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1 Influence of the architecture of magma-poor hyperextended rifted
2 margins on orogens produced by the closure of ‘narrow’ versus ‘wide’
3 oceans

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12 **Abstract**

13 Orogens resulting from the closure of ‘narrow oceans’ such as the Alps or the Pyrenees
14 usually lack voluminous syn-subduction and syn-orogenic magmatism. Such orogenies
15 are essentially controlled by mechanical processes in which the initial architecture of the
16 original rifted margins strongly controls the architecture of the orogen. In this paper, we
17 first provide a synthesis of the structure, dimensions and lithology of hyperextended rift
18 systems and oceans, based on recent seismic and petrologic data. Secondly, we investigate
19 how rift-related inheritance influences the crustal geometry and mantle geochemistry of
20 orogens related to the closure of ‘narrow oceans’, and compare them to orogens resulting
21 from the closure of ‘wide/mature oceans’.

22 Our results show that ‘narrow oceans’ usually lack mature spreading system forming
23 Penrose-type oceanic crust (i.e. 6–7-km-thick basaltic oceanic crust typical of steady-state
24 spreading systems; see Anonymous, 1972), in contrast to ‘wide oceans’. However, there
25 is statistically no difference in the structural and lithological architecture of their passive
26 continental margins. Thus, the main difference between ‘narrow’ and ‘wide oceans’ is
27 whether the margins are separated by a significant amount of oceanic crust and under-
28 lying depleted mantle. In addition, due to the lack of significant magmatism during the
29 closure of ‘narrow oceans’, the mantle wedge is likely to remain relatively fertile compared
30 to the wedge above long-lasting subductions of ‘wide oceans’. This difference in mantle

31 composition may dictate the magmatic budget of subsequent orogenic collapse or rifting
32 events.

33 1 Introduction

34 Collisional orogens are often regarded as the result from the telescoping of former rifted
35 margins following subduction of a wide oceanic domain (e.g. Uyeda, 1981; Willett et al.,
36 1993; Ernst, 2005; Handy et al., 2010). Long-lasting, Pacific-type subduction systems are
37 typically associated with volcanic arcs and high temperature/low pressure metamorphism
38 in the hanging wall, both of which strongly modify the architecture, lithology and thermal
39 state of the initial margin (Miyashiro, 1961, 1967; Ernst et al., 1970; Gerya, 2011). There-
40 fore, following the closure of a ‘wide ocean’, at least one side of the orogen is significantly
41 overprinted by subduction-induced processes. This may explain why, except for a few stud-
42 ies (De Graciansky et al., 2011; Butler et al., 2006; Butler, 2013; Mohn et al., 2011, 2014;
43 Beltrando et al., 2014; Tugend et al., 2015, and references therein), little attention has been
44 paid to the potential impact of the initial architecture of the intervening rifted margins.
45 However, orogens such as the Alps, the Pyrenees and the Variscides of Western Europe
46 supposedly result from the closure of ‘narrow’, possibly ‘embryonic’ oceans (i.e. devoid of
47 a mature seafloor spreading system; Vlaar and Cloetingh (1984); Pognante et al. (1986);
48 Rosenbaum and Lister (2005); Mohn et al. (2010)) and lack evidence for voluminous arc
49 magmatism contemporaneous with subduction. These orogenies were essentially controlled
50 by mechanical processes, where the architecture of the initial margins dictated largely the
51 architecture of the resulting orogen (see Roca et al. (2011); Beltrando et al. (2014) for
52 reviews). Therefore, the knowledge of the structure, dimensions and lithology of rifted
53 margins and oceans is of primary importance for the understanding and the modelling of
54 collisional orogens.

55 This paper aims to provide a synthesis of the first-order characteristics of hyperextended
56 rift systems and oceans, at a scale and resolution compatible with lithospheric-scale thermo-
57 mechanical numerical codes. First of all, based on recent seismic and petrologic data, we
58 synthesize the structure and lithology of magma-poor hyperextended rifted margins and
59 oceans, and provide a compilation of the dimensions of the different structural domains
60 comprising rifted margins. We compare the characteristics between ‘narrow’/‘immature’
61 and ‘wide’/‘mature oceans’. We define ‘narrow oceans’ as rift systems that reached at
62 least the stage of hyperextension (see Sutra et al. (2013) and Doré and Lundin (2015) for
63 the definition and a review of *hyperextension*), but remained less than ~ 300 km wide

64 (the reasons for this limit are discussed below). We use the term ‘mature’ for oceans that
65 comprise a self-sustaining, steady-state seafloor spreading system, as opposed to ‘imma-
66 ture’ or ‘embryonic’ oceans whose development stopped at the stage of hyperextension or
67 exhumation. Second, we investigate how the first-order rift-related inheritance of narrow
68 oceans may influence orogens resulting from their closure, and compare these ‘immature
69 orogens’ with classical ‘mature orogens’ produced by the closure of wide/mature oceans
70 (Uyeda, 1981).

71 **2 Characteristics of magma-poor hyperextended rifted mar-** 72 **gins**

73 Observations of rift systems indicate that during the early stages of rifting extension is
74 often distributed across a relatively wide zone containing multiple rifts (Withjack et al.,
75 1998; Skogseid, 2010). These observations can be compared with the numerical models by
76 Braun and Beaumont (1989); Beaumont and Ings (2012); Chenin and Beaumont (2013);
77 Harry and Grandell (2007), which show that, while the crust is decoupled from the mantle,
78 extension may be accommodated by several inherited crustal weak zones at the same time.
79 Consequently, a series of ‘failed rift’ basins, which may be offset from the locus of the
80 final breakup, develop in the early stages of extension. Once the crust is coupled to the
81 mantle, extension localizes in one basin while the others are abandoned. During this stage,
82 the crust and lithospheric mantle are progressively thinned, until the eventual lithospheric
83 breakup occurs.

84 Recent studies suggest that most rifted margins whose necking and hyperextension
85 phases were magma-poor display a similar succession of comparable domains (Figure 1),
86 regardless of whether or not they achieved lithospheric breakup and steady-state seafloor
87 spreading (Sutra et al., 2013; Péron-Pinvidic et al., 2013). In this section, we characterize
88 the first-order architecture and lithology of such rift systems and quantify the width of
89 their distal sub-domains based on natural examples.

90 **2.1 Primary architecture**

91 Since the extensive seismic and drilling surveys of the 90’s, rifted margins are no longer
92 regarded as a simple series of tilted blocks adjacent to Penrose-type oceanic crust (6–7-
93 km-thick basaltic oceanic crust typical of steady-state spreading systems; see Anonymous,
94 1972). Indeed, the drilling of exhumed mantle offshore Iberia led to the reconsideration

95 of the architecture of rifted margins (Boillot et al., 1980). In order to characterize the
96 architecture of hyperextended rifted margins, Sutra et al. (2013) distinguished several
97 domains based on morphological criteria (Figure 1). On the one hand, the *proximal domain*
98 corresponds to un-thinned or minor extended ($\sim 30\text{--}35$ km thick) continental crust and
99 is characterized by parallel, roughly flat basement and Moho topographies. On the other
100 hand, a typical *oceanic domain* is made of homogeneous, Penrose-type oceanic crust, about
101 6–7 km thick. Here again, basement and Moho are parallel. The *distal domain*, between
102 them, records most of the rift-related deformation.

103 The distal domain can be divided into several sub-domains (Figure 1), namely: (a) a
104 *necking domain* characterized by the abrupt thinning of the continental crust from $\sim 30\text{--}$
105 35 km down to ~ 10 km, which translates to a deepening of top basement and shallowing of
106 the Moho (the latter only on seismic sections in depth); (b) a *hyperextended domain*, where
107 continental crust is thinned from ~ 10 km down to 0 km. The transition between the neck-
108 ing and hyperextended domain corresponds to a sudden decrease in the dip of the Moho
109 on depth seismic sections; (c) a so-called *exhumation domain* may exist in magma-poor
110 rifted margins when lithospheric sub-continental mantle is exhumed at the seafloor (Boillot
111 et al., 1987; Whitmarsh et al., 2001; Manatschal and Müntener, 2009). Mantle exhuma-
112 tion is associated with serpentinization down to a depth of 4–6 km (Boillot et al., 1989;
113 Escartín et al., 2001; Minshull et al., 1998), and the transition from serpentinized mantle
114 to fresh peridotite translates to a progressive increase in seismic velocity (Horen et al.,
115 1996; Miller and Christensen, 1997; Skelton et al., 2005). Mantle exhumation may also
116 be accompanied with the emplacement of discontinuous extrusive magmatic rocks from an
117 immature spreading system, which is characterized by incomplete development of oceanic
118 layers 2 and 3 (e.g. Desmurs et al., 2002; Müntener et al., 2004; Jagoutz et al., 2007). Note
119 that allochthonous blocks of continental crust may be found on top of the serpentinized
120 exhumed mantle of the exhumation domain. Thus, on seismic sections, the exhumation
121 domain appears as a more or less structured surface either devoid of a Moho reflector, or
122 with a discontinuous seismic Moho reflector when accompanied with magmatism (Whit-
123 marsh et al., 2001). The transition from the exhumation domain into the oceanic domain is
124 highlighted by an abrupt step-up in the basement (Bronner et al., 2011), which reflects the
125 decrease in the density of the lithospheric column resulting from the production of thicker
126 oceanic crust through increased magmatic activity. In contrast, in the case of magma-rich
127 margins, the transition from thickened magmatic crust to steady-state seafloor spreading
128 is expressed as a step down onto oceanic crust. The first-order morphology of magma-poor

129 rifted margins is highlighted by the red lines on Figure 1.

