.2.4. Summary and predictions

1219 The evolution of the Norwegian margin is consistent with that of our idealized  
1220 model in that: (1) it displays a sequential localization of deformation, starting with  
1221 discontinuous Permo–Triassic extension distributed over at least the future  
1222 proximal and necking domains, and probably also over the future hyperextended  
1223 domain. Distributed deformation stopped in the Late Triassic and focused in the  
1224 future necking domain during the Late Jurassic (i.e., necking phase); (2) The  
1225 main structures allowing for crustal thinning (the *inner-* and *outer necking*  
1226 *breakaway complexes* of Osmundsen and Péron-Pinvidic, 2018) are crustal-  
1227 scale detachment faults, possibly associated with minor normal faults; (3) The  
1228 necking phase was often associated with erosion at both the inner-and outer  
1229 necking breakaway complexes; (4) the late stage of the necking phase was  
1230 associated with an erosional unconformity (the *BCU*) that affected both the  
1231 future necking domain, and at least the inner part of the hyperextended domain.



1232 Based on our model, we interpret the BCU either as an unconformity that marks  
1233 the erosion of the keystone block during the necking phase, comparable with the  
1234 unconformity interpreted at the top of the necking TS in the South China Sea; or  
1235 as a onlap surface marking the onset of rapid deepening of the hyperextended  
1236 domain following extreme crustal thinning and/or the onset of thermal relaxation.  
1237 In both cases, we expect that the overlying Early Cretaceous sediments display  
1238 sedimentary facies testifying to a rapid deepening associated with the  
1239 dismembering of the keystone block during hyperextension. This interpretation is  
1240 supported by the massive thickness of Cretaceous deposits (locally > 7 km) in  
1241 both the Møre and Vøring basins (Brekke, 2000; Faleide *et al.*, 2010). Besides,  
1242 the pattern of subsidence–uplift/erosion–subsidence observed on the Halten  
1243 Terrace (necking domain) can be compared to that of the Briançonnais domain  
1244 and explained by a flexural rebound following early necking of the upper mantle  
1245 with respect to the crust (Chenin *et al.*, 2019).

1246

### 1247 5.3. The southeastern margin of the Gulf of Aden: the Socotra region

#### 1248 5.3.1. Geological setting

1249 Socotra Island is located at the eastern edge of the Gulf of Aden southern  
1250 margin. The western part of the island was interpreted to expose part of a former  
1251 necking domain, while its eastern part would correspond to the outer part of a  
1252 proximal domain (Ahmed *et al.*, 2014; Nonn *et al.*, 2017). In between, Socotra  
1253 Island is crossed by an ENE–WSW transfer zone (Hadibo Transfer Zone; HTZ  
1254 on the map of Figure 11). Socotra is one of the rare places in the world where a  
1255 necking domain and the outer part of a proximal domain crop out. Emergence of  
1256 these regions, which usually lie in relatively deep-marine environments, is

1257 related to a late-rift regional uplift (the *syn-OCT uplift* of Leroy *et al.*, 2012, and  
1258 references therein, where *OCT* stands for *Ocean–Continent Transition*).

1259 The oldest sediments recovered in Socotra are Early Triassic and Jurassic in the  
1260 eastern part of the island, and Cretaceous in the west (Samuel *et al.*, 1997).  
1261 These were deposited during a phase of discontinuous, broadly distributed  
1262 extension (stretching phase) that lasted from the Permo-Triassic to the Eocene  
1263 in the future Gulf of Aden region. Localized extension (necking) initiated during  
1264 the Priabonian (ca. 34 Ma; Leroy *et al.*, 2012, and references therein). The  
1265 timing of hyperextension is not constrained. The onset of seafloor spreading  
1266 occurred during the Burdigalian (ca. 20 Ma in eastern Socotra; Fournier *et al.*,  
1267 2007) and was accompanied by regional uplift and significant erosion of both  
1268 conjugate proximal margins (the *syn-OCT uplift* mentioned above).

1269

### 1270 5.3.2. The calibrated (onshore) proximal and inner necking domains

1271 At Socotra, necking was associated with significant tectonic exhumation and  
1272 denudation that removed the entire pre-rift sedimentary cover and part of the  
1273 exposed basement both in the east and in the west of the island (Pik *et al.*,  
1274 2013) We note that this geometry is compatible with basement exhumation  
1275 related to the activity of crustal-scale detachment faults (dotted blue lines on  
1276 transects A–A' and C–C' of Figure 11). This episode of tectonic exhumation  
1277 occurred earlier in the east (ca. 38 Ma) compared to the west (ca. 32 Ma; Pik *et*  
1278 *al.*, 2013).

1279 From late Priabonian to Rupelian time, sediments were deposited in graben-type  
1280 depocenters displaying subsidence and progressive deepening adjacent to the  
1281 detachment faults (Nonn *et al.*, 2017, and references therein). Subsidence  
1282 climax was reached during Rupelian–Chattian time (deep-marine deposits).

1283 Then, a *syn-OCT uplift* brought Socotra Island to emergence (Leroy *et al.*, 2012,  
1284 and references therein).

1285

### 1286 5.3.3. The uncalibrated (offshore) outer necking and hyperextended domains

1287 Offshore Socotra, Nonn *et al.* (2017) recognized usually two crustal-scale  
1288 extensional faults that exhumed basement to the seafloor over several  
1289 kilometers along dip (up to 10 km in transect B–B' in Figure 11). However, in  
1290 easternmost Socotra, they recognized only one major detachment fault that  
1291 exhumed basement over 13 km along dip.

1292 Further outboard, top basement deepens abruptly from ca. 2 km to 4 km below  
1293 sea level over ca. 10 km. This morphostructural step is interpreted to mark the  
1294 transition between the necking and hyperextended domain and can be regarded  
1295 as a mega fault scarp (Nonn *et al.*, 2017; Ribes *et al.*, 2019a). The  
1296 hyperextended domain is structured by a series of oceanward dipping  
1297 detachment faults (transect B–B' and D–D' in Figure 11), which prompted Nonn  
1298 *et al.* (2017) to interpret the Socotra margin as a lower plate. This hypothesis is  
1299 consistent with their interpretation of a residual keystone and of continent-ward-  
1300 dipping faults in the hyperextended domain of the conjugate (Oman) margin.

1301

### 1302 5.3.4. Summary and predictions

1303 The Socotra necking domain provides thermochronological data testifying to a  
1304 phase of necking starting at ca. 34 Ma (Leroy *et al.*, 2012), which was  
1305 accompanied with basement exhumation and emergence (Pik *et al.*, 2013) that  
1306 presumably occurred along detachment faults. Stratigraphy indicates a rapid  
1307 deepening of sedimentary deposits in adjacent depocenters. The climax of

1308 subsidence rate was reached during the Rupelian–Chattian (ca. 28 Ma). Yet,  
1309 according to our model, the highest subsidence rates in the necking domain  
1310 occur during the hyperextension phase. In the case of Socotra, this would imply  
1311 that the necking phase was relatively short (< 34 – 28, namely < 6 My).

1312 The occurrence of a mega-fault scarp at the transition between the necking and  
1313 hyperextended domain of a lower-plate margin is not expected in our idealized  
1314 model. However, whether this mega-fault scarp formed during  
1315 necking/hyperextension or corresponds to a late rift feature associated with the  
1316 syn-OCT uplift of the proximal margin described by Leroy *et al.* (2012, and  
1317 references therein) remains to be determined. That the Socotra margin is not an  
1318 upper plate remains also to be confirmed.

1319 Relying on our model, we expect that the presence of Priabonian (syn-necking)  
1320 and older sediments in the hyperextended domain is limited to the top of the  
1321 tilted blocks, and that they display shallow-water facies and possibly erosional  
1322 unconformities. We expect the hyperextension phase to start during the  
1323 Rupelian–Chattian and to be associated with a rapid increase in water depth,  
1324 which would be recorded by sedimentary facies both on top and between the  
1325 tilted blocks in the hyperextended domain.

1326

## 1327 6. Discussion

### 1328 6.1. Tectono-stratigraphic model vs. reality: first-order trends and simplifications

1329 As a generic template, our tectono-stratigraphic model is necessarily simplified with  
1330 respect to the evolution of rift systems worldwide. Below, we discuss the relevance of  
1331 our simplifications and the reliability of the first-order features we identified as  
1332 characteristic of rift phases or domains.

1333 6.1.1. The stretching phase

1334 In our model, we consider that the stretching phase is achieved via broadly  
1335 distributed, synchronous extension over the entire rift domains. Yet it has been  
1336 proven that, for many rift systems and rifted margins, the stretching phase  
1337 included several extension pulses that were offset in time and/or space (e.g.,  
1338 Alves *et al.*, 2002; Bosworth *et al.*, 2005; Péron-Pinvidic and Osmundsen, 2018;  
1339 Sinclair, 1993). Due to the absence of localized deformation during the  
1340 stretching phase, it is very difficult to distinguish stretching-related structures  
1341 from those generated by orogenic collapse or far-field effects of other tectonic  
1342 events. Conversely, the onset of the subsequent phase of localized deformation  
1343 (i.e., necking) is relatively easy to identify.

1344 Besides, the distribution of deformation during the stretching phase is largely  
1345 controlled by inheritance (Manatschal *et al.*, 2015), and thus specific to each  
1346 tectonic system. As our model aims to provide a template as general as possible  
1347 to interpret the tectono-stratigraphic record of distal margins, it is sensible to limit  
1348 our description of the stretching phase to a very first-order, considering that the  
1349 detailed stretching evolution can only be unraveled on the basis of regional data.

1350

1351 6.1.2. The necking phase

1352 6.1.2.1. Formation of a keystone

1353 The formation of a keystone as part of the necking phase is predicted by  
1354 both analogue and numerical models. It has also been suggested in the  
1355 balanced kinematic restorations from Sutra *et al.* (2013) and Chao *et al.*  
1356 (2021). Several field and seismic studies also provide good evidence for it  
1357 (e.g., Hauptert *et al.*, 2016, and references therein). It is, however,  
1358 extremely difficult to prove the systematic formation of a keystone during

1359 rifting since this phase is transient and the keystone is thought to be  
1360 subsequently dismembered during hyperextension. One exception may be  
1361 the Briançonnais domain which, because of its particular position between  
1362 two overstepping en-échelon rift segments, may have "fossilized" the  
1363 necking stage just prior to the reorganization of the segmented Alpine rift  
1364 system into a unique rift basin, and preserved it during later Alpine  
1365 convergence (Lescoutre and Manatschal, 2020; Manatschal *et al.*, 2021).

1366

#### 1367 6.1.2.2. Formation of tens of kilometer-scale detachment faults

1368 Although only rarely described until the last decade, a growing body of  
1369 evidence suggests the existence of major detachment faults that would  
1370 exhume basement over up to several tens of kilometers along dip (Chao *et*  
1371 *al.*, 2021; Osmundsen and Péron-Pinvidic, 2018; Ribes *et al.*, 2020; Sutra  
1372 and Manatschal, 2012). Reasons why such structures have been seldom  
1373 identified so far may be: (1) the poor calibration of the necking domain  
1374 because it lies in relatively deep-water settings in thermally equilibrated  
1375 rifted margins; (2) our preconceived idea on rifted margins being formed of  
1376 tilted blocks (Figure 12a); and (3) the difficulty to distinguish between the  
1377 top of pre-rift basement and top of new real estate basement (i.e.,  
1378 basement exhumed to the seafloor via an extensional detachment fault).  
1379 Further studies are needed to determine whether such structure represent  
1380 the rule or the exception at rift systems/rifted margins.

