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Influence of the architecture of magma-poor hyperextended rifted margins on orogens produced by the closure of narrow versus wide oceans

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ABSTRACT

Orogens resulting from the closure of narrow oceans, such as the Alps or the Pyrenees, usually lack voluminous synsubduction and synorogenic magmatism. Such orogenies are essentially controlled by mechanical processes in which the initial architecture of the original rifted margins strongly controls the architecture of the orogen. In this paper we first provide a synthesis of the structure, dimensions, and lithology of hyperextended rift systems and oceans, based on recent seismic and petrologic data. We then investigate how rift-related inheritance influences crustal characteristics and mantle geochemistry of orogens related to the closure of narrow oceans, and compare them to orogens resulting from the closure of wide and/or mature oceans. Our results show that narrow oceans usually lack a mature spreading system forming Penrose-type oceanic crust (i.e., 6–7-km-thick basaltic oceanic crust typical of steady-state spreading systems; see Anonymous, 1972), in contrast to wide oceans. However, there is statistically no difference in the structural and lithological architecture of their passive continental margins. Thus, the main difference between narrow and wide oceans is whether the margins are separated by a significant amount of oceanic crust and underlying depleted mantle. In addition, due to the lack of significant magmatism during the closure of narrow oceans, the mantle wedge is likely to remain relatively fertile compared to the wedge above long-lasting subduction of wide oceans. This difference in mantle composition may dictate the magmatic budget of subsequent orogenic collapse or rifting events.

INTRODUCTION

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

This paper aims to provide a synthesis of the primary characteristics of hyperextended rift systems and oceans at a scale and resolution compatible with lithospheric-scale thermomechanical numerical codes. Based on recent seismic and petrologic data, we synthesize the structure and lithology of magma-poor hyperextended rifted margins and oceans, and provide a compilation of the dimensions of the different structural domains composing rifted margins. We compare the characteristics between narrow and/or immature and wide and/or mature oceans. We define narrow oceans as rift systems that reached at least the stage of hyperextension (for definitions and reviews of hyperextension, see Sutra et al., 2013; Doré and Lundin, 2015), but remained <~300 km wide (the reasons for this limit are discussed in the following). We use the term mature for oceans that comprise a self-sustaining, steady-state seafloor-spreading system, as opposed to immature or embryonic oceans, whose development stopped at the stage of hyperextension or exhumation. We then investigate how the primary rift-related inheritance of narrow oceans may influence orogens resulting from their closure, and compare these immature orogens with classic mature orogens produced by the closure of wide and/or mature oceans (Uyeda, 1981).

CHARACTERISTICS OF MAGMA-POOR HYPEREXTENDED RIFTED MARGINS

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

Recent studies suggest that most rifted margins whose necking and hyperextension phases were magma poor display a similar succession of comparable domains (Fig. 1), regardless of whether they achieved lithospheric breakup and steady-state seafloor spreading (Sutra et al., 2013; Péron-Pinvidic et al., 2013). In this discussion we characterize the primary architecture and lithology of such rift systems and quantify the width of their distal subdomains based on natural examples.

Primary Architecture

Since the extensive seismic and drilling surveys of the 1990s, rifted margins are no longer regarded as a simple series of tilted blocks adjacent to Penrose-type oceanic crust (i.e., 6–7-km-thick basaltic oceanic crust typical of steady-state spreading systems; see Anonymous, 1972). The drilling of exhumed mantle offshore Iberia led to the reconsideration of the architecture of rifted margins (Boillot et al., 1980). In order to characterize the architecture of hyperextended rifted margins, Sutra et al. (2013) distinguished several domains based on morphological criteria (Fig. 1). On the one hand, the proximal domain corresponds to unthinned or minor extended (~30–35 km thick) continental crust and is characterized by parallel, roughly flat basement and Moho topographies. On the other hand, a typical oceanic domain is made of homogeneous, Penrose-type oceanic crust, ~6–7 km thick. Here again, basement and Moho are parallel. The distal domain, between them, records most of the rift-related deformation.