130 **2.2 Dimensions and maturity of rift systems**

131 In order to assess the dimensions of rift systems and their distal sub-domains, in addition to
132 their maturity, we first compile the *width* and the lithology of basement rocks for several
133 narrow and wide extensional systems around the world (Table 1). In this paper, the
134 width of an extensional system refers to the distance between the two conjugate necking
135 points (Figure 1). From our compilation (Table 1), we suggest that extensional systems
136 narrower than 300 km are usually immature, that is devoid of a self-sustained, steady-
137 state spreading system, and are floored with either thinned continental crust, exhumed
138 mantle and/or embryonic oceanic crust (Table 1). Note, however, that for rift systems
139 whose development is accompanied by a plume, such as the eastern Gulf of Aden and the
140 southern Red Sea, the rifted margins may be narrower and seafloor spreading may start
141 earlier.

142 Second, we measure the width of the distal sub-domains (namely the necking, hyperex-
143 tended and exhumation domains) for a selection of published seismic sections (Table 2).
144 For this compilation, we only consider dip seismic lines imaging extensional systems formed
145 in a single, unidirectional and continuous rifting event (i.e. no significant time lag without
146 extension). Note that the necking domain of the ISE1 seismic section seems anomalously
147 wide compared to the other data points (see Figure 2 b). Its unique architecture displaying
148 a ‘double neck’ (see seismic section 2 in Table 2) suggests that it did not form in a single
149 extensional event, since deformation tends to become more and more localized as extension
150 progresses. Therefore, we remove this data point in the statistical diagrams c, d, e, f and
151 g of Figure 2, but we include it in our discussion with the asterisk.

152 In addition, we restrict our measurements to domains with minor magmatic additions.
153 Note that we did not include the width of the exhumation domain for rift basins that did
154 not achieve lithospheric breakup, because in this case the full width of the exhumation
155 domain is not realized.

156 Figure 2 (c) shows that the width of the distal domain ranges from about 130 to 240 km
157 (or to *350 km), with an average of 170 km (*185 km) and a median of 165 km (*165 km).
158 Of all the distal sub-domains, the necking and exhumation domains have the highest vari-
159 ability in width (see Figure 2 d and f). The width of the necking domain ranges from
160 about 10 to 100 km (or to *210 km) with an average \sim 55 km (*60 km), a median \sim 50 km
161 and a standard deviation of 25 km (*40 km). The width of the exhumation domain varies

162 between about 20 km to 110 km, with an average of 60 km, a median of ~ 70 km and
163 a standard deviation of ~ 30 km. In contrast, the width of the hyperextended domain
164 seems more constant, varying between 20 and 70 km and with an average of 50 km (me-
165 dian ~ 60 km; standard deviation 15 km). In addition, Figure 2 (e) indicates that no
166 correlation exists between the width of any of the distal sub-domains and the total width
167 of the distal domain. This means that magma-poor hyperextended rifted margins, and
168 hence each of their comprising distal sub-domains, have a specific range of width and can
169 therefore be described with an ‘average structural architecture’ (see the middle panel of
170 Figure 3 a).

171 Figure 2 (g) suggests that the width of both the necking and hyperextended domains
172 is unrelated to the maturity of the extensional system, thus to whether it is a ‘narrow’
173 or a ‘wide’ ocean. Therefore, the distal sub-domains of narrow/immature oceans have a
174 structural architecture similar to those of wide/mature oceans (compare the middle panels
175 of Figure 3 a and b).

176 **2.3 First-order lithological architecture**

177 Although rifting affects an initially more or less horizontally homogeneous lithosphere
178 (‘layer-cake’ lithosphere), tectonic processes and local fluid-rock interactions (including
179 magma) may increasingly modify the lithology and thermal state, and thus the rheology,
180 of the intervening lithosphere as extension progresses.

181 When rifting is not triggered by magmatism (for example, upwelling of a mantle plume),
182 the lithology of the crust is not significantly modified by extension in the proximal do-
183 main and in the necking domain. Therefore, the continental crust of both domains can be
184 approximated by a quartzo-feldspathic material, as usually considered by numerical mod-
185 ellers. Note that, in the necking domain, the ductile layers are mechanically attenuated
186 during extension, so that the crust is fully brittle in the hyperextended domain (Pérez-
187 Gussinyé et al. (2003); Sutra et al. (2013); Manatschal et al. (2015); see the bottom panel
188 of Figure 3 a and b). In addition, the subcontinental mantle underlying the proximal and
189 necking domains is not significantly modified and can be approximated by an average ‘in-
190 herited mantle’ consisting of peridotite with highly variable composition (~ 50 – 70% olivine
191 (ol), ~ 1 – 30% clinopyroxene (cpx), ~ 20 – 30% orthopyroxene (opx) and ~ 1 – 4% spinel
192 (sp); Müntener et al. (2010); Picazo et al. (2016)) and pyroxenite (ol as low as 0% , cpx up
193 to 80% , opx up to 60% and 1 – 2% spinel/garnet).

194 In contrast, in hyperextended and exhumation domains, fluid-rock interactions form hy-

195 drous minerals, similar to the fluid-rock interactions observed in present-day mid-oceanic
196 ridges (Mével, 2003; Bach et al., 2004; Boschi et al., 2006; Picazo et al., 2012, 2013).
197 Hydrothermal circulation is responsible for the formation of hydrous minerals within hy-
198 perextended crust (e.g. sericite and illite; Pinto (2014) and references therein), as well as
199 in exhumed mantle (e.g. serpentinite, chlorite, talc; see Hess (1955); Christensen (1970);
200 Früh-Green et al. (2004); Picazo et al. (2013)). Therefore, the rheology of the hyperex-
201 tended continental crust and the top 4–6 km of lithospheric mantle in the exhumation
202 domain can be approximated by a phyllosilicate-type rheology.

203 Significant crustal thinning is also associated with partial melting of the asthenosphere
204 (Latin and White, 1990). The resulting melts are initially not, or only little extracted,
205 but impregnate the overlying lithospheric mantle in the hyperextended domain (Müntener
206 et al., 2010). Thus, the subcontinental mantle underlying the hyperextended crust and
207 exhumed mantle displays a ‘fertile’ composition with a significant amount of plagioclase
208 and clinopyroxene, and evidence for melt rock reaction (e.g. Müntener and Piccardo,
209 2003; Müntener et al., 2010). As a consequence, immature rift systems whose development
210 stopped at the stage of hyperextension are likely to retain this fertile mantle composition
211 (Figure 3 b, bottom panel).

212 In contrast, during lithospheric breakup, melt starts to be extracted and a sustainable
213 magmatic system is established marking the onset of steady-state seafloor spreading. The
214 resulting oceanic crust is homogeneous in both composition and thickness (basaltic to gab-
215 broic and ~ 6 –7 km thick; Anonymous (1972)). The process of partial melting associated
216 with the creation of oceanic crust depletes the underlying mantle in the most fusible el-
217 ements, leaving an average composition of 57% ol, 13% cpx, 28% opx and 2% sp (i.e. a
218 ‘depleted DMM’ (DMM = Depleted MORB Mantle); see Workman and Hart (2005) and
219 Figure 3 a, bottom panel). Note that, even in the case of a so-called *magma-poor rift-*
220 *ing*, onset of steady-state seafloor spreading is presumably triggered by a magmatic pulse
221 (Bronner et al., 2011). Therefore, we expect the mantle underlying the most distal part of
222 the margins of mature oceans to be depleted in fusible elements, conversely to the mantle
223 underlying immature oceans.

224 We calculated the density of fertile and depleted mantle 700°C to 1300°C in steps of
225 100°C, at pressure from 0.5 to 3 GPa (Table 3) based on their average modal composition
226 (Müntener et al., 2010; McCarthy and Müntener, 2015), their physical properties (Hacker
227 et al., 2003) and the geotherm of each mantle-type domain (Sclater et al., 1980). At
228 pressure of 0.5 GPa, calculations show that fertile plagioclase peridotite is less dense than

229 depleted harzburgite. From 1.0 to 2 GPa, depleted mantle is lighter by about 20–30 kg.m⁻³
230 compared to fertile mantle. Note that the peridotites considered in this calculation are
231 relatively rich in Fe, but if they would be richer in Mg, the density contrast would be of
232 the order of 50–80 kg.m⁻³. At high pressures, because of the higher modal abundance
233 of garnet in fertile rocks, the fertile peridotite is denser than the residual harzburgite by
234 about 30–40 kg.m⁻³.