1381

#### 1382 6.1.3. The hyperextension phase: formation of upper- vs. lower plate margins

1383 The formation of upper- vs. lower-plate margins at magma-poor rift systems is  
1384 predicted by numerical modelling (Lavie and Manatschal, 2006; Peron-Pinvidic

1385 and Naliboff, 2020) and supported by morphological observations (Hauptert *et*  
1386 *al.*, 2016; Nirrengarten *et al.*, 2016; Péron-Pinvidic *et al.*, 2015; Ranero and  
1387 Pérez-Gussinyé, 2010; Sutra *et al.*, 2013). Note that flip in the polarity of upper-  
1388 vs lower plate margins may occur across transform faults along strike of rift  
1389 systems (Decarlis *et al.*, 2017).

1390 Significant thermal weakening prior to/during the hyperextension phase may  
1391 instead generate a symmetrical evolution (Buck, 1991). However, magma-  
1392 rich/volcanic rift systems are not necessarily symmetrical and may go through a  
1393 comparable early rift evolution as magma-poor ones (e.g., Eldholm *et al.*, 1995;  
1394 Péron-Pinvidic *et al.*, 2013). Thus, whether magma-rich rift systems develop an  
1395 upper- vs. lower-plate architecture remains to be determined.

1396

#### 1397 6.1.4. The mantle exhumation phase: “symmetrical” pattern of detachment faulting

1398 It is generally accepted that mantle exhumation is achieved via extensional  
1399 detachment faulting, however the details of this process are difficult to unravel  
1400 based on field observations and direct seismic imaging only. Several pieces of  
1401 evidence suggest that mantle exhumation is achieved via flip-flop detachment  
1402 faulting, rather than via a unique detachment fault, namely: (1) the fact that  
1403 seafloor spreading initiates usually in the center of an exhumed mantle or proto-  
1404 oceanic domain; (2) the relatively symmetrical pattern of magnetic anomalies  
1405 over exhumed mantle with respect to the final seafloor spreading axis (Gillard *et*  
1406 *al.*, 2016b, and references therein); (3) the symmetrical pattern of over-tilted  
1407 sedimentary sequences in the exhumed mantle domain (Gillard *et al.*, 2016a,  
1408 2015).

1409

## 1410 6.2. Characteristics of rift systems

### 1411 6.2.1. Spatial and temporal scaling of margin domains and rifting events

1412 In Table 1, we summarize the duration of the rift phases, as well as the width of  
1413 the rift domains for the examples presented above. We did not attempt to  
1414 provide duration ranges for the stretching phase because it is a phase of diffuse  
1415 deformation, and thus hard to distinguish from the "noise" generated by potential  
1416 preceding orogenic collapse or far-field effects of adjacent tectonic events. In  
1417 most examples, necking appears to have been achieved rapidly (< 10 My). One  
1418 exception is the Norwegian rift system, for which a late Mid-Jurassic to latest  
1419 Jurassic necking phase (ca. 25 My?; Table 1) is usually considered (Peron-  
1420 Pinvidic and Osmundsen, 2018, and references therein). Yet the Norwegian  
1421 rifting is often described as a succession of extension phases separated by  
1422 tectonic pauses, rather than one continuous polyphase extensional event (see  
1423 Fig. 20 in Zastrozhnov *et al.*, 2020). Short necking phase durations are  
1424 consistent with numerical models by Chenin *et al.* (2020), which explore a wide  
1425 range of crust and mantle rheologies and predict the duration of the necking  
1426 phase to range between 3.3 for strong lithospheres and 8 My for weak  
1427 lithospheres submitted to a bulk extension rate of  $10^{-15} \text{ s}^{-1}$ . The necking domain  
1428 width of all examples above ranges from 20 to 100 km, consistent with the range  
1429 observed by Chenin *et al.* (2017).

1430 Hyperextension also appears to have been achieved rapidly in most examples  
1431 (< 10 My). It may have taken longer at the Norwegian margin, however the  
1432 transition between the phase of hyperextension and a potential phase of mantle  
1433 exhumation is not constrained, which makes it a poorly calibrated example. The  
1434 hyperextended domain widths of all examples above (from 25 to 60 km; Table 1)  
1435 fall within the range observed by Chenin *et al.* (2017). Considering that the  
1436 hyperextended domain is the domain that displays the smaller variability in width  
1437 (Chenin *et al.*, 2017) and is characterized by relatively consistent angles of



1438 aperture between top basement and Moho amongst rifted margins (Nirrengarten  
1439 *et al.*, 2016), we can also expect the hyperextension phase to be relatively  
1440 consistent among rifting events. This hypothesis needs however to be  
1441 confirmed.

1442 The duration of the mantle exhumation phase is poorly constrained from the  
1443 examples presented above; however, we expect it to be highly variable given  
1444 the variability of exhumed mantle domain widths reported by Chenin *et al.*  
1445 (2017). At Socotra, the small width (ca. 10 km) of the exhumation domain  
1446 reported by Nonn *et al.* (2017) falls below the range observed by Chenin *et al.*  
1447 (2017). We can either conclude that the latter range is underestimated because  
1448 mantle exhumation can occur over a very short period/distance; or that mantle  
1449 may not have been exhumed at the Socotra margin, and thus that the domain  
1450 between the hyperextended crust and first mature oceanic crust may rather be  
1451 interpreted as a proto-oceanic domain like in the South China Sea. The latter  
1452 interpretation may be supported by the presence of outer highs, which are  
1453 regarded as magmatic extrusions by Nonn *et al.* (2017). The duration of mantle  
1454 exhumation may be largely controlled by the capacity of underlying  
1455 asthenosphere to produce magma (temperature, fertility, upwelling rate  
1456 conditions), and thus the rapidity of steady-state seafloor spreading to take over,  
1457 however these hypotheses need to be tested.

1458

#### 1459 6.2.2. Spatio-temporal evolution of accommodation and related sedimentary facies

1460 There are two ways to create accommodation, namely vertical deepening and  
1461 horizontal widening. Accommodation space is more or less efficiently and rapidly  
1462 filled with sediments depending on the ratio between sedimentation- vs.  
1463 accommodation creation rate (Tugend *et al.*, 2014). In the following, we describe

1464 the relative contribution of vertical vs horizontal accommodation creation in the  
1465 different rift domains and provide insight into the likely evolution of depositional  
1466 environment. For the sake of simplicity, we consider a sediment-rich, siliciclastic-  
1467 dominated setting like in the NW-South China Sea, for which the tectonic history  
1468 is efficiently captured by the sedimentary record and its relationship to tectonic  
1469 structures (e.g., faults; Figure 12). Sediment-starved rift systems will be  
1470 discussed in section 6.4.1.

#### 1471 6.2.2.1. The proximal domain

1472 The proximal domain is well-calibrated by drilling, seismic imaging, and  
1473 field data. The low heave and throw of normal faults and associated limited  
1474 crustal thinning (continental crust remains usually > 25 km thick; Figure 3a)  
1475 results in limited tectonic accommodation creation (grey curves in Figure  
1476 13a). Besides, minor lithospheric thinning beneath the proximal domain  
1477 accounts for the lack of significant thermal subsidence.

1478 The proximal domain is usually characterized by accommodation-limited  
1479 settings during the entire rifting period, which means an excess of  
1480 sediments with respect to the overall available accommodation space  
1481 (Prosser, 1993). As a result, facies usually range between subaerial and  
1482 shallow-marine *sensu lato*, while a significant part of sediments bypass the  
1483 domain and feed more distal basinal areas. Most common sediments within  
1484 proximal half-graben depocenters include siliciclastic, salt and/or carbonate  
1485 platform deposits. Massive salt deposition may also characterize proximal  
1486 domain depocenters, in particular in arid rift settings (Prosser, 1993).

1487

#### 1488 6.2.2.2. The necking domain

1489 The necking domain is characterized by both moderate horizontal widening  
1490 (up to a few tens kilometers; e.g., Figure 8b2) due to the creation of new  
1491 real estate via extensional detachment faults, and moderate- to high crustal  
1492 (30–10 km) and lithospheric thinning (blue curves in Figure 13a). The  
1493 necking stage is thought to display a largely undeformed keystone block  
1494 between two crustal-scale necking faults (Figure 5c).

1495 The tectono-stratigraphic evolution during the necking phase is complex  
1496 and may vary among rift systems depending on the inherited rheological  
1497 architecture of the involved lithosphere. In the early stages of necking, the  
1498 necking domain remains in shallow water (e.g., Iberia margin; Mohn *et al.*,  
1499 2015, and references therein, mid-Norwegian margin; Faleide *et al.*, 2010,  
1500 and references therein, northern North Sea; McLeod *et al.*, 2002) or may  
1501 even be locally and temporarily uplifted above sea level (e.g., Socotra  
1502 margin; Pik *et al.*, 2013, Campos Basin; Lewis *et al.*, 2014, East-India  
1503 margin; Hauptert *et al.*, 2016, Mont Blanc; Ribes *et al.*, 2020, Porcupine  
1504 Basin; Whiting *et al.*, 2020, NW-South China Sea; Chao *et al.* 2021). At this  
1505 stage, the necking domain is confined and may still be accommodation-  
1506 limited. Sediment deposits may include shallow-marine carbonates and/or  
1507 shallow-marine- or terrestrial sandstones (e.g., Newfoundland margin;  
1508 Wielens *et al.*, 2006, Norwegian margin; Faleide *et al.*, 2010, Porcupine  
1509 Basin; Whiting *et al.*, 2020). Syn-necking sandstones are possibly derived  
1510 from immediately adjacent exhumed basement domes/core complexes  
1511 (e.g., the *Grès Singuliers* close to Mont Blanc; Ribes *et al.*, 2020). They  
1512 may contrast with those deposited during the stretching TS, which are  
1513 usually sourced from more regional sources.

1514 Subsequently, when significant new real estate has been created and/or  
1515 significant thermal subsidence has occurred and/or the flexural rebound of

1516 the underlying lithospheric mantle has stopped, the necking domain may  
1517 rapidly evolve into open-marine and sediment-limited settings (Figure 6c).  
1518 Depositional environments evolve from continental and/or shallow-marine  
1519 during the stretching and early necking TS, to deep marine/continental  
1520 slope environments during the late necking/hyperextension and  
1521 subsequent TS (e.g., shale deposition over the syn-necking shallow-water  
1522 *Grès Singulier* in the Alpine necking domain, Ribes *et al.*, 2020; turbidites  
1523 deposition over syn-necking (Tithonian) shallow-water deposits in the distal  
1524 Iberian margin, Mohn *et al.*, 2015; Upper Jurassic deep-marine shales  
1525 deposited over shallow marine sand-dominated Bathonian deposits and  
1526 Callovian shallow-marine shales in the necking zone of the Norwegian  
1527 margin, Faleide *et al.*, 2010 and Bell *et al.*, 2014, early Cretaceous shales  
1528 in the Newfoundland necking domain, Wielens *et al.*, 2006; sediment  
1529 starvation contemporaneous with strain localization at the Horda Platform  
1530 in the North Sea; Tillmans *et al.*, 2021).

1531 Note that the deposits of the stretching TS are partially dismembered an  
1532 potentially tilted by syn-necking detachment and normal faults, whereas  
1533 deposits of the late necking TS are continuous.

1534

#### 1535 6.2.2.3. The hyperextended domain

1536 The hyperextended domain is characterized by substantial horizontal  
1537 widening, especially on the lower plate, due to the creation of new real  
1538 estate via successive, in-sequence detachment faults (red curves in Figure  
1539 13a; see also Fig. 7 in Manatschal *et al.*, 2001, and Figs. 10c and 10d in  
1540 Epin and Manatschal, 2018). It also displays extreme crustal thinning from  
1541 ca. 10 to > 0 km (Figure 3a), which is accompanied by immediate tectonic

1542 subsidence, and followed by significant thermal subsidence owing to major  
1543 lithosphere thinning.