235 The density contrasts observed between fresh versus altered peridotites and depleted vs.
236 fertile mantle are significant enough to influence tectonic processes as shown for example
237 for basin subsidence (Kaus et al., 2005). In particular the location of subduction initiation
238 during the inversion of an ocean may be controlled by such density contrasts.

239 **3 First-order architecture of collisional orogens**

240 Every ocean will eventually undergo subduction and be sutured after collision of its con-
241 tinental margins. On a first-order, a collisional orogen can be regarded as a three-part
242 system comprising (Figure 4 a2 and b2): (1) two buttresses; (2) an accretionary wedge;
243 and (3) a subducted part.

244 In this section, we discuss how these different components may correlate with specific
245 parts of the initial rift system. In order to avoid the complexity induced by magmatic
246 overprinting, we focus on the well-studied Pyrenean and Alpine orogens, both of which
247 largely lack voluminous syn-subduction and syn-orogenic magmatism and are therefore
248 essentially controlled by mechanical processes. Moreover, as both are relatively recent,
249 their orogenic architecture is fairly well-preserved.

250 Oceanic lithosphere, due to its high density, tends to be efficiently subducted. According
251 to Stern (2004), most of the ophiolites preserved within orogens correspond to obducted
252 remnants of buoyant oceanic crust from small and young oceanic basins, usually former
253 forearcs or backarcs, rather than mature oceanic crust. Besides, a significant proportion
254 of sediments is also usually subducted, while the remaining part accumulates in the ac-
255 cretionary prism (Clift et al., 2004; Stern, 2011, and references therein). A significant
256 proportion of the subducted material has to be integrated to the orogenic root in order
257 to account for the isostasy of collisional orogens, however the deep architecture of colli-
258 sional orogens is very poorly constrained (Butler, 2013) as illustrated by the diversity in
259 the interpretation of the deep part of the ECORS-CROP seismic section (Figure 5 b).

260 Besides, both the Alpine and the Pyrenean orogens display a similar architecture, where
261 the external domain is made of little deformed continental basement and the internal part

262 of a complex stacking of material originating from the distal margin (De Graciansky et al.,
263 2011; Bellahsen et al., 2014; Butler et al., 2006; Butler, 2013; Schmid et al., 2004; Beltrando
264 et al., 2014; Casteras, 1933; Mattauer, 1968; Jammes et al., 2009; Muñoz, 1992). Recent
265 studies (Mohn et al., 2014; Tugend et al., 2015) show consistently that the external parts
266 of the orogen (the buttresses B on Figure 4) correspond to the little deformed necking
267 zones of the former continental margins. In contrast, the internal part corresponds to the
268 accretionary prism (A on Figure 4), which is comprised of thinned continental basement
269 remnants, ophiolites and/or exhumed mantle and thick sequences of highly deformed sed-
270 iments (see Beltrando et al. (2014) for a review). This accreted material is essentially
271 derived from the sedimentary cover, but according to Andersen et al. (2012), some parts of
272 the hyperextended and exhumation domains may also be integrated into the accretionary
273 wedge, as evidenced by the remnants of exhumed mantle in the Caledonian orogen (see also
274 Chew and Van Staal (2014). This proposition is also verified in the Alps and in the Pyre-
275 nees, where most ophiolites are remnants from ‘inherited’ or ‘refertilized’ subcontinental
276 mantle rather than from a steady-state spreading system (Lemoine et al., 1987; Müntener
277 et al., 2010; Picazo et al., 2016).

278 What controls the distribution of accreted versus subducted material remains poorly
279 constrained. From the seismic refraction profile across offshore Iberia and the interpretation
280 of an adjacent transect calibrated by drill holes shown in Figure 5 (a), we speculate that
281 the boundary between sediments and hydrated material on the one hand (green and blue
282 colors on the refraction profile) and the unaltered underlying mantle on the other may act
283 as a decollement separating the accreted and subducted part once the hyperextended and
284 exhumation domains reach the subduction trench.

285 The heat budget is significantly different between mature and immature subductions/orogens.
286 Indeed, in mature subduction systems, a large amount of heat is advected by the magma
287 transported from the mantle to the crust, inducing arc magmatism and batholith formation
288 in the hanging wall of the subduction (Uyeda, 1981; Stern, 2002). The associated fluids
289 and magma transported into the overlying crust weaken it considerably (Gerya, 2011),
290 which has a major impact on the architecture of the subsequent orogen (Figure 4 a).

291 In contrast, although high-grade metamorphism is recorded in the Alps (for instance in
292 the Tauern window and the Lepontine dome), heat transfer from the mantle is limited.
293 The metamorphic belts of Alpine-type systems are created by stacking upper continental
294 crustal slivers and internal heat production by radioactive decay (Burg and Gerya, 2005).
295 The resulting metamorphism may have some impact on the rheology of the buried crust

296 (Bellahsen et al., 2014), but is not significant enough to affect the overall architecture
297 of the orogen (Figure 4 b). Therefore, in our study, we neglect the effects of low-grade
298 metamorphism in immature orogens.

299 In summary, when the closure of a (narrow) ocean is devoid of significant magmatism,
300 the orogenesis is essentially controlled by mechanical processes and its first-order geometry
301 can be related to specific portions of the initial rift system (Figure 4): (1) the buttresses
302 correspond to the proximal plus necking domains (see Figure 1); (2) the accretionary prism
303 to part of the hyperextended and/or exhumation domains, in addition to part of the overall
304 basin sediments; and (3) the subducted part to most of the hyperextended and exhumation
305 domains, the oceanic lithosphere and part of the distal and oceanic sediments.

306 4 Discussion

307 Since the advent of plate tectonics and the understanding of first-order subduction pro-
308 cesses, collisional orogens are usually regarded as the result of the telescoping of continen-
309 tal margins after long-lasting subduction of a wide oceanic domain (Uyeda, 1981; Willett
310 et al., 1993; Ernst, 2005; Handy et al., 2010). Such orogens are characterized by paired
311 metamorphic belts, namely a low temperature/high pressure belt corresponding to the ac-
312 cretionary wedge; and a high temperature/low pressure belt related to arc metamorphism
313 and/or magmatism (Miyashiro, 1961; Dewey and Horsfield, 1970; Brown, 2009, and refer-
314 ences therein). However, several collisional orogens such as the Alps and the Pyrenees lack
315 voluminous magmatism contemporaneous to subduction, that is, are devoid of remnants
316 of arcs, forearcs and backarcs and of high temperature/low pressure metamorphic assem-
317 blages in the upper plate. Yet, both orogens result from the closure of relatively narrow (<
318 400–600 km for the Alpine Tethys, < 200 km for the Pyrenean rift system), hyperextended
319 (Pyrenean rift system) or possibly embryonic (Alpine Tethys) ‘oceans’ (Rosenbaum et al.,
320 2002; Rosenbaum and Lister, 2005; Mohn et al., 2010; Lemoine et al., 1987). This may be
321 the main cause for the lack of significant magmatic products. Indeed, because significant
322 dehydration of the basaltic crust and serpentinized mantle starts only from a depth of 100–
323 200 km (Peacock et al., 1994; Rüpke et al., 2004), a magmatic arc is unlikely to develop
324 before the slab is subducted to this depth (e.g. Jarrard, 1986; England et al., 2004). There-
325 fore, there must be a critical width for rift systems, below which their subduction is devoid
326 of significant magmatic activity. In such cases, the subsequent orogenies may be essentially
327 controlled by mechanical processes, in which the structural and lithologic architecture of
328 the intervening margins may be the dominant factor in controlling the architecture of the

329 orogen.

330 4.1 ‘Narrow oceans’ and ‘magma-poor’ subduction

331 As slabs subduct with an average dip of 50–60° in the upper mantle (Stevenson and Turner,
332 1977; Tovish et al., 1978; Billen, 2008), the slab must be at least 130 km long to reach a
333 depth of 100 km. Adding to this twice the 55 km average length of each necking domain
334 (section 2.1), which is usually not subducted (section 3), magma generation seems very
335 unlikely during closure of rift systems narrower than 240 km. Furthermore, subduction
336 must last long enough for a sufficient amount of volatiles to induce hydrous partial melting
337 (Peacock, 1991; Gaetani and Grove, 1998). This is consistent with the compilation by
338 Jarrard (1986), which shows that the length of the subducted slab associated with the
339 youngest arc (the Philippines arc, 6 m.y. old) is at least 170 km long. As a consequence,
340 we expect rift basins narrower than 300 km to be devoid of significant magmatism expressed
341 at the surface. In the following, we refer to these as *narrow oceans*, as opposed to *wide*
342 *oceans* larger than $\sim 1,000$ km.