1544 The hyperextended domains of conjugate margins correspond to the  
1545 former keystone blocks that used to be framed by the two major necking  
1546 faults during the necking phase (Lavier and Manatschal, 2006; Chao *et al.*  
1547 2021). At that time (necking TS), the keystone was at shallow water depths  
1548 or even emerged (cf. previous section; Figure 6c). During the subsequent  
1549 hyperextension phase, it was intensively delaminated via extensional  
1550 detachment faults. As a result, both the stretching and necking TS are  
1551 largely dismembered in the hyperextended domain (Figure 6b). From the  
1552 late necking or early hyperextension TS onward, sediment deposits testify  
1553 to a switch from an accommodation-limited to a sediment-limited  
1554 environment, which is expressed by a rapid deepening of sedimentary  
1555 facies up to deep-marine or abyssal conditions (Figure 6c; e.g., Masini *et*  
1556 *al.*, 2011; Mohn *et al.*, 2015). The available accommodation during  
1557 subsequent rift stages is thus essentially inherited from the unfilled  
1558 accommodation created during hyperextension, and to a lesser extent due  
1559 to lithosphere thermal relaxation.

1560 The sediments of the late hyperextension TS are the first to be continuous  
1561 along dip of the hyperextended domain.

1562

#### 1563 6.2.2.4. The exhumed mantle domain

1564 The exhumed mantle domain is characterized by extreme horizontal  
1565 widening since this domain is entirely made of new real estate. When the  
1566 distal rift basin is not isolated, mantle is exhumed in deep-marine or  
1567 abyssal settings (e.g., Epin *et al.*, 2021; Steinmann, 1905) and is

1568 subjected to significant thermal subsidence once it has moved away from  
1569 the spreading center.

1570 The first sediments deposited onto the exhumed mantle belong to the  
1571 mantle exhumation TS (Figure 6a). Siliciclastic sedimentary sources may  
1572 be lacking since the exhumed mantle domain is usually more than 100 km  
1573 away from the emerged continent (Chenin *et al.*, 2017). Instead, deposition  
1574 may be dominated by axial sedimentation (i.e., perpendicular to the margin  
1575 dip-direction). Deposits may either arise from local gravitational reworking,  
1576 turbidity or contour currents, pelagic sedimentation, or chemical  
1577 precipitation (e.g., oversaturated SiO<sub>2</sub> rich fluids precipitating cherts), all of  
1578 which usually characterize abyssal depositional environments (below the  
1579 Carbonate Compensation Depth or CCD). Exceptions may occur at  
1580 transform margins, where the exhumed mantle domain may be directly fed  
1581 by the adjacent continent.

1582

### 1583 6.3. Projections and correlations from proximal to distal domains

#### 1584 6.3.1. New aspects of the interpretation of distal domains

1585 Due to the significantly greater depth of the distal margin with respect to the  
1586 proximal domain, the necking zone has for a long time represented an  
1587 “interpretation frontier” between a well-calibrated domain continent-ward, and a  
1588 much more enigmatic and largely inaccessible domain oceanward (Doran and  
1589 Manatschal, 2017). The first attempts of distal margin interpretation naturally  
1590 considered that observations made at the proximal domain can be transposed to  
1591 the distal one (Figure 12a). Therefore, rift-related deformation was regarded as  
1592 synchronous along the entire margin and achieved via planar high angle normal  
1593 faulting between rigid tilted blocks (Waltham *et al.*, 1993). The understanding of

1594 where, when and how deformation and sediments are distributed during rifting  
1595 underwent a significant paradigm shift since the 1990's (Figure 12b). It is now  
1596 generally acknowledged that rifting is a localizing process that focuses toward  
1597 the locus of future lithospheric breakup (Pérez-Gussinyé *et al.*, 2020; Ribes *et*  
1598 *al.*, 2019b; Whiting *et al.*, 2020; Luo *et al.*, 2021; Chao *et al.* 2021), and thus the  
1599 base of post-tectonic deposits should be regarded as younging oceanward.  
1600 Practically, (1) syn-tectonic sediments of the distal domain are likely to  
1601 correspond (timewise) to post-tectonic deposits in the proximal domain; and (2)  
1602 sediment deposits of early rift phases are going to be increasingly fragmented  
1603 oceanward (Figure 12b).

1604

### 1605 6.3.2. Correlating timelines across rifted margins

1606 In our tectono-stratigraphic model, we associated each of the successive  
1607 deformation phases at play during rifting to one Tectonic Sequence (TS). A  
1608 Tectonic Sequence encompasses all stratigraphic packages synchronous to the  
1609 corresponding deformation phase, therefore the base and the top of each TS  
1610 are by definition timelines along-dip of a given transect. Due to the polyphase  
1611 and localizing nature of rifting, their correlation across the margin is difficult.  
1612 Below we provide a set of characteristic timelines that may be generally  
1613 recognized at rifted margins and discuss what the formerly defined *Rift Onset*  
1614 *Unconformity* and *Breakup Unconformity* (Falvey, 1974) and *Necking*  
1615 *Unconformity* (Chenin *et al.*, 2015) become in our new model.

#### 1616 6.3.2.1. The Base of the Syn-Rift megasequence vs. the Rift Onset

##### 1617 Unconformity

1618 In the mid-1970's, the onset of rifting was thought to be immediately  
1619 preceded by a phase of large-scale (several hundreds of km)

1620 doming/arching of the continental crust that produces widespread erosion  
1621 in the future region of rifting and breakup (Falvey, 1974). By the late  
1622 1970's, such a phase of thermal doming/dynamic topography was thought  
1623 to be characteristic of the onset of *active/volcanic rifting* only, while  
1624 *passive/magma-poor rifting* was supposed to start with subsidence (Sengör  
1625 and Burke, 1978), as predicted by the Pure-Shear Model of McKenzie  
1626 (1978). More recently, Menzies *et al.* (2002) concluded that pre-rift dynamic  
1627 topography was not systematic enough to be diagnostic of volcanic rifting.  
1628 Dynamic topography produces a major unconformity over a wide domain  
1629 (> 600 km; Campbell, 2005; Figure 13b). This unconformity can be used as  
1630 a proxy for the Base Syn-Rift timeline (BSR in Figures 6, 12a and 12b)  
1631 when the related doming precedes the stretching phase.

1632 Whether or not initial doming takes place, the stretching phase is  
1633 characterized by widely distributed normal faulting that forms numerous  
1634 half-graben depocenters. In sediment-rich rift systems, the BSR  
1635 corresponds within the basins to a shift from pre-tectonic to half-graben  
1636 wedge-shaped deposits. Such a shift may be identifiable on both seismic  
1637 sections and at the outcrop scale.

1638 The formation of high-angle normal faults is often associated with the uplift  
1639 of the corresponding footwall and local erosion of its crest (Yielding, 1990)  
1640 over a width usually less than 5 km (Figure 13b). This erosional  
1641 unconformity and its correlative conformities can be used as a proxy for the  
1642 age of the BSR.

1643

1644 6.3.2.2. The Base of the Necking Tectonic Sequence vs. the Necking

1645 Unconformity



1646 Chenin *et al.* (2015) defined the necking unconformity as the transition from  
1647 syn- to post-tectonic deposits in the proximal domain (for sufficiently  
1648 sediment-rich rift systems). In our model this “unconformity” corresponds to  
1649 the timeline at the base of the necking TS (BNk in Figures 6 and 12a and  
1650 12b).

1651 The “necking unconformity” is difficult to correlate across more distal  
1652 domains. In the future necking and hyperextended domains, we notice that  
1653 the necking phase is often associated with the formation of erosional  
1654 unconformities. In the necking domain, erosion occurs at the breakaway of  
1655 the necking fault (e.g., Norwegian margin; Bell *et al.*, 2014) and/or along  
1656 upper- to mid-crustal basement exhumed by this fault (e.g., Mont Blanc;  
1657 Ribes *et al.*, 2020). The new real estate created by activity on detachment  
1658 faults corresponds also to a discordance since sedimentary deposits  
1659 cannot be older than local basement exhumation age. Distinguishing  
1660 between exhumed basement and eroded top basement may be difficult,  
1661 especially because one does not exclude the other.

1662 In-between the main necking fault systems, the keystone block (future  
1663 hyperextended domains) seems to remain at shallow water depths during  
1664 (part of) the necking phase, or to be even temporarily emergent. Illustrative  
1665 examples include the early Jurassic karst of the Briançonnais (section  
1666 4.2.2; Claudel and Dumont, 1999), Tithonian offshore shelf deposits at the  
1667 Iberian hyperextended domain (section 4.3.3; Mohn *et al.*, 2015, and  
1668 references therein), lack of deposition in the central part of the  
1669 northwestern South China sea (section 5.1.4 and Figure 9e; Chao *et al.*  
1670 *subm.*), shallow water deposits on the Halten Terrace at the mid-Norwegian  
1671 margin (section 5.2.2; Bell *et al.*, 2014), and basement subaerial  
1672 exhumation at the Socotra margin (section 5.3.2; Pik *et al.*, 2013). The

1673 width of such necking-related uplift is of the order of a few tens of  
1674 kilometers (Figure 13b; Chenin *et al.*, 2019, and references therein).

1675 Thus, the necking phase may be associated with several unconformities  
1676 and/or discordances in both the necking and hyperextended domains.  
1677 Although these may not be synchronous, they are largely limited to the  
1678 necking TS, which seems to usually encompass only a short period of time  
1679 (section 6.2.1), and thus can be reasonably correlated with the “necking  
1680 unconformity” of the proximal domain.

1681

### 1682 6.3.2.3. The Base of the Post-Rift megasequence vs. the Breakup

#### 1683 Unconformity

1684 Falvey (1974) suggested that continental breakup results in a rapid  
1685 increase in the subsidence rate at the two conjugate margins. The resulting  
1686 rapid marine transgression produces either a backstep of the onlapping  
1687 sediments towards the continent when accommodation creation rate  
1688 dominates over sedimentation rate at the coast, or a progressive onlap  
1689 otherwise. In both cases it would produce an unconformity, often named  
1690 the *Breakup Unconformity*. This hypothesis was supported by dynamic  
1691 thermo-mechanical modelling (Braun and Beaumont, 1989). Although the  
1692 existence of such discordance has been highlighted in several margins  
1693 (e.g., Iberia–Newfoundland; Tucholke *et al.*, 2007, Santos Basin, Grand  
1694 Banks and Beaufort Sea; Hubbard, 1988, Amerasian Basin; Embry and  
1695 Dixon, 1990), its cause and synchronicity became debated from the 2000’s  
1696 (e.g., Masini *et al.*, 2013; Wilson *et al.*, 2001). Gillard *et al.* (2015)  
1697 highlighted that, in the exhumed mantle domain, each detachment fault  
1698 was associated with a specific discordance, ruling out the existence of a  
1699 unique *breakup unconformity* at the scale of an entire margin. More

1700 recently, Pérez-Gussinyé *et al.* (2020) advocated, based on dynamical  
1701 thermo-mechanical numerical modeling, that the “breakup unconformity”  
1702 was restricted to the most distal part of the margin because lithospheric  
1703 breakup is not associated with a sudden stress drop.

1704 The Base of the Post-rift timeline (BPR in Figures 6, 12a and 12b) and  
1705 onset of seafloor spreading are often difficult to define, either because of  
1706 scattered and sporadic incipient magmatic activity in magma-poor rift  
1707 systems (Bronner *et al.*, 2011; Epin *et al.*, 2019; Epin *et al.*, 2021; Jagoutz  
1708 *et al.*, 2007), or because of excess magma prior to onset of steady-state  
1709 seafloor spreading in magma-rich rift systems (Tugend *et al.*, 2020). When  
1710 the first-formed oceanic crust can be identified, the age of breakup  
1711 corresponds to the age of the first overlying sediments in most sediment-  
1712 rich rift systems (in sediment-starved systems the first overlying sediments  
1713 provide only a minimum age). On seismic sections, the corresponding  
1714 seismic reflection could theoretically be correlated across the entire margin.  
1715 It may correspond to a discordance in the most distal part of the margin  
1716 since it expresses the transition from syn- to post tectonic settings,  
1717 however it is likely a conformable reflection in the proximal domain since  
1718 deformation therein ceased earlier than the time of breakup. In the field, the  
1719 BPR is practically impossible to identify based on geometrical criteria. In  
1720 the past, it was considered to be marked by the biostratigraphic hiatus  
1721 related to the backstep/progressive onlap mentioned above. However,  
1722 given that breakup is not associated with a sudden stress drop in the  
1723 proximal domain (Pérez-Gussinyé *et al.*, 2020), the contemporality  
1724 between this paraconformity recorded in the proximal domain and the BPR  
1725 is unlikely.

1726 It is important not to automatically interpret unconformities at proximal  
1727 margins as marking the BPR because numerous unconformities in the  
1728 proximal domain may relate either to local tectonic events, or to large-  
1729 scale, unrelated processes (e.g., climate, or far-field influence of a large-  
1730 scale tectonic event). Indeed, as the top of the proximal domain remains at  
1731 shallow depths due to its thick crust ( $30 \pm 5$  km) and the absence of  
1732 significant lithospheric thinning, it may be very sensitive to eustatic  
1733 changes (Prosser, 1993). Therefore, the timing of breakup cannot be  
1734 accurately determined from seismic reflector relationship the proximal and  
1735 necking domains.

1736

#### 1737 6.4. Model limitations

1738 The tectono-stratigraphic evolution presented in this contribution provides a powerful  
1739 template to guide the seismic interpretation of distal margins and to unravel the  
1740 complexity of rift systems exposed in orogens, where they are usually dismembered  
1741 by thrust tectonics. However, as a model it remains a simplification, and thus it  
1742 should not be used without taking the regional specificities of individual rift systems.  
1743 Below we briefly review the main potential pitfalls to keep in mind when using this  
1744 template:

##### 1745 6.4.1. Completeness and quality of the tectono-stratigraphic record

1746 The efficient record of the tectono-stratigraphic evolution depends primarily on  
1747 the ratio between sedimentation- and accommodation creation rate (e.g.,  
1748 Ravnås and Steel, 1998 and references therein; Martins-Neto and Catuneanu,  
1749 2010). A sediment-starved system will incompletely or even not record syn-  
1750 tectonic deformation (Figure 14a; e.g., Barr, 1991). Note that the sedimentary  
1751 budget available during rifting depends largely on the climate, in that humid

1752 conditions favor increased erosion, and hence prolific sediment sources,  
1753 conversely to arid climates that favor sediment starvation (Prosser, 1993).

1754 The presence of pre- or syn-rift low frictional materials like salt and/or clay may  
1755 also be a problem since they can induce deformation and accommodation  
1756 creation not necessarily related to the rifting event itself (e.g., Rowan and Ratliff,  
1757 2012).

1758

#### 1759 6.4.2. Presence of magmatic systems

1760 The presence of magmatic systems may represent a difficulty because  
1761 (1) magmatic crust may be hard to distinguish from pre-rift continental crust  
1762 (Osmundsen et al., 2016; Tugend et al., 2020); (2) magmatic intrusions may  
1763 perturb or overprint the tectono-sedimentary record, rendering its unravelling  
1764 difficult (Zhang et al., 2021); and (3) magma-related heat may modify the  
1765 rheology of the lithosphere involved, and thus lead to different deformation  
1766 modes compared to magma-poor rifting; in particular, significant magmatic  
1767 activity and related diking may trigger both crustal and lithospheric breakup  
1768 before the continental crust was thinned to 0 km (Bialas et al., 2010; Ebinger  
1769 and Casey, 2001; Sun et al., 2019). Otherwise, magmatic lava flows may  
1770 behave comparably to sediments, and can thus be interpreted in a similar way  
1771 (e.g., Chao et al., 2021)

1772

#### 1773 6.4.3. The three-dimensional aspect of rift systems

1774 Rift systems are often segmented by transform faults (Figure 14b1), especially  
1775 during crustal thinning (necking and hyperextension; Nirrengarten *et al.*, 2018).  
1776 Transform systems can separate segments with distinct margin architectures, as

1777 exemplified by the differing domains width between the northern and central  
1778 segment of the Iberia margin (Sutra and Manatschal, 2012) and between the  
1779 southern (Møre), central (Vøring) and northern (Lofoten) segment of the mid-  
1780 Norwegian margin (Péron-Pinvidic and Osmundsen, 2018). Transfer faults can  
1781 also delimit segments with different timings of extension (e.g., the sequential  
1782 northward propagation of crustal extension along the Iberian margin; Alves *et*  
1783 *al.*, 2009, the sequential evolution along the Norwegian–Greenland conjugate  
1784 margins; Gernigon *et al.*, 2020, or the westward younging of extension across  
1785 the transform fault of the Socotra margin; Pik *et al.*, 2013). Therefore, the timing  
1786 of necking, hyperextension and mantle exhumation, as well as the age of the  
1787 various unconformities mentioned above, vary presumably from one segment to  
1788 another (Nirrengarten *et al.*, 2018; Péron-Pinvidic *et al.*, 2013).

1789 Individual faults have a limited extent along strike (Lescoutre and Manatschal,  
1790 2020; Lymer *et al.*, 2019; Nonn *et al.*, 2017; Péron-Pinvidic *et al.*, 2007) and  
1791 major structures may change their tectonic role along the margin (e.g., local  
1792 merging of breakaway complexes in Osmundsen and Péron-Pinvidic, 2018;  
1793 Figure 10a).

1794 The correlation of timelines is particularly tricky in propagating rift systems (e.g.,  
1795 the V-shaped South China Sea, Gulf of Aden, Bay of Biscay, Porcupine Basin,  
1796 early stages of the southern North-Atlantic rifting, etc.; Nirrengarten *et al.*, 2018,  
1797 and references therein; Luo *et al.*, 2021). Indeed, the migration of deformation  
1798 toward both the rift center and the tip of the propagator renders timelines oblique  
1799 to the conjugate margins (Figure 14b2).

1800

1801 6.4.4. Multistage rifting (out-of-sequence deformation)

1802 Rifting may not always initiate on a thermally and gravitationally equilibrated  
1803 lithosphere (Manatschal *et al.*, 2021; Zhang *et al.*, 2020) and/or the pre-rift  
1804 sediments may display variable thicknesses inherited from previous tectonic  
1805 events. In Western Europe, for instance, the Alpine Tethys rifting overprinted an  
1806 area that was formerly affected by the Meliata-Vardar rifting, which is expressed  
1807 by a thickening of the pre-Alpine Tethys rifting sedimentary sequence to the east  
1808 (Manatschal *et al.*, this vol. and references therein). In Grand Banks, the Jeanne  
1809 d’Arc stretching basin underwent post-Iberia–Newfoundland rifting extension  
1810 related to the opening of the Orphan Basin (e.g., Skogseid, 2010). We refer to  
1811 these spatially overprinting but temporally distinct extension phases as  
1812 *multistage rifting* (Cadenas *et al.*, 2020). The consequence of multistage rifting  
1813 may be the local incorporation and/or reactivation of older tectono-stratigraphic  
1814 features into the ongoing rift system (Figure 14c). Examples include the Permian  
1815 and Triassic basins incorporated in the Jurassic hyperextended domain of the  
1816 Alpine rift system (Epin and Manatschal, 2018); and Jurassic salt-bearing basins  
1817 in the Cretaceous hyperextended domain of the Basque-Cantabrian basin  
1818 (Manatschal *et al.*, 2021; Miró *et al.*, 2021).

1819

## 1820 7. Conclusion

1821

1822 In this contribution, we propose a tectono-stratigraphic model describing the primary  
1823 spatio-temporal evolution of active deformation during rifting. We show that the different  
1824 rift domains of rifted margins are associated with specific (1) periods of tectonic activity;  
1825 (2) styles of deformation and related tectono-stratigraphic architecture; (3) amounts of  
1826 horizontal and vertical accommodation creation; and (4) depositional environments. We  
1827 demonstrate that the sequential localization of extension during rifting results in both a  
1828 younging of the base of passive infill and an increasing fragmentation of early rift-related

1829 deposits oceanward. We propose a panel of Tectonic Sequences that may be correlated  
1830 more or less easily and continuously across rifted margins. This model provides a new  
1831 template to help correlate observations made at the well-calibrated and understood  
1832 proximal domain with the more inaccessible and enigmatic distal domain. It highlights the  
1833 necessity of jettisoning outdated dogma on rifted margins and changing our routines  
1834 when interpreting seismic sections and correlating remnants of dismembered rifted  
1835 margins exposed in orogens.

1836

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1843

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1845 Pauline Chenin: Writing – original draft, Conceptualization, Investigation

1846 Gianreto Manatschal: Writing – review & editing, Conceptualization, Investigation, Funding  
1847 acquisition

1848 Jean-François Ghienne: Writing – review & editing, Conceptualization, Investigation

1849 Peng Chao: Writing – review & editing, Conceptualization, Investigation

1850

1851 Data Availability Statement:

1852 Data sharing is not applicable to this article as no new data were created or analyzed in this  
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TABLE CAPTION

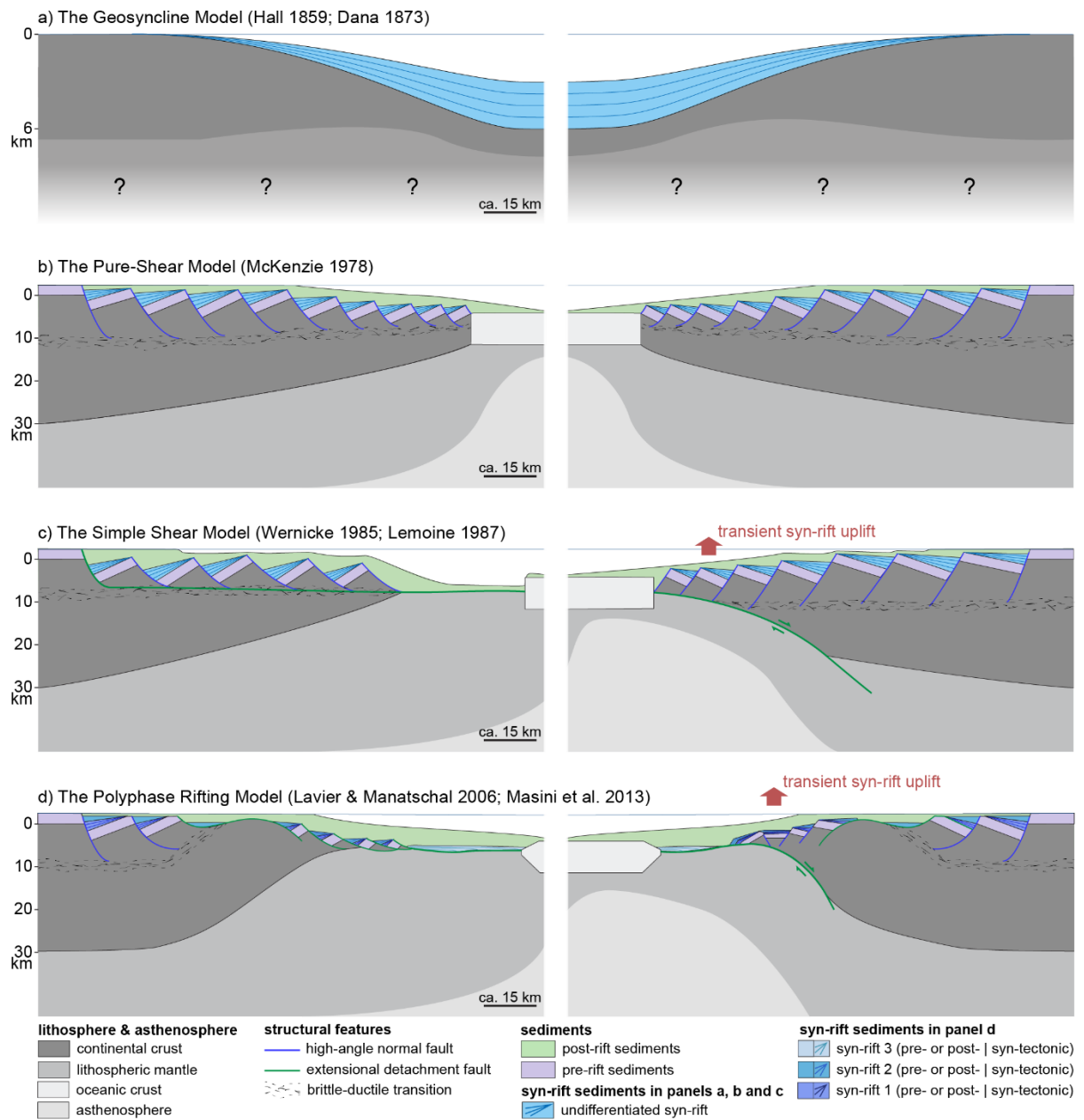
	Necking		Hyperextension		Mantle exhumation		Proto-oceanic				
	phase duration	domain width	phase duration	domain width	phase duration	domain width	Phase duration	domain width			
Alps	6 My (Ribes et al. 2019)	ca. 10 My (Manatschal et al. this vol.)	unknown	5 My (Ribes et al. 2019)	ca. 10 My (Manatschal et al. this vol.)	unknown	18 My (Ribes et al. 2019)	ca. 10 My (Manatschal et al. this vol.)	unknown	X	unknown
Iberia–Newfnd.	ca. 10 My (Mohn et al. 2015, with ref.)	ca. 100 (E) & 45 (W) km (Sutra et al. 2013)	ca. 5 My (Mohn et al. 2015, with ref.)	ca. 60 & 45 km (Sutra et al. 2013)	ca. 25 My (Mohn et al. 2015, with ref.)	ca. 80 (E) & 90(W) km (Sutra et al. 2013)			X		X
NW SCS	7 My (Chao et al. subm.)	N margin: ca. 60 km (Chao et al. Subm.)	4 My (Chao et al. Subm.)	ca. 25 (N) & 30 (S) km (Chao et al. Subm.)			X	X	2 My (Chao et al. subm)		X
Norway	Late Mid-Jurassic–base Cretaceous (ca. 25 My? Péron-Pinvidic et al. 2018, with ref.)	20–80 km (Osmundsen & Péron-Pinvidic 2018)	Base Cretaceous–Late Cretaceous (ca. 60 My? Péron- Pinvidic et al. 2018, with ref.)	unknown	unknown	unknown	unknown	unknown	Late Cretaceous–Paleocene (ca. 30 My? Péron-Pinvidic et al. 2018, with ref.)		unknown
Socotra	< 6 My? (This study)	unknown	ca. 10 My? (This study)	20–30 km (Nonn et al. 2017)	unknown	ca. 10 km (Nonn et al. 2017)			X		X