343 In addition to this ‘flux melting’, arc magma generation is also driven by decompres-
344 sion melting of the hot asthenosphere rising to compensate the down-dragging of mantle
345 wedge material by the slab (Iwamori, 1998; Jagoutz et al., 2011; Sisson and Bronto, 1998).
346 Yet, decompression melting becomes important only when vigorous convection is active in
347 the mantle wedge, which only develops after significant subduction (Peacock et al., 1994;
348 Conder et al., 2002). Thus, it is unlikely to occur during the closure of a narrow ocean.

349 4.2 Characteristics of ‘narrow’ versus ‘wide’ oceans

350 As highlighted in Table 1, extensional systems narrower than 300 km are usually devoid
351 of a mature, self-sustaining spreading system, thus of ‘normal’ oceanic crust. As a result,
352 their seafloor is either comprised of thinned continental crust, exhumed mantle and/or
353 embryonic oceanic crust. On the contrary, mature, steady-state oceanic systems are usually
354 characterized by a homogeneous, Penrose-type oceanic crust. Of course, there is a wide
355 range of oceanic crust types between these ‘embryonic’ and ‘Penrose’ end-members. For
356 instance, in ultra-slow spreading systems, a thin (2–5 km thick) ‘oceanic crust’ comprised
357 of both magmatic and a-magmatic segments may be steadily emplaced for millions of
358 years (Dick et al., 2003). Yet, like normal oceanic crust, oceanic crust formed at ultra-slow
359 spreading ridges is most likely to be efficiently subducted during the closure of wide oceans
360 (Stern, 2004), thus it will *a priori* not influence the subsequent orogeny.

361 As we showed in section 2.1, there is statistically no relationship between the size or
362 maturity of a magma-poor hyperextended rift system and the architecture of its margins.
363 Therefore, on a first-order, the main difference between narrow and wide oceans is the
364 existence of a significant amount of oceanic crust and underlying depleted mantle (compare
365 Figure 3 a and b). In contrast, the mantle underlying narrow embryonic oceans is likely
366 to retain its fertile composition resulting from the impregnation by asthenospheric melts
367 during hyperextension (Müntener et al., 2010).

368 **4.3 Subduction of ‘narrow’ versus ‘wide oceans’ and subsequent oroge-** 369 **nies**

370 During short-lived subduction associated with the closure of narrow ‘oceans’, the slab
371 remains at relatively shallow angle (Billen, 2008) and no self-sustaining subduction will
372 likely develop due to insufficient slab pull (Hall et al., 2003; Gurnis et al., 2004), thus
373 development of small-scale convection above the subducting slab is unlikely (Peacock et al.,
374 1994). Furthermore, the small length of the slab may not allow for a significant amount of
375 volatiles to reach the critical depth for entering the hot part of the mantle wedge (Rüpke
376 et al., 2004; Grove et al., 2006). In such circumstances, the generation of arc magmas is
377 limited and hydration of the mantle wedge is likely to be the dominant process (Peacock
378 et al., 1994).

379 In addition, because of the low dip angle of the slab, both slab pull and the potential
380 effect of mantle flow on the slab are limited, making the development of backarc basins
381 unlikely (Uyeda, 1981; Heuret and Lallemand, 2005). This is supported by the worldwide
382 compilation of backarc deformation style by Heuret and Lallemand (2005), which high-
383 lights that no young subduction is associated with strongly extensional backarc settings.
384 Therefore, the lithosphere underlying orogens resulting from the closure of narrow oceans
385 is likely to be relatively fertile and hydrated (Figure 6 b).

386 In contrast, protracted subduction associated with the closure of wide oceans is likely to
387 become self-sustained, in particular, due to the eclogitization of the subducting slab, which
388 makes it denser than the encompassing asthenospheric mantle (Doin and Henry, 2001; Aoki
389 and Takahashi, 2004). Long lasting subduction develops usually vigorous convection in the
390 mantle wedge, which efficiently transports volatiles derived from the dehydration of the
391 slab to great depth (Peacock et al., 1994). The resulting partial melting creates thickened
392 sialic crust at the surface (magmatic arcs), induces high-temperature metamorphism in the
393 encompassing upper plate and depletes the underlying mantle in fusible elements (Uyeda,

394 1981).

395 When a significant amount of oceanic lithosphere is subducted, the strong slab pull, and
396 potentially the effect of dynamic mantle flow dragging on the slab may induce a backward
397 migration of the lower plate with respect to the upper plate (slab rollback), which may help
398 form backarc basins (Uyeda, 1981; Heuret and Lallemand, 2005). In such cases, back-arc
399 extension may be associated with seafloor spreading and underlying mantle depletion as
400 well. Note that, while vigorous convection within the mantle wedge tends to homogenize
401 its composition, a lower mantle fertility is still to be expected beneath an orogen resulting
402 from the closure of a wide ocean. This assumption is supported by the depleted mantle
403 wedge composition of the Pacific subduction compiled by Woodhead et al. (1993) and the
404 decrease in mantle fertility with increasing distance to the arc region observed in the Lau
405 and Mariana backarc regions (Martinez and Taylor, 2002). Therefore, the thermal and
406 lithological architecture of orogens related to the closure of wide oceans may largely differ
407 from thermal and lithological architecture of their initial margins, as opposed to orogens
408 consequent upon the closure of narrow oceans (compare Figure 6 a and b).

409 **4.4 Impact on subsequent collapse or rifting magmatic budget**

410 The difference in mantle composition resulting from the closure of narrow/embryonic
411 oceans versus wide oceans may dictate the magmatic budget of subsequent extensional
412 events such as post-orogenic collapse or rifting. Indeed, the depleted mantle beneath
413 orogens related to mature subduction systems may not allow for voluminous magma pro-
414 duction, in contrast to the fertile mantle underlying orogens produced by the closure of
415 narrow oceans. This hypothesis may account for both the a-magmatic collapse of the Scan-
416 dinavian Caledonides and the large amount of magmatism during the Variscan orogenic
417 collapse.

418 The Variscides of Western Europe result from the closure of several narrow ‘oceans’
419 (McKerrow et al., 2000a; Franke, 2006), in addition to the suturing of the wide (> 2,000 km;
420 Torsvik, 1998; Nance and Linnemann, 2008) Rheic Ocean (see Matte (2001) and Kröner
421 and Romer (2013) for reviews). Only the closure of the Rheic Ocean did presumably form a
422 significant magmatic arc (Franke, 2006). The orogenic collapse of the Variscan topography
423 was followed by significant magmatic activity, which resulted in widespread, more or less
424 acidic intrusions within the crust, and formed a thick mafic crustal underplating across
425 most of the orogenic area (Bois et al., 1989; Rey, 1993; Costa and Rey, 1995; Schaltegger,
426 1997; Petri, 2014).

427 In contrast, the Scandinavian Caledonides between Norway and Greenland resulted es-
428 sentially from the closure of the wide ($> 2,000$ km; van Staal et al., 2012) Iapetus Ocean
429 (McKerrow et al., 2000b). At this latitude, the two-sided subduction of this ocean formed
430 at least one major volcanic arc now exposed in Norway (Mac Niocaill et al., 1997; McK-
431 errow et al., 2000b). Note that, further South, the British and Appalachian Caledonides
432 involved many more micro-continents, narrow oceanic tracks and volcanic arcs (see Roberts
433 (2003); Chew and Van Staal (2014) and Figure 1 in van Staal et al. (2012)), comparably
434 to the Variscides of Western Europe (see Figure 2 in Franke (2006)). Thus we restrict our
435 consideration of the Scandinavian Caledonides to the norther part between Norway and
436 Greenland, where accretion of terranes was extremely limited. The Caledonian topography
437 underwent a phase of orogenic collapse, which was essentially achieved through mechanical
438 deformation without significant magmatic activity north of the Elbe lineament (McClay
439 et al., 1986; Andersen, 1998; Meissner, 1999; Fossen et al., 2014).

440 An alternative hypothesis to account for the low magmatic budget of the Caledonian
441 orogenic collapse relies on the depleted composition of the mantle underlying the two con-
442 tinents involved in the orogeny, namely Laurentia and Baltica. Indeed, both are comprised
443 of Archean cratonic cores (the North American and East European craton, respectively),
444 which are characterized by a thick, cold and depleted lithospheric mantle (Bernstein et al.,
445 1998; Griffin et al., 2003; Beyer et al., 2004).

446 The importance of the magmatic event associated with a post-orogenic collapse has
447 direct consequences on the characteristics of the lithosphere, since it may erase all the
448 structural inheritance in the lower crust (Bois et al., 1989; Rey, 1993), introduce major
449 compositional and thermal heterogeneities in the upper and middle crust (Costa and Rey,
450 1995; Vanderhaeghe and Teyssier, 2001), and significantly deplete the underlying mantle
451 (McCarthy and Müntener, 2015). This inheritance is also much more likely to be expressed
452 in subsequent tectonic events, for instance influencing the localization and controlling the
453 magmatic budget of later rifting events. In particular, this could account for the differing
454 behaviour of the North Atlantic rift with respect to the Caledonian and Variscan orogenic
455 lithospheres highlighted by Chenin et al. (2015). It could also explain the variability in
456 the behaviour of the Gondwanan rifts with respect to the former orogens affecting the
457 supercontinent (i.e. paralleling, cutting across or circumventing; see Krabbendam and
458 Barr (2000)). The characteristics of the narrow hyperextended rift systems and wide
459 oceans morphology, lithology, subduction and resulting orogens are summarized in Table 4
460 and Figure 6.

461 5 Conclusion

462 In this paper, we show that each distal rift domain, as defined by Sutra et al. (2013),
463 has a specific range of width. In contrast to the necking and exhumation domains, whose
464 range of width is large, the width of the hyperextended domain is relatively consistent
465 amongst rifted margins (about 50 km). Both the width of the necking domain and of the
466 hyperextended domain are independent of the total width of the margin (i.e. the distance
467 from the coupling point to the lithospheric breakup point) and of the maturity of the rift
468 system (i.e. whether or not steady-state, self-sustaining seafloor spreading is achieved).

469 As ‘narrow oceans’ are usually devoid of mature spreading systems (then called ‘imma-
470 ture’) in contrast to wide, mature oceans, the main difference between these end-members
471 is whether their margins are separated by a wide domain of ‘normal’ oceanic crust. Further-
472 more, narrow and immature ‘oceans’ are likely to be underlain by fertile mantle resulting
473 from melt impregnation during the phase of hyperextension conversely to wide and mature
474 oceans, which have a typical mid-oceanic ridge-type depleted mantle.

475 During the subduction of narrow oceans, the slab is expected to remain at shallow angle
476 (Stevenson and Turner, 1977; Tovish et al., 1978; Billen, 2008). Thus, subduction is very
477 unlikely to become self sustained (Hall et al., 2003; Gurnis et al., 2004) or to develop
478 vigorous small-scale convection in the mantle wedge (Peacock et al., 1994), as opposed to
479 subduction of wide oceans. Furthermore, subduction of narrow oceans does not produce
480 significant magmatic activity, thus no mantle depletion, because insufficient volatiles reach
481 a sufficient depth to allow partial melting (Peacock, 1991; Rüpke et al., 2004). Therefore,
482 hydration is likely to be the dominant process in the mantle wedge.

483 Conversely, protracted subduction associated with the closure of wide oceans develop
484 vigorous convection in the mantle wedge, forms magmatic arcs, and is potentially associated
485 with seafloor spreading in the backarc region (Uyeda, 1981; Peacock et al., 1994). These
486 processes deplete the underlying mantle in fusible elements, create new crustal material
487 and are associated with high temperature/low pressure metamorphism.

488 As a result, orogens resulting from the closure of narrow oceans may be essentially con-
489 trolled by mechanical processes, without significant compositional or thermal perturbation,
490 and with a major influence of the inherited characteristics of the intervening margins. In
491 contrast, orogens produced by the closure of wide oceans may be significantly controlled
492 by subduction-induced processes.

493 Because of the lack of magmatic activity during the closure of narrow oceans, the mantle
494 underlying the resulting orogens is likely more fertile than the mantle underlying orogens

495 due to the closure of wide oceans. This difference in fertility may dictate the magmatic
496 budget of a subsequent extensional event, such as a post-orogenic collapse or an episode of
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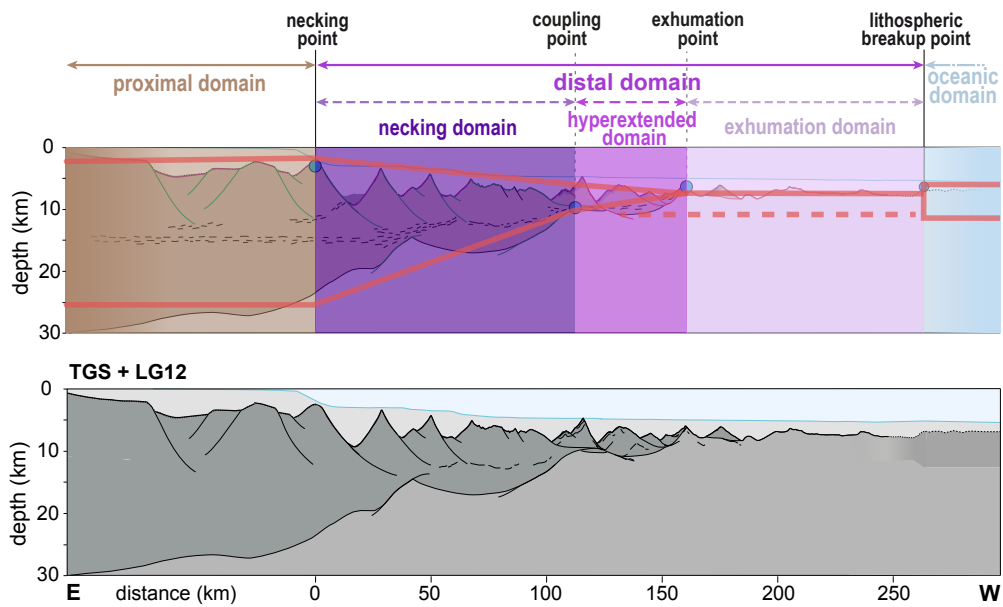


Figure 1: Definition of rift (sub-)domains based on Sutra et al. (2013) and Chenin et al. (2015) and their first-order morphology (red lines). The dashed line represents the ‘seismic Moho’ (i.e. a sharp increase in P velocity from $< 7 \text{ km.s}^{-1}$ to $> 7.8 \text{ km.s}^{-1}$; see Mengel and Kern (1992)), where it differs from the petrologic Moho (i.e. the crust–mantle boundary).