**Table 1:** Summary of the duration of the different rift phases and width of the different rift domains at the rift systems presented in this study.

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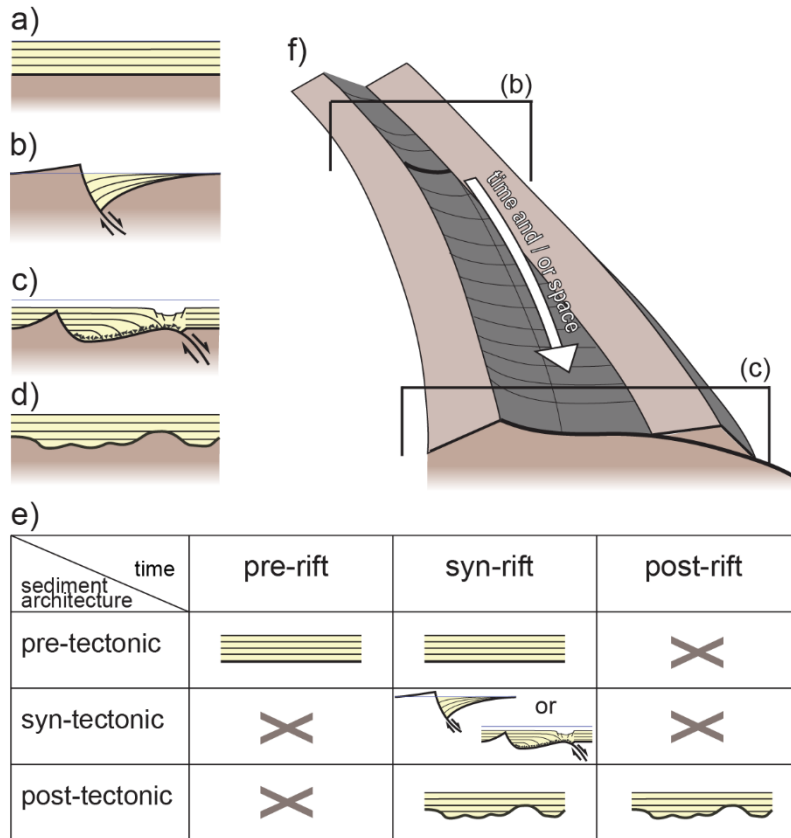
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FIGURE CAPTIONS



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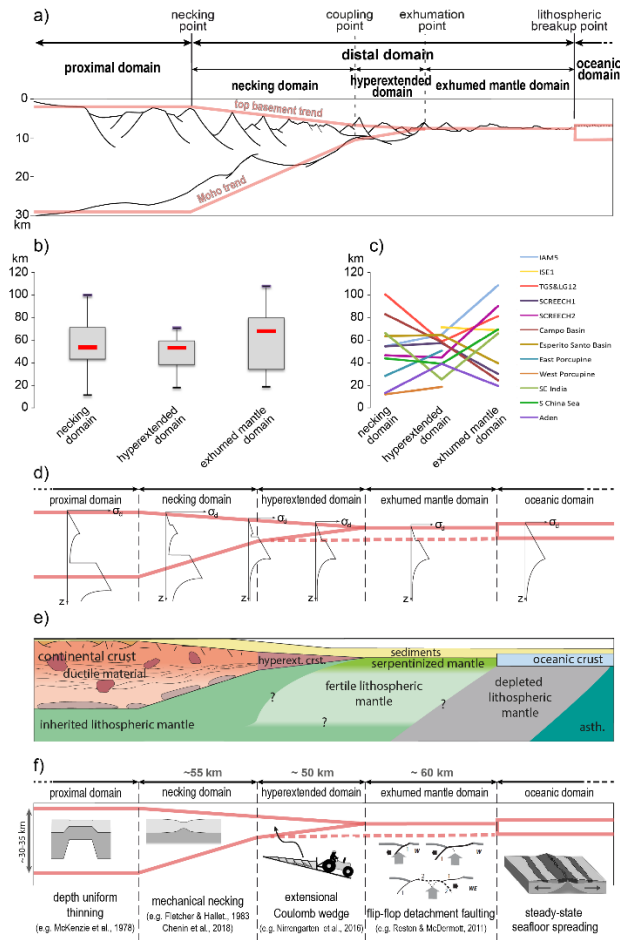
**Figure 1:** Schematic illustration of the evolution of stratigraphic models for rifted margins: a) the Geosyncline Model (Dana, 1873; Hall, 1859); b) the Pure-Shear or Depth-Uniform Thinning Model by McKenzie (1978); c) the Simple-Shear Model by Wernicke (1985); and d) the Polyphase Stretching–Thinning–Exhumation Model by Lavier and Manatschal (2006). Note that the scale for sedimentary deposits is vertically exaggerated compared to that of the crust.



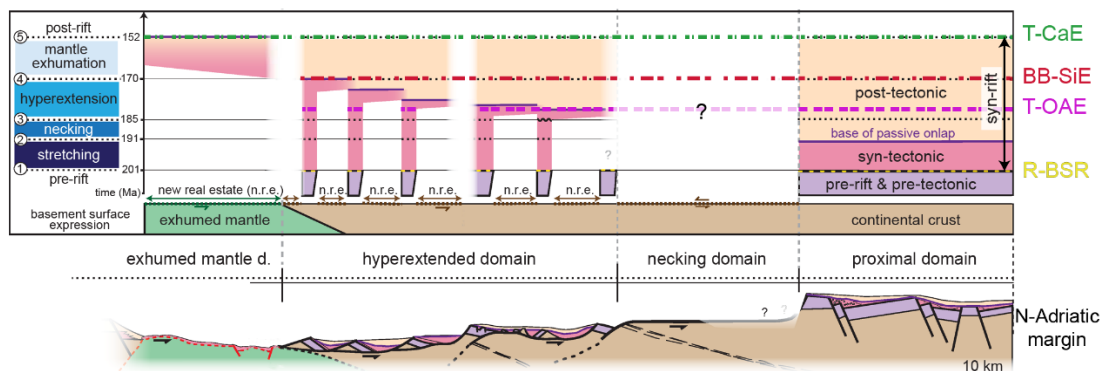
**Figure 2:** Schematic stratigraphic architectures depending on the tectonic context during sediment deposition: a) pre-tectonic; b) syn-tectonic half-graben wedge-shaped; c) syn-tectonic supra-detachment wedge-shaped; d) post-tectonic. e) relationships between the pre-, syn-, post-rift time intervals and the pre-, syn-, post-tectonic sedimentary architectures; f) schematic illustration of the evolution of a high-angle normal fault into a low-angle extensional detachment fault as extension progresses (modified from Péron-Pinvidic et al., 2007). Note that all schematics represent idealized architectures deposited in sediment-rich systems.

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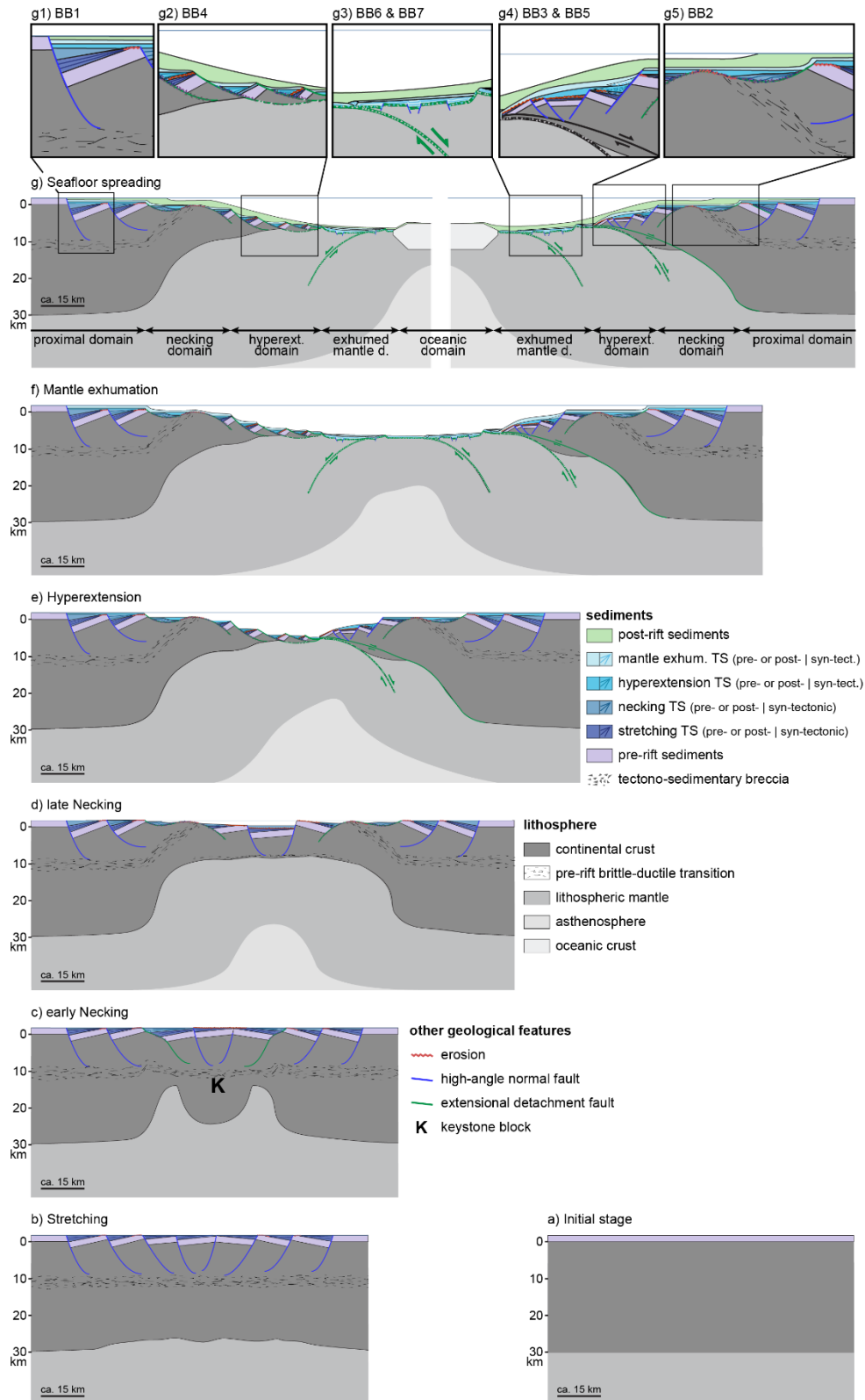
**Figure 3:** a) Schematic cross section highlighting the primary morphology of magma-poor rifted margins (from Chenin *et al.*, 2017); b) Width of the different distal domains in magma-poor rifted margins (from Chenin *et al.*, 2017); c) Width relationships among the different distal domains in magma-poor rifted margins (from Chenin *et al.*, 2017); d) First-order rheological evolution across a typical magma-poor rifted margin represented by schematic depth-dependent stress profiles;  $\sigma_d$ : deviatoric stress;  $z$ : depth; e) Schematic cross section displaying the first-order lithology across a typical magma-poor rifted margin (modified from Chenin *et al.*, 2017). f) Schematic cross section displaying the dominant deformation mechanism associate with the different rift domains of magma-poor rifted margins.