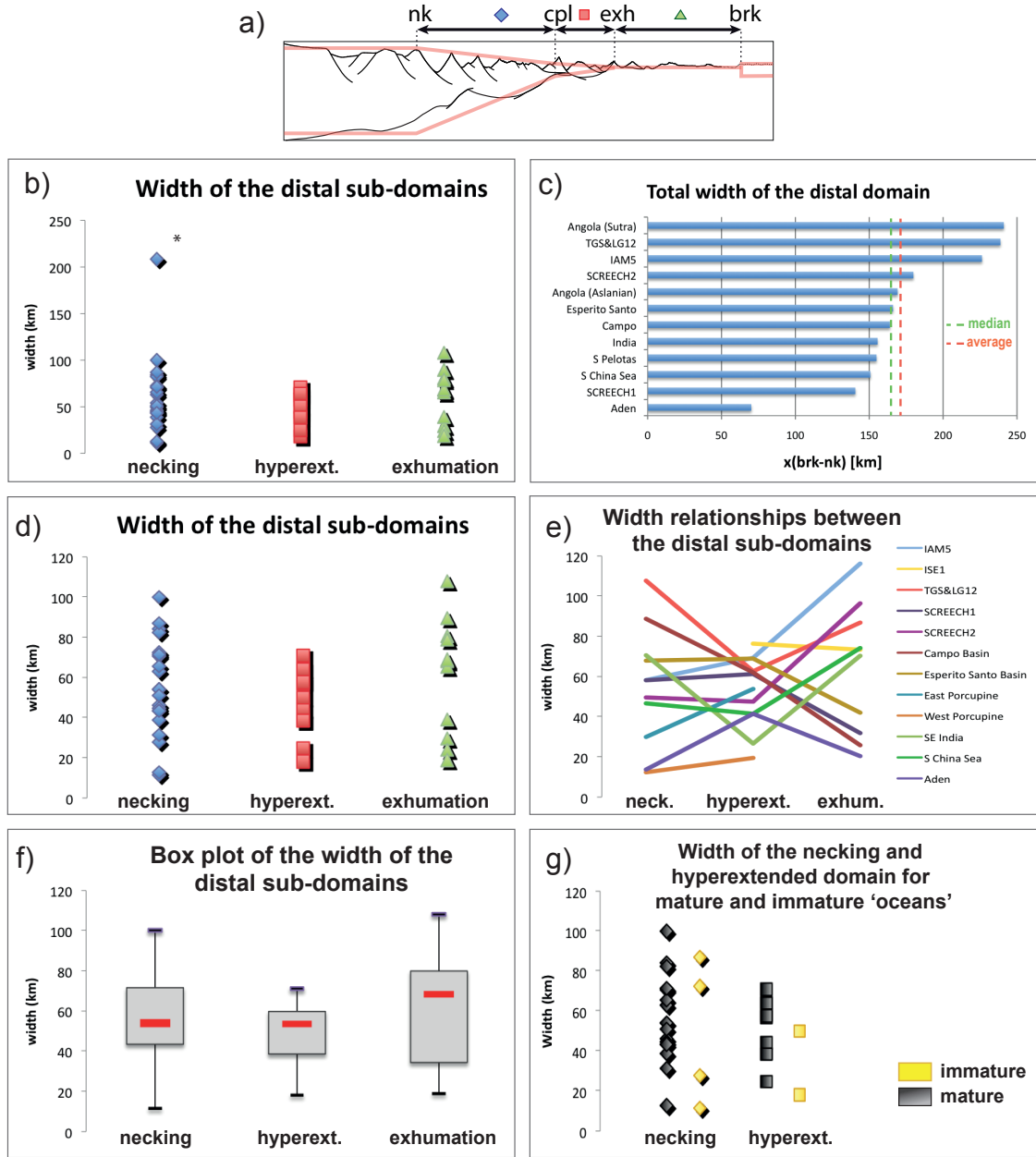


Figure 2: (a) Definition of the necking (blue diamond), hyperextended (red square) and exhumation (green triangle) domains; nk: necking point; cpl: coupling point; exh: exhumation point; brk: lithospheric breakup point. (b) Width of the necking, hyperextended and exhumation domains for all selected margins; (c) Total width of the distal domain for several rifted margins; (d) Width of the necking, hyperextended and exhumation domains without the anomalously wide necking domain of the ISE1 section (see text for discussion); (e) Width relationships between the necking, hyperextended and exhumation domains; (f) Box plot of the width of the necking, hyperextended and exhumation domains; (g) Width of the necking and hyperextended domains for immature and mature oceans.

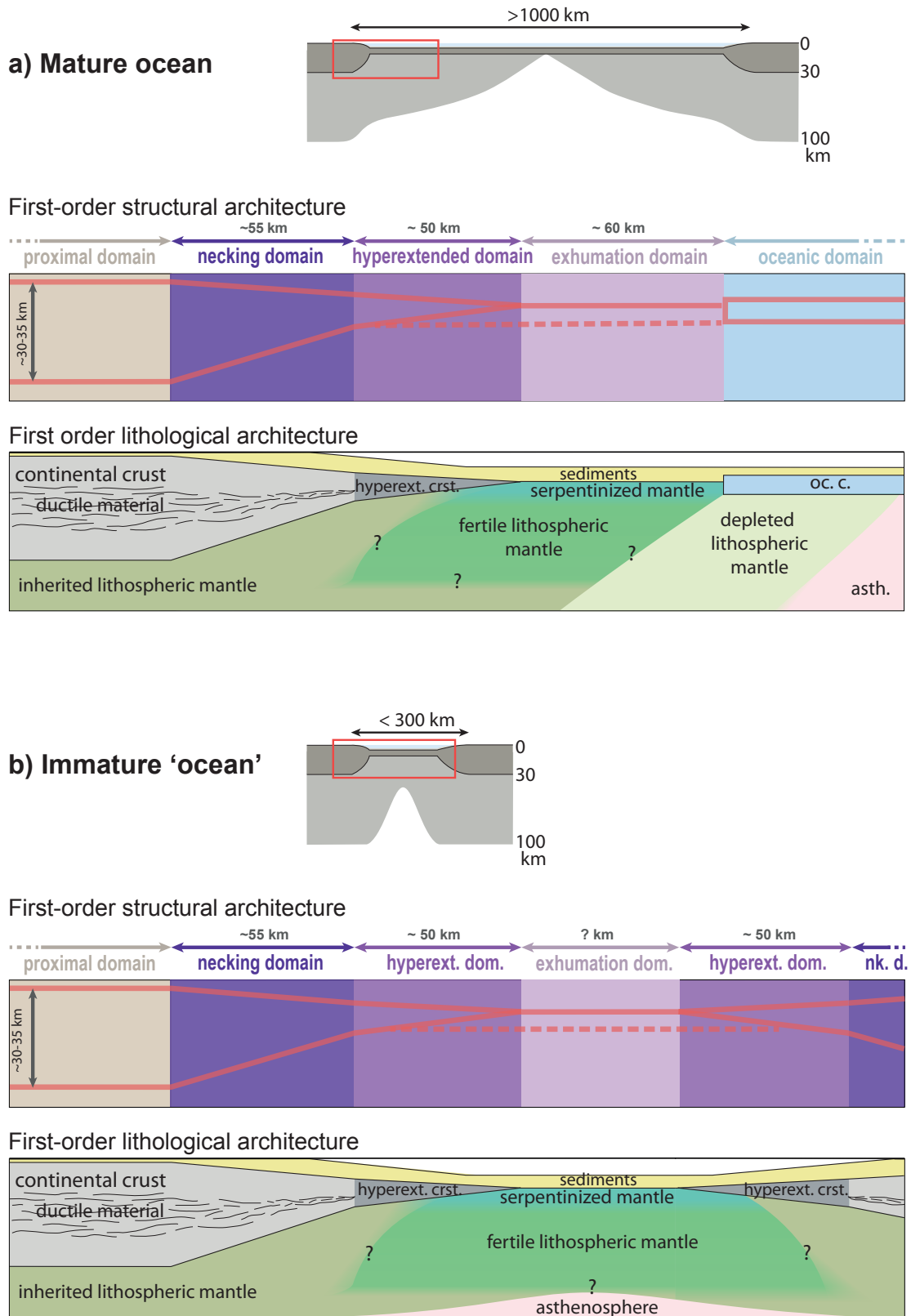


Figure 3: First-order architecture of magma-poor hyperextended rifted margins from (a) wide, mature oceans; and (b) narrow, immature oceans. Abbreviations: nk. d.: necking domain; hyperext. crst.: hyperextended continental crust; oc. c.: oceanic crust; asth.: asthenospheric mantle. See text for discussion.

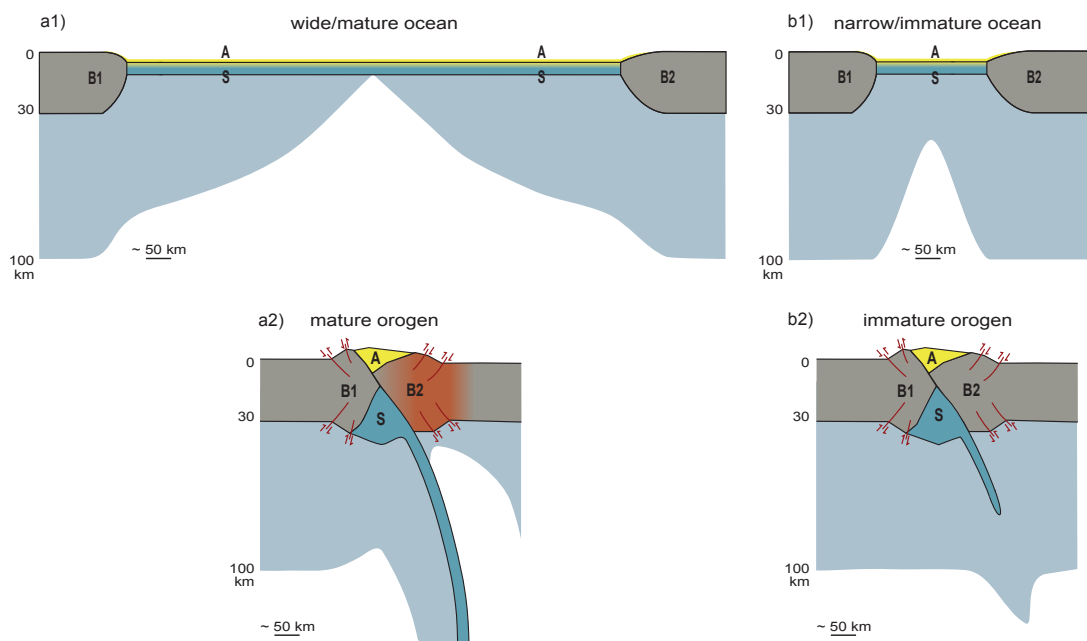


Figure 4: First-order architecture of a (a1) wide mature ocean; and (b1) narrow immature ocean. First-order architecture of (a2) a mature orogen following the subduction of a wide ocean; and (b2) an immature orogen produced by the closure of a narrow ocean; In both cases, continental collision involves two buttresses B (external parts of the orogen), an accretionary prism A (internal part of the orogen) and a subducted part S.