**Figure 4:** Wheeler diagram representing the spatio-temporal distribution of pre-, syn- and post-tectonic sedimentary sequences across the Adriatic margin of the Alpine Tethys (modified from Ribes *et al.*, 2019b). The dotted lines in the Wheeler diagram represent timelines that correspond to dated global or local events, namely: BB-SiE: Bajocian-Bathonian bio-Siliceous Event; R-BSR: Rhaetian Base of the Syn-rift sequence; T-CaE Tithonian Carbonate Event; T-OAE: Toarcian Oceanic Anoxic Event. Abbreviation n.r.e. stands for *new real estate*.

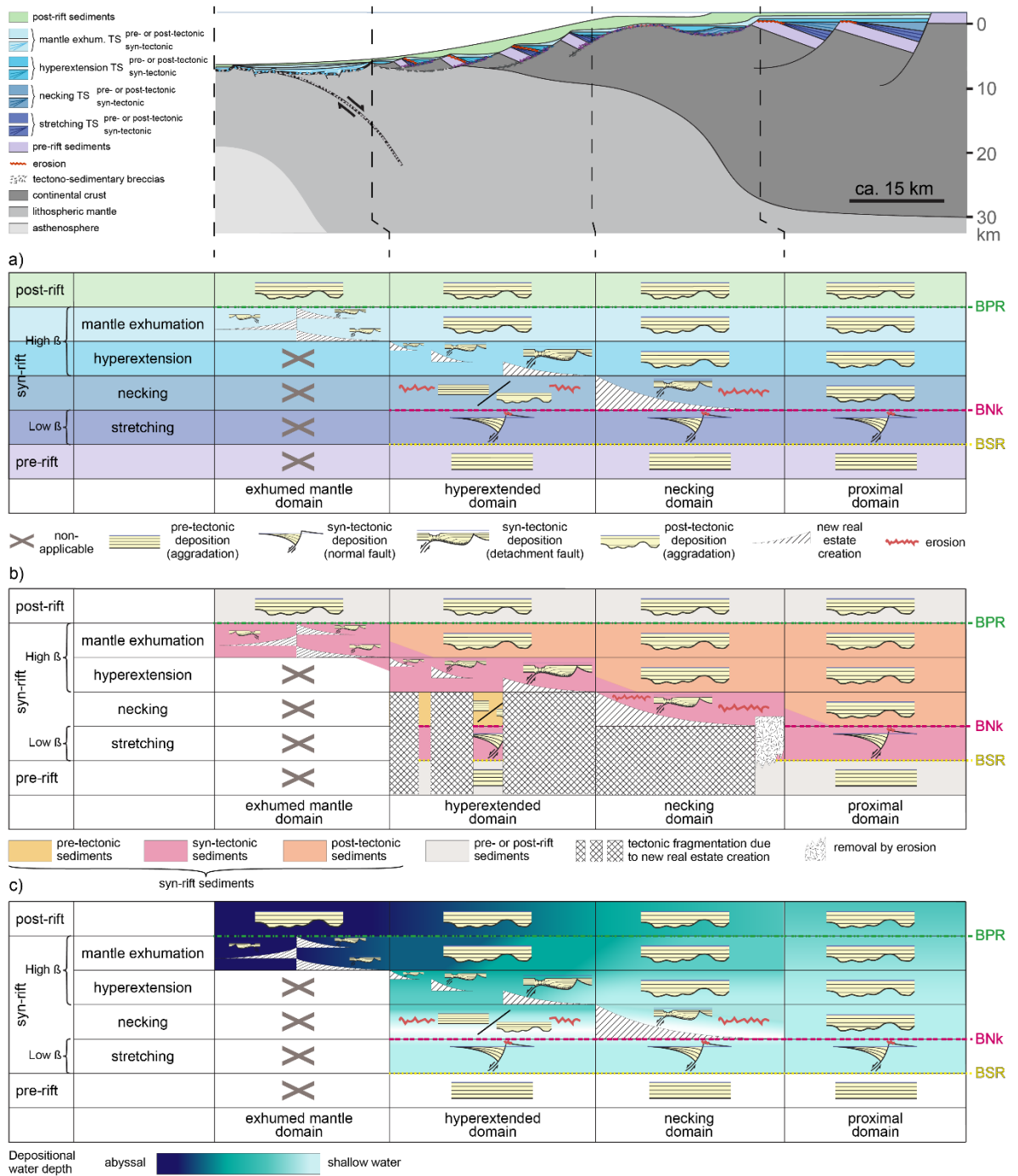
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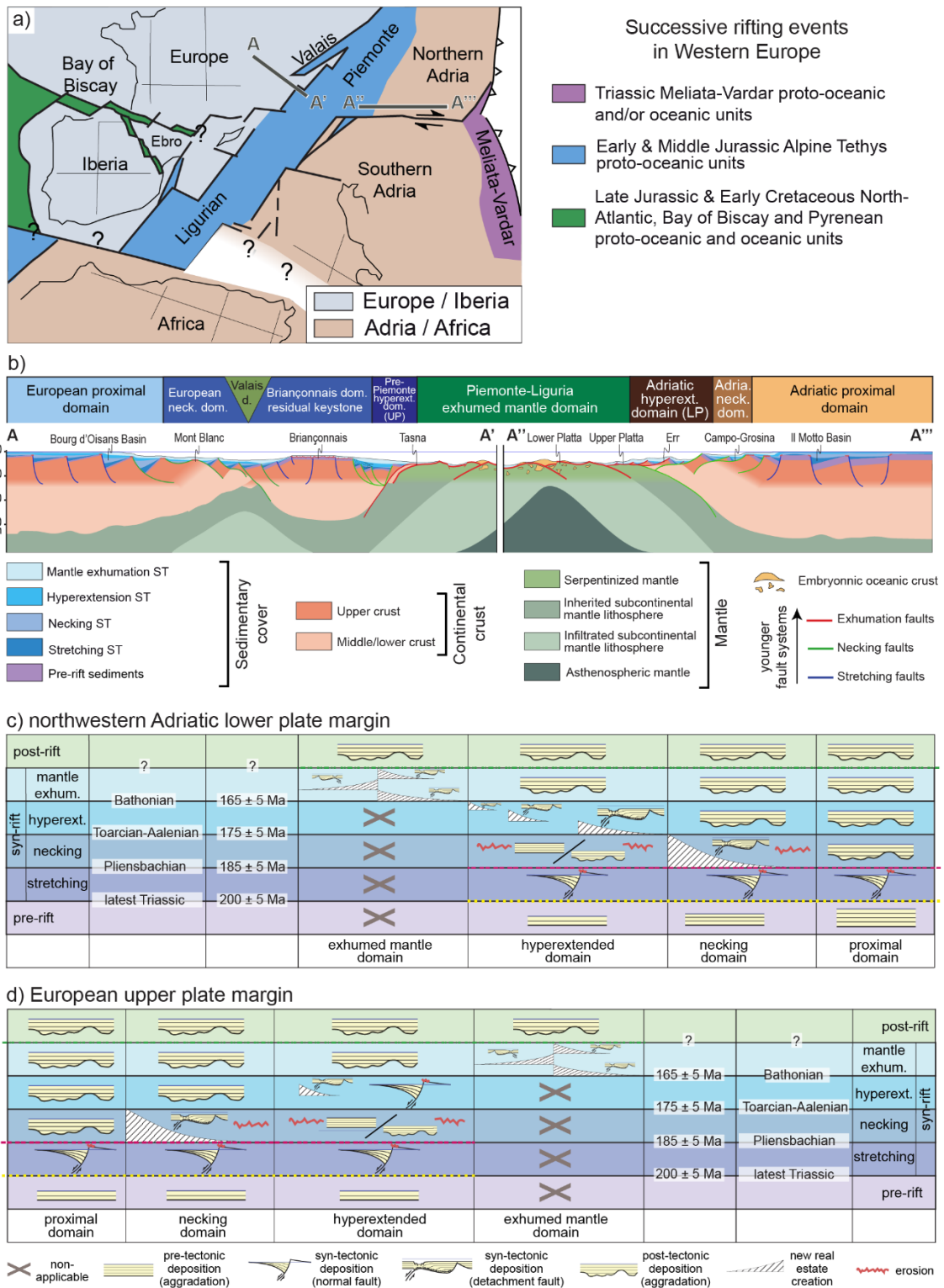


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**Figure 5:** Schematic evolution of an idealized magma-poor rifting (crustal surface and pre-rift sediment length are conserved among the different stages): a) pre-rift setting; b) stretching phase; c) early necking phase; d) late necking phase; e) hyperextension phase; f) mantle exhumation phase g) seafloor spreading phase. Insets g1-g5 zoom on characteristic “Building Blocks” (BB1-BB7) of rifted margins (see Manatschal *et al.*, this volume, for a description of the different Building Blocks).



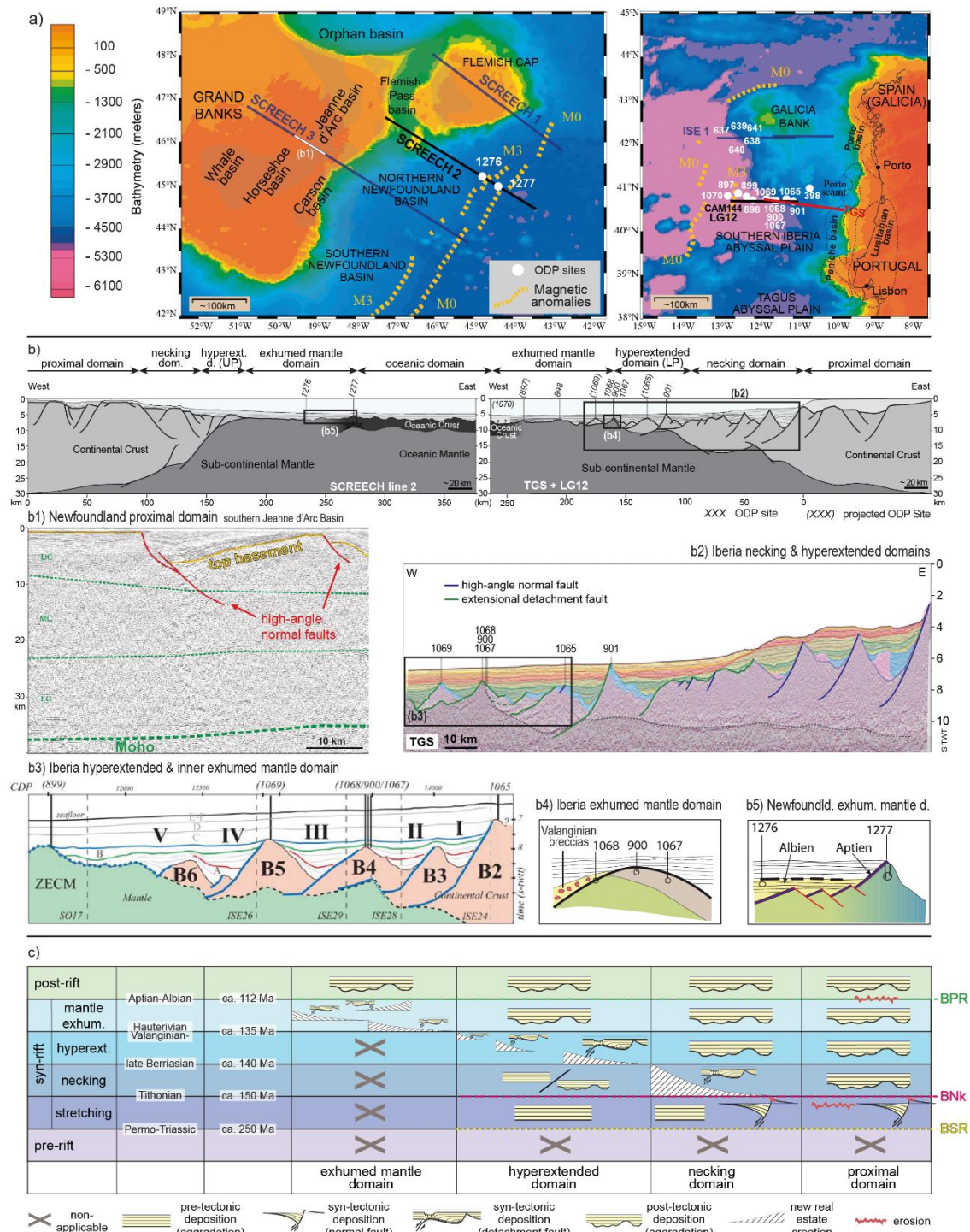
**Figure 6:** Tables summarizing the main characteristics of each system tract in the different rift domains of our idealized model: a) main structures and related stratigraphic architecture; b) distribution of pre-, syn- and post-tectonic sedimentation and tectonic fragmentation; c) evolution of the depositional water depth.



**Figure 7:** a) Schematic map showing the succession of rifting events in Western Europe (modified from Manatschal and Müntener, 2009).; b) Reconstruction of a synthetic section across the former Alpine Tethys rift system during the Late Jurassic margins. See panel (a) for location and note that section A—A' across the European margin is offset from section A''—A''' across the Adriatic margin. Abbreviations: hyperext.: hyperextended; d.: domain; Low. Aust. N.: Lower Austroalpine nappes; OCT: Ocean–Continent Transition; c) and d) Tables synthesizing the sedimentary architectures in each system tract in the different rift domains of the northwestern Adriatic and European margin, respectively.

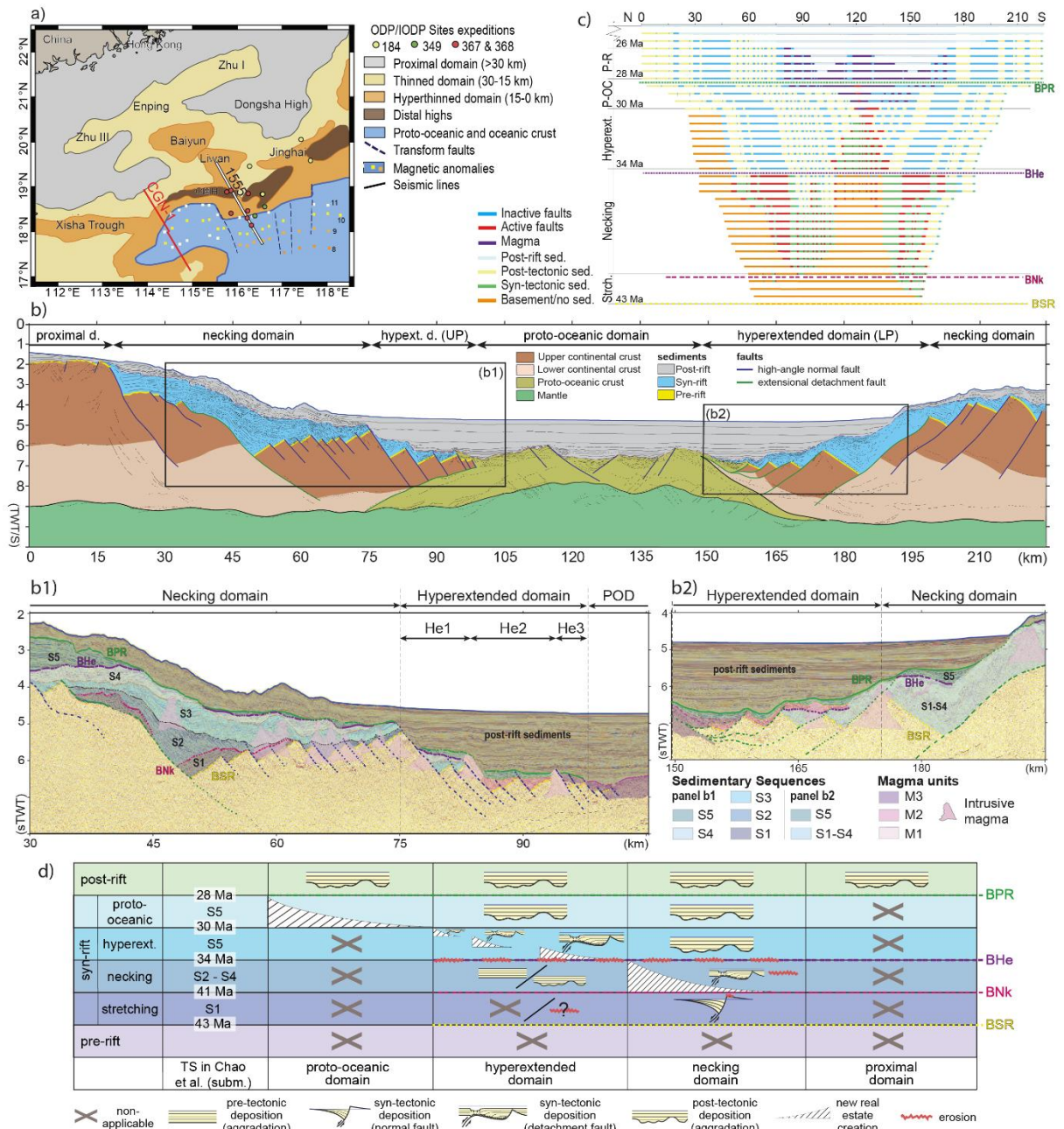
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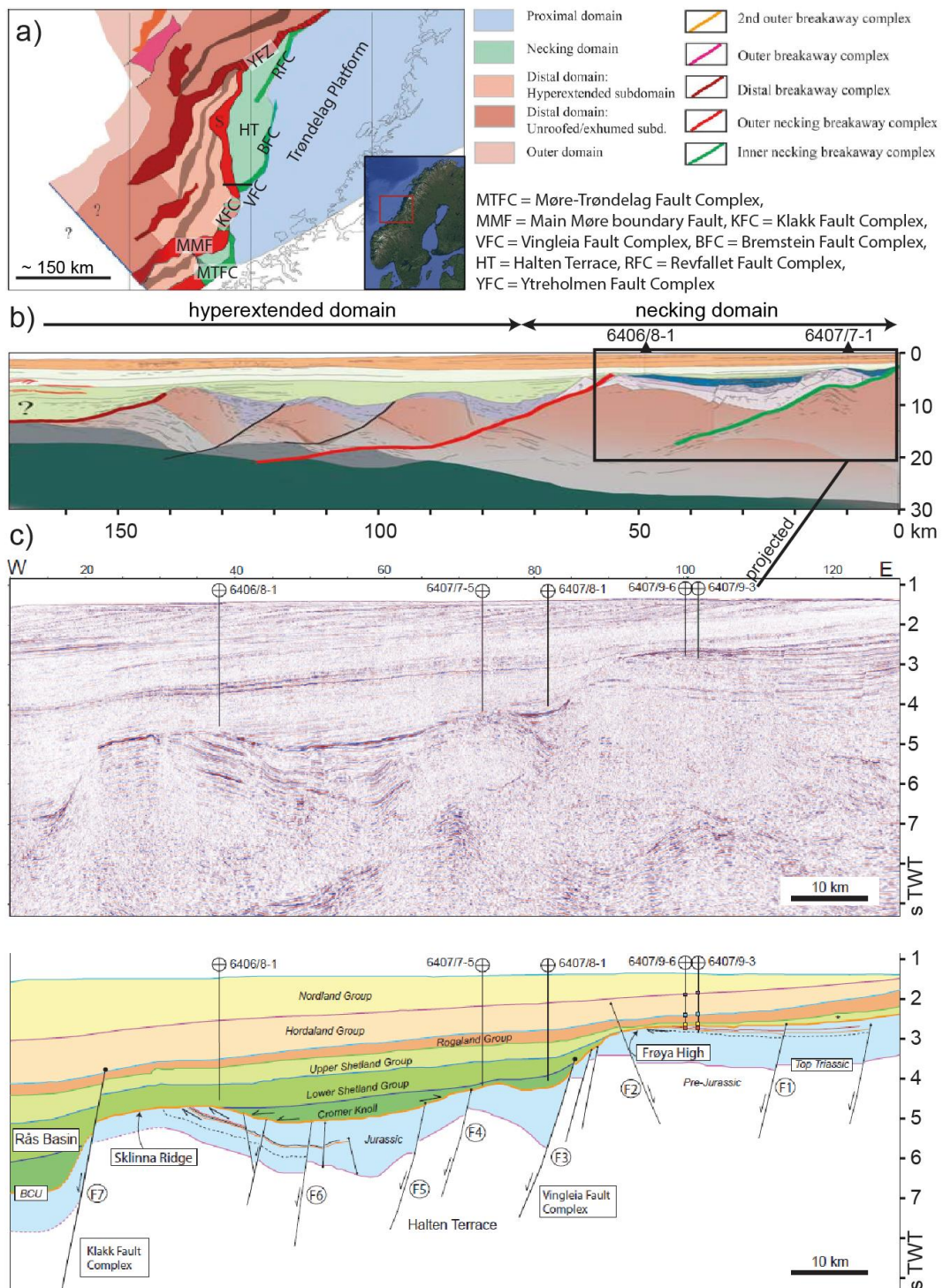
**Figure 8:** a) map of the Iberia–Newfoundland rifted margins and location of the public seismic profiles and drill holes (modified from Sutra *et al.*, 2013); b) interpreted SCREECH 2, TGS and LG 12 seismic sections across the conjugate Iberia–Newfoundland rifted margins (modified from Sutra *et al.*, 2013); b1) close up on the southern Jeanne d’Arc Basin on the SCREECH 3 seismic transect (modified from Deemer *et al.*, 2009); b2) stratigraphic overview of the Iberian necking and hyperextended domains (from Merino-Perez, 2016, and references therein); b3) stratigraphic interpretation of the CAM 144 seismic line imaging the Iberian hyperextended domain (from Péron-Pinvidic *et al.*, 2007); b4) and b5) stratigraphic overview from the exhumed mantle domain of Iberia and Newfoundland, respectively (modified from Sutra *et al.*, 2013); c) Table synthesizing the sedimentary architectures in each system tract in the different rift domains of the Iberian margin.