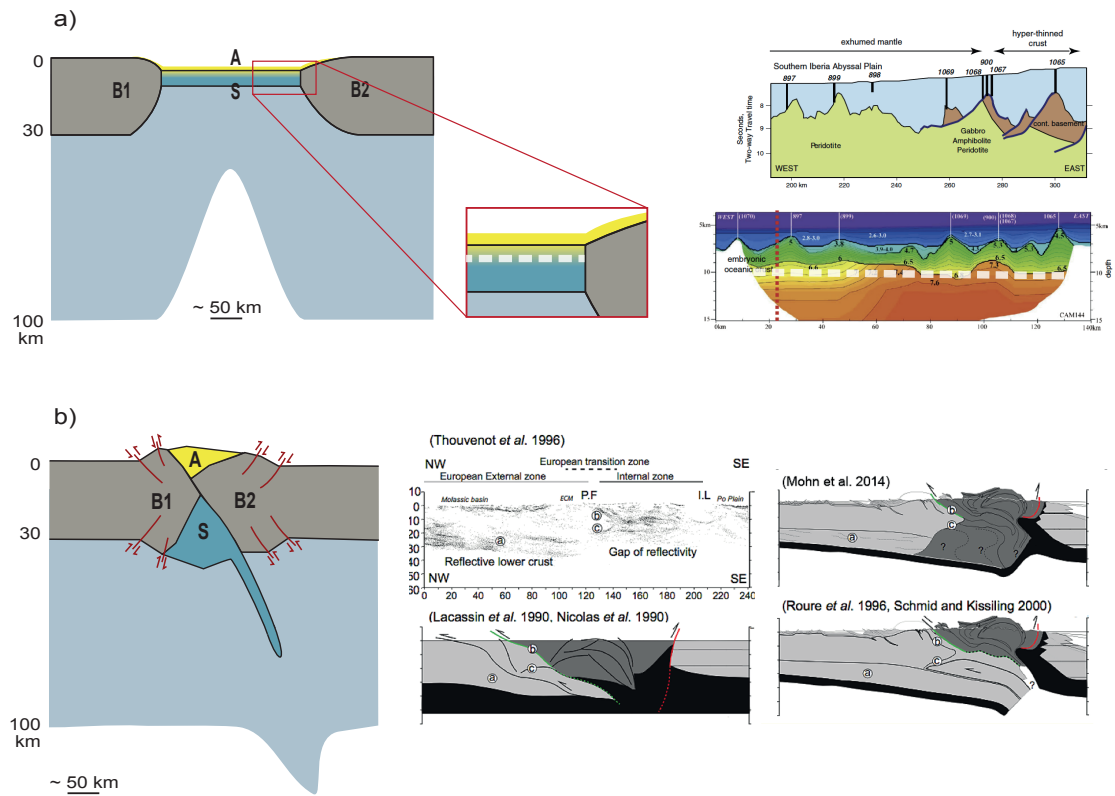


Figure 5: (a) The decoupling level between the subducted and accreted material during subduction corresponds possibly to the hydration front based on the geological interpretation of the reflection Lusigal 12 seismic section and the Velocity model of the adjacent CAM 144 seismic section by Beltrando et al. (2014). (b) Uncertainty about the nature of the deep structure of collisional orogens highlighted by the various interpretations of the ECORS-CROP seismic profile through the Alpine orogen (from Mohn et al., 2014).

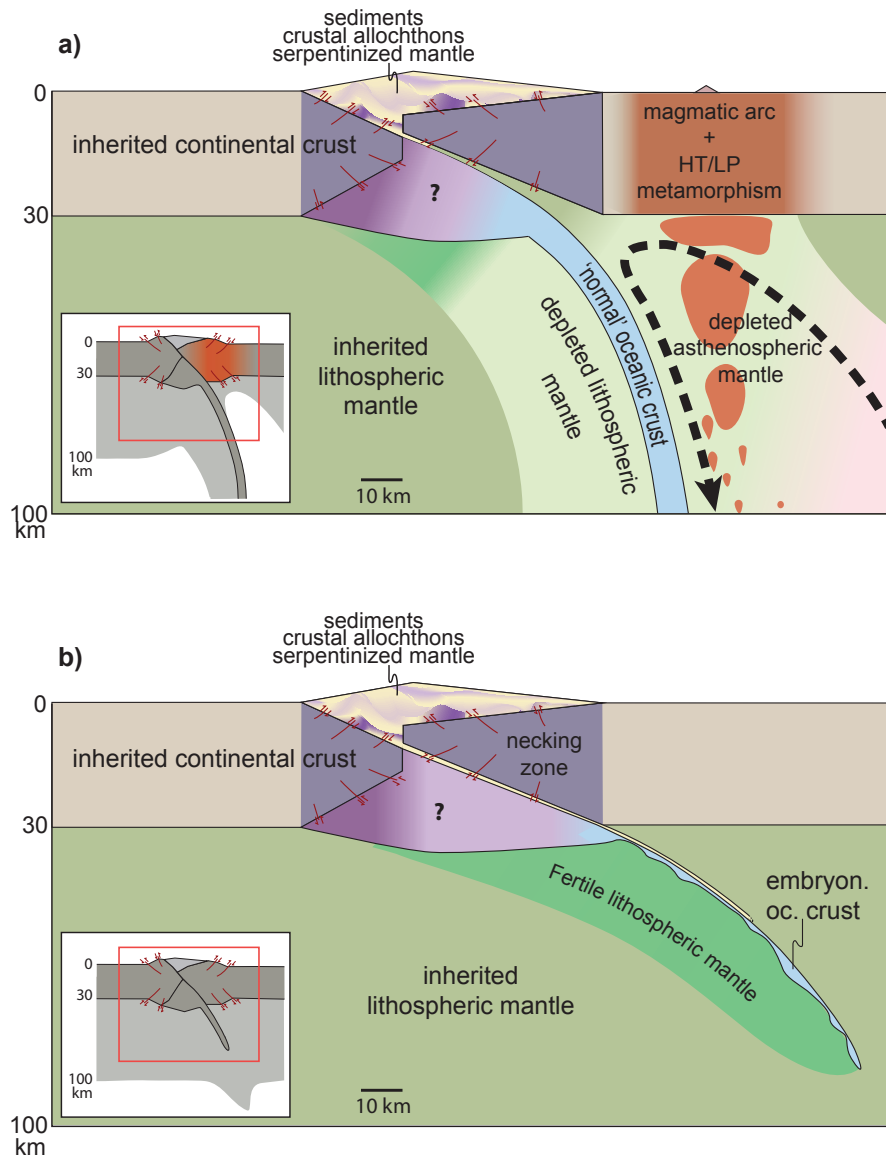
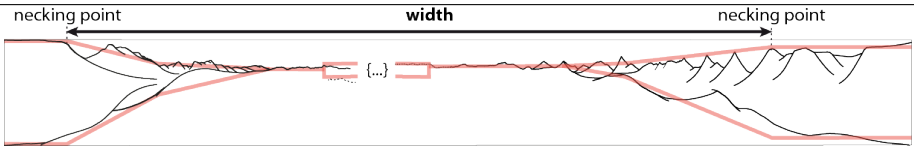
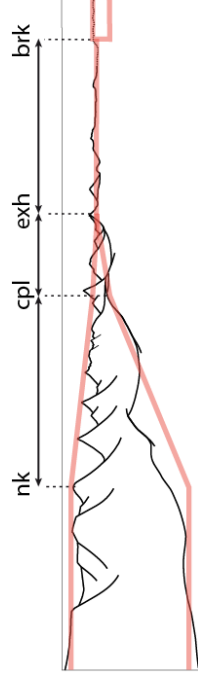
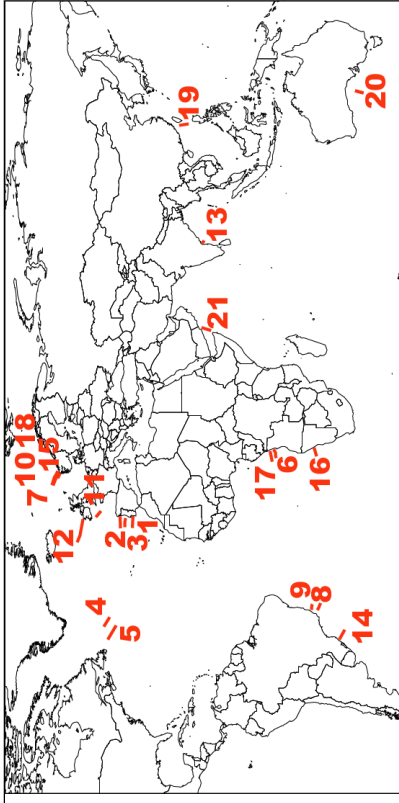


Figure 6: Architecture of orogens following the closure of a (a) wide and mature ocean; and (b) narrow and immature ocean. The inserts highlight the scale of the zones represented. Abbreviations: embryon. oc.: embryonic oceanic crust; HT/LP: high temperature/low pressure.



| | Width | Basin floor | References |
|------------------|--------------------|--|----------------------------|
| Atlantic Ocean | ~ 3,000 - 5,000 km | oceanic crust | |
| Porcupine Basin | < 200 km | thinned cont. crust (& exhumed mantle?) | Reston et al. (2004) |
| Rockall Trough | ~ 250 km | thinned cont. crust | Klingelhöfer et al. (2005) |
| Hatton Basin | ~ 200 km | thinned cont. crust | Edwards (2002) |
| Orphan Basin | ~ 400 km | thinned cont. crust | Chian et al. (2001) |
| Northern Red Sea | 250–350 km | no or embryonic oceanic crust | Bosworth et al. (2005) |
| Southern Red Sea | 350–450 km | no or embryonic oceanic crust | Bosworth et al. (2005) |
| Gulf of Aden | < 200 km | (embryonic?) oc. crust | Bosworth et al. (2005) |
| Indian Ocean | up to ~ 20,000 km | oceanic crust | |
| Pacific Ocean | up to ~10,000 km | oceanic crust | |

Table 1: Width and nature of basement floor of several present-day extensional systems. The *width* refers to the distance between the two conjugate necking points (nk).



| Seismic section | x(cpl-nk) (km) | x(exh-cpl) (km) | x(brk-exh) (km) | x(brk-nk) (km) | Moho and basement linedrawing |
|----------------------------------|-------------------|--------------------|--------------------|-------------------|-------------------------------|
| 1 IAM5 (Afilhado et al., 2008) | 54 | 64 | 108 | 226 | |
| 2 ISE1 (Sutra et al., 2013) | 208* | 71 | 68 | 347 | |
| 3 TGS+LG12 (Sutra et al., 2013) | 100 | 58 | 81 | 239 | |
| 4 SCREECH1 (Sutra et al., 2013) | 54 | 57 | 30 | 141 | |
| 5 SCREECH2 (Sutra et al., 2013) | 46 | 44 | 90 | 180 | |
| 6 Angola (Unterneh et al., 2010) | 71 | | | 241 | |

| Seismic section | x(cpl-nk) | x(exh-cpl) | x(brk-exh) | x(brk-nk) | Moho and basement linedrawing |
|---|-----------|------------|------------|-----------|-------------------------------|
| 7 Norway (TrII Nirrengarten et al., 2014) | 84 | | | | |
| 8 Campos Basin (Zalán et al., 2012) | 82 | 58 | 24 | 164 | |
| 9 Esperito Santo Basin (Zalán et al., 2012) | 63 | 64 | 39 | 166 | |
| 10 Norway (Osmundsen and Ebbing, 2008) | 45 | | | | |
| 11 E-Porcupine (McDermott et al., 2014) | 28 | 50 | | | |
| 11 W-Porcupine (McDermott et al., 2014) | 11 | 18 | | | |
| 12 E-Rockall (Welford et al., 2010) | 72 | | | | |
| 12 W-Rockall (Welford et al., 2010) | 87 | | | | |
| 13 India (Radhakrishna et al., 2012) | 65 | 25 | 65 | 156 | |
| 14 S-Pelotas (Stica et al., 2014) | 71 | | | | |
| 15 Norway (Kvarven, 2013) | 43 | | | | |
| 16 Namibia (Gladzenko et al., 1998) | 51 | | | | |

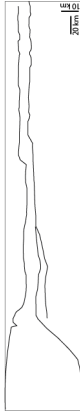
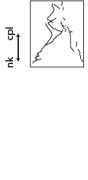
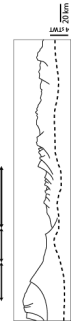


| Seismic section | x(cpl-nk) | x(exh-cpl) | x(brk-nk) | Moho and basement linedrawing | |
|--|-----------|------------|-----------|---|---|
| 17 Angola (Aslamian et al., 2009) | 39 | | 169 |  | |
| 18 Norway (Osmundsen and Ebbing, 2008) | 31 | | |  | |
| 19 China Sea (Lester et al., 2014) | 43 | 39 | 69 | 151 |  |
| 20 South Australia (Direen et al., 2008) | | | 79 | |  |
| 21 Aden (Leroy et al., 2010) | 13 | 39 | 19 | 70 |  |

Table 2: Width of the marginal domains for several hyperextended to oceanic rift systems. nk: necking point; cpl: coupling point; exh: exhumation point; brk: breakup point (see Figure 1).

| | | | | | | | | | |
|-------------|--------------|----------------|--------------|--------------|----------------|--------------|--------------|----------------|--------------|
| temp. (°C) | 700 | | | 800 | | | 900 | | |
| mantle type | harz. | fertile lherz. | | harz. | fertile lherz. | | harz. | fertile lherz. | |
| P (GPa) | (plg) | | | (plg) | | | (plg) | | |
| 0.5 | 3.260 | 3.209 | | 3.248 | 3.197 | (spl) | 3.235 | 3.185 | (spl) |
| 1 | 3.275 | 3.224 | (spl) | 3.263 | 3.213 | 3.276 | 3.250 | 3.201 | 3.264 |
| 1.5 | 3.295 | 3.303 | | 3.283 | 3.291 | | 3.271 | 3.279 | |
| 2 | 3.309 | 3.317 | | 3.297 | 3.306 | | 3.300 | (grt) | 3.310 |
| 2.5 | 3.323 | (grt) | 3.331 | 3.311 | (grt) | 3.320 | 3.299 | 3.332 | 3.308 |
| 3 | 3.336 | 3.368 | | 3.325 | 3.357 | | 3.313 | 3.346 | |
| temp. (°C) | 1000 | | | 1100 | | | 1200 | | |
| mantle type | harz. | fertile lherz. | | harz. | fertile lherz. | | harz. | fertile lherz. | |
| P (GPa) | (plg) | | | (plg) | | | (plg) | | |
| 0.5 | 3.222 | 3.173 | (spl) | 3.209 | 3.161 | (spl) | 3.195 | 3.148 | (spl) |
| 1 | 3.237 | 3.189 | 3.251 | 3.225 | 3.177 | 3.239 | 3.212 | 3.165 | 3.226 |
| 1.5 | 3.258 | 3.267 | | 3.246 | 3.254 | | 3.233 | 3.242 | |
| 2 | 3.273 | (grt) | 3.282 | 3.261 | (grt) | 3.270 | 3.249 | (grt) | 3.258 |
| 2.5 | 3.288 | 3.320 | 3.297 | 3.276 | 3.309 | 3.285 | 3.264 | 3.297 | 3.273 |
| 3 | 3.300 | 3.334 | | 3.288 | 3.323 | | 3.277 | 3.312 | |
| temp. (°C) | 1300 | | | | | | | | |
| mantle type | harz. | lherz. | | | | | | | |
| P (GPa) | | | | | | | | | |
| 0.5 | | | | | | | | | |
| 1 | | | | | | | | | |
| 1.5 | | | | | | | (spl) | | |
| 2 | 3.236 | 3.246 | | | | | | | |
| 2.5 | 3.250 | 3.262 | | | | | | | |
| | | | | | | | (grt) | | |
| 3 | 3.265 | 3.300 | | | | | | | |

Table 3: Density in the mantle as a function of temperature, pressure and composition calculated with the algorithms of Hacker et al. (2003). Depleted harzburgite (harz.): 80% olivine, 20% opx, both with a Mg# of 90-91; Fertile plagioclase peridotite (plg fertile lherz.): 9% plagioclase (An80), 60% olivine (Fo90), 24% opx (Mg# 90), 7% cpx (Mg# 90); Fertile spinel peridotite (spl fertile lherz.): 2% spinel (spinel, hercynite, magnetite), 58% olivine (Fo90), 26% opx (Mg# 90), 14% cpx (Mg# 90); Fertile garnet peridotite (grt fertile lherz.): 9% garnet (pyrope 66%, almandin 24%, grossular 10%), 60% olivine (Fo89), 24% opx (Mg#89), 7% cpx (Mg# 90).

| | Narrow ‘ocean’ | Wide ocean |
|---------------------------|---|---|
| Spatial extent | \lesssim 350–400 km | $>$ 1,000 km |
| Oceanic crust | unsteady seafloor spreading \Rightarrow rough, heterogeneous in thickness and composition; no mantle depletion | steady-state seafloor spreading \Rightarrow smooth, homogeneous in thickness and composition; depleted underlying mantle |
| Magmatic activity | none or very little \Rightarrow no mantle depletion | moderate to large \Rightarrow mantle depletion |
| Subduction geometry | shallow angle | shallow or deep angle |
| Subduction sustainability | transitory | self-sustained |
| Mantle wedge convection | minor | vigorous |
| Magmatic arc | none | yes; new over-thickened crust creation and associated HT metamorphism and underlying mantle depletion |
| backarc basin | none | possible; may be associated with seafloor spreading and underlying mantle depletion |
| Orogen type | collisional | accretionary followed by collisional |
| Mantle wedge composition | hydrated and fertilized with sediments | Depleted in fusible elements |
| Post-orogenic collapse | possibly magmatic | purely extensional |

Table 4: Summary of the characteristics of narrow versus wide extensional systems, subduction processes and orogens (see text for discussion).