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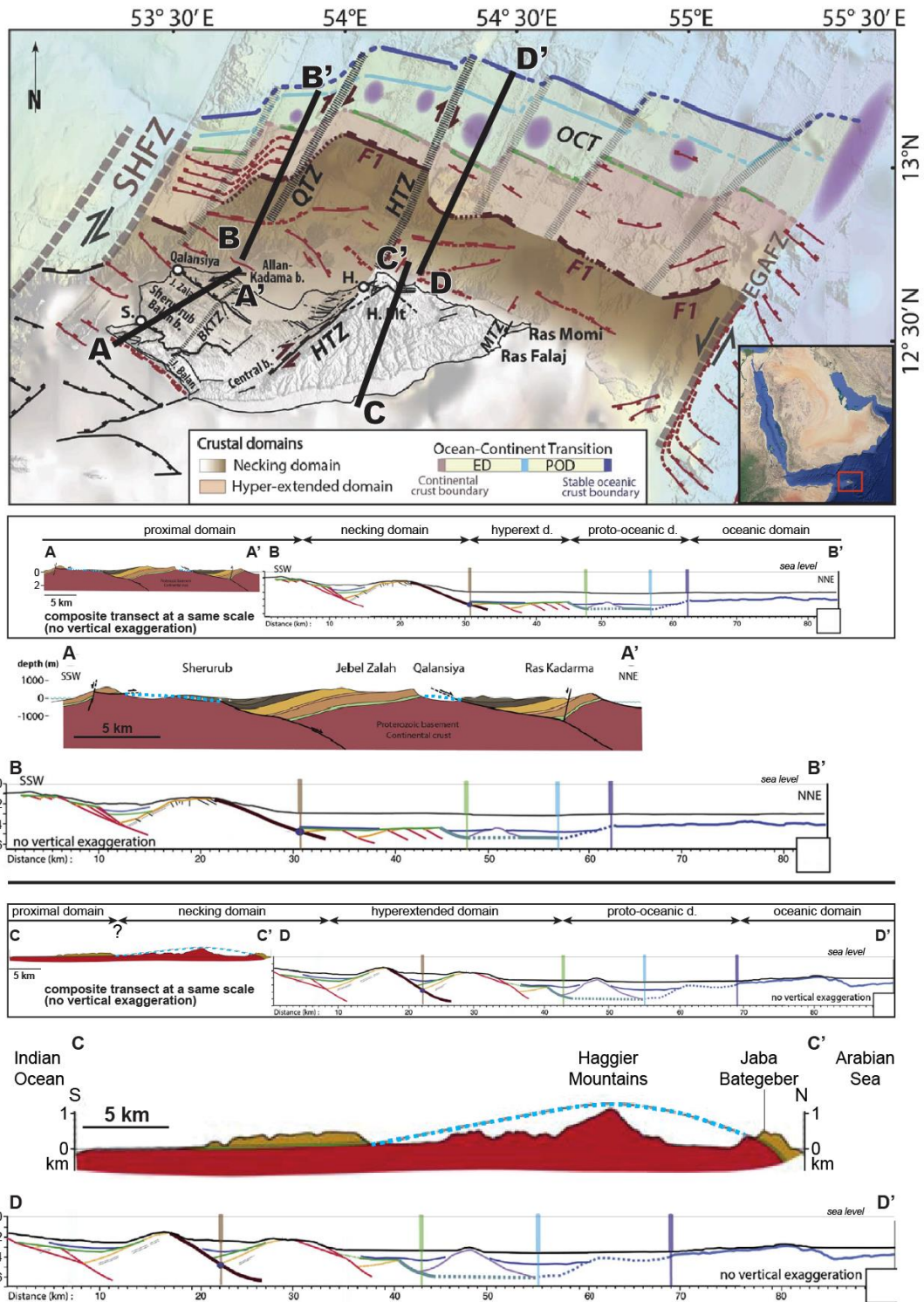
**Figure 9:** a) Rift domain map showing the location of the CGN-1 transect and IODP Sites (modified from Nirrengarten *et al.* 2020); b) linedrawing of the CGN-1 line by Chao *et al.* (2021); b1) and b2) close-up of the outer necking - inner hyperextended domains of the northern and southern margin, respectively; c) spatio-temporal evolution of active vs inactive faulting and sedimentation along the studied transect (from Chao *et al.* 2021); f) Table synthesizing the sedimentary architectures of each system tract in the different rift domains of the northern margin of the NW-SCS. Abbreviation POD stands for Proto-Oceanic Domain.





**Figure 10:** a) Map of the rift domains and intervening breakaway fault complexes at the mid-Norwegian margin (modified from Osmundsen and Péron-Pinvidic, 2018). Inset map from Google Earth; b) Interpreted seismic transect across the necking and hyperextended domains south of the Halten Terrace (no vertical exaggeration; modified from Osmundsen and Péron-Pinvidic, 2018; see panel a for location); c) uninterpreted and interpreted seismic section highlighting pervasive erosion of the Jurassic sequence at the breakaway of both the Vingleia and Klakk fault complexes (from Bell *et al.*, 2014).

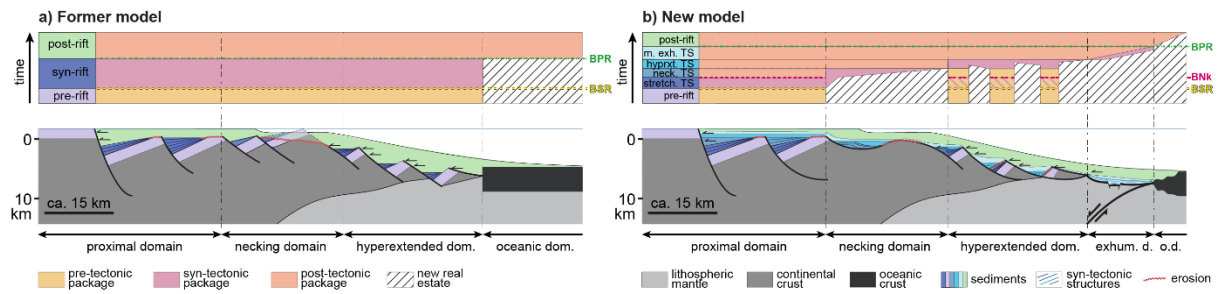
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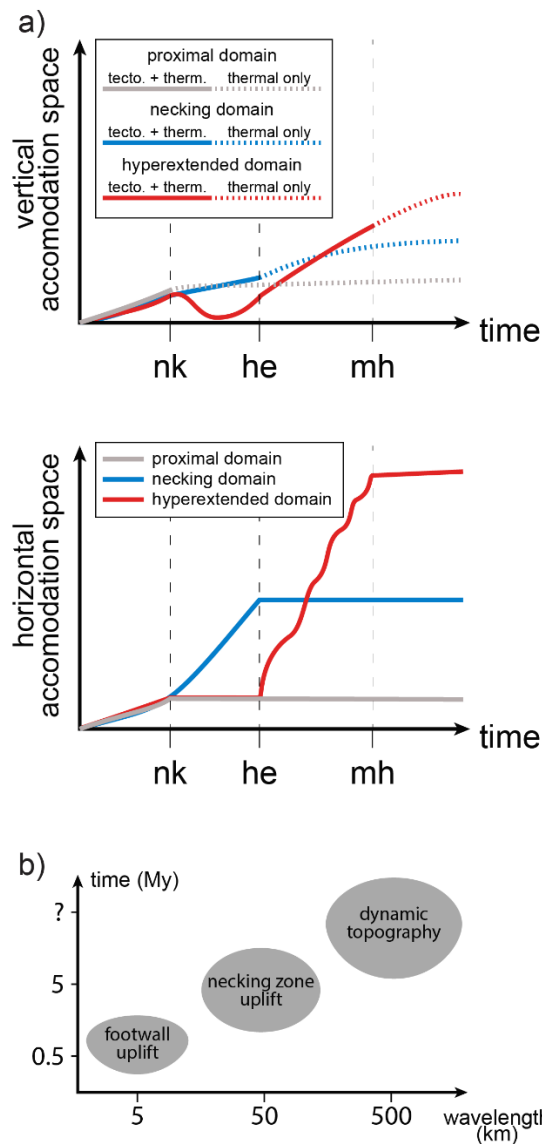
**Figure 11:** Map of the rift domains in the region of Socotra and interpreted transects onshore (A-A' from Nonn *et al.*, 2017; C-C' from (Fournier *et al.*, 2007) and offshore (B-B' and D-D' from Nonn *et al.*, 2017). Transects are first shown at the same scale and without vertical exaggeration for reference. Inset map from Google Earth.





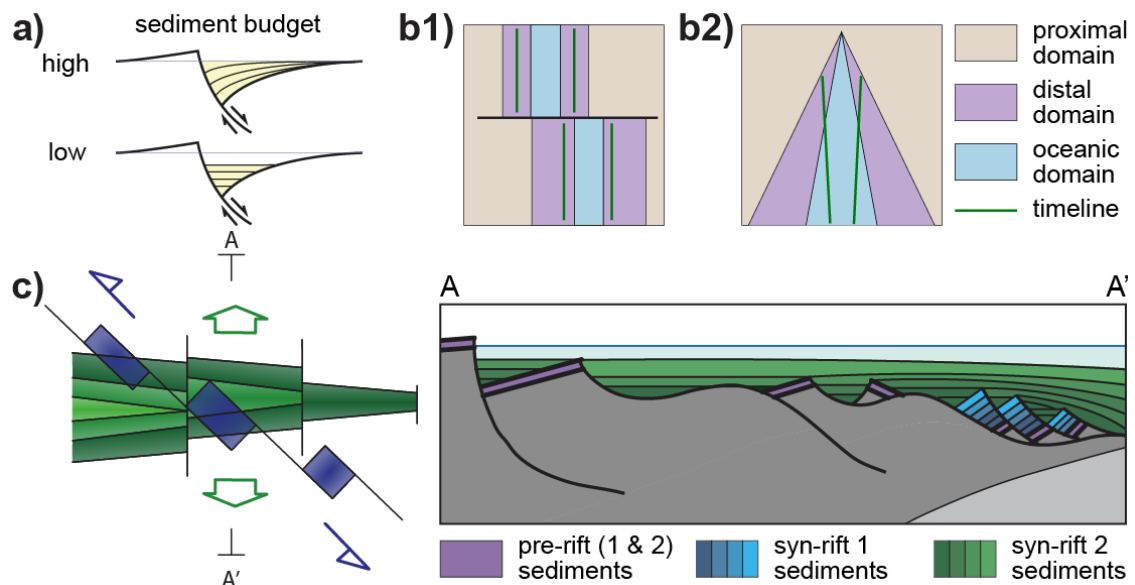
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**Figure 12:** Schematic view of a rifted margin interpretation and corresponding Wheeler diagram according to a) former tectono-stratigraphic models; b) the tectono-stratigraphic presented in this study.



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**Figure 13:** a) Graphs displaying the evolution of vertical and horizontal accommodation creation with time for the different rift domains. Note that vertical accommodation creation may result either from tectonics with a minor thermal component, or from thermal relaxation only. b) Characteristic wavelengths and time of uplift for different tectonic or geodynamic processes.



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**Figure 14:** Main pitfalls to be kept in mind when applying a tectono-stratigraphic model to the interpretation of rift systems: a) impact of the ratio between sedimentation- and subsidence rate: sediment-starved systems will incompletely or even not record tectonic activity. b1) impact of segmentation and b2) propagating extension on the correlation of timelines across rift systems. c) impact of spatially overprinting extensional events (i.e., “multi-stage rifting”), which may incorporate sedimentary sequences unrelated to the rifting event considered (modified from Miró *et al.* 2021).