The distal domain can be divided into several subdomains (Fig. 1): (1) a necking domain characterized by the abrupt thinning of the continental crust from ~30–35 km down to ~10 km that translates to a deepening of top basement and shallowing of the Moho (the latter only on seismic sections in depth); (2) a hyperextended domain, where continental crust is thinned from ~10 km down to 0 km. The transition between the necking and hyperextended domain corresponds to a sudden decrease in the dip of the Moho on depth seismic sections. (3) A so-called

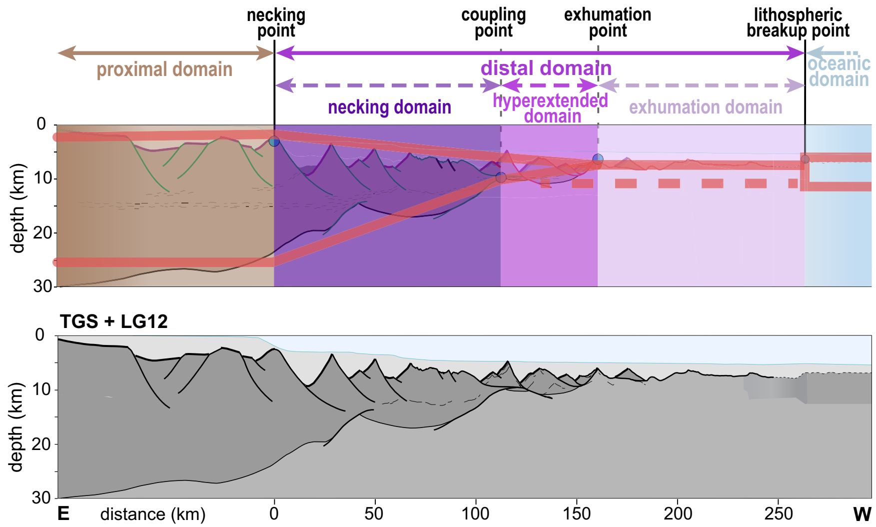
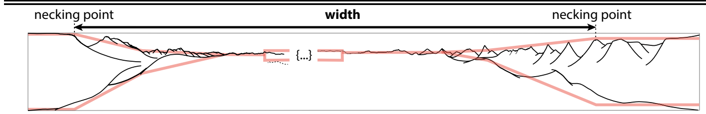


Figure 1. Definition of rift domains or subdomains based on Sutra et al. (2013) and Chenin et al. (2015) and their primary morphology (red lines). The dashed line represents the seismic Moho (i.e., a sharp increase in P-wave velocity from <7 km s^{-1} to >7.8 km s^{-1}; see Mengel and Kern, 1992), where it differs from the petrologic Moho (i.e., the crust-mantle boundary). LG12: see Beslier (1996). TGS: credit TGS-NOPEC.

exhumation domain may exist in magma-poor rifted margins when lithospheric subcontinental mantle is exhumed at the seafloor (Boillot et al., 1987; Whitmarsh et al., 2001; Manatschal and Müntener, 2009). Mantle exhumation is associated with serpentinization down to a depth of 4–6 km (Boillot et al., 1989; Escartín et al., 2001; Minshull et al., 1998), and the transition from serpentinized mantle to fresh peridotite translates to a progressive increase in seismic velocity (Horen et al., 1996; Miller and Christensen, 1997; Skelton et al., 2005). Mantle exhumation may also be accompanied by the emplacement of discontinuous extrusive magmatic rocks from an immature spreading system, which is characterized by incomplete development of oceanic layers 2 and 3 (e.g., Desmurs et al., 2002; Müntener et al., 2004; Jagoutz et al., 2007). Note that allochthonous blocks of continental crust may be found on top of the serpentinized exhumed mantle of the exhumation domain. Thus, on seismic sections, the exhumation domain appears as a more or less structured surface either devoid of a Moho reflector, or with a discontinuous seismic Moho reflector when accompanied with magmatism (Whitmarsh et al., 2001). The transition from the exhumation domain into the oceanic domain is highlighted by an abrupt step up in the basement (Bronner et al., 2011), which reflects the decrease in the density of the lithospheric column resulting from the production of thicker oceanic crust through increased magmatic activity. In contrast, in the case of magma-rich margins, the transition from thickened magmatic crust to steady-state seafloor spreading is expressed as a step down onto oceanic crust. The primary morphology of magma-poor rifted margins is highlighted by the red lines in Figure 1.

TABLE 1. WIDTH AND NATURE OF BASEMENT FLOOR OF SEVERAL PRESENT-DAY EXTENSIONAL SYSTEMS



	Width (km)	Basin floor	Source
Atlantic Ocean	~3000–5000	oceanic crust	
Porcupine Basin	<200	thinned continental crust (and exhumed mantle?)	Reston et al. (2004)
Rockall Trough	~250	thinned continental crust	Klingelhöfer et al. (2005)
Hatton Basin	~200	thinned continental crust	Edwards (2002)
Orphan Basin	~400	thinned continental crust	Chian et al. (2001)
Northern Red Sea	250–350	no or embryonic oceanic crust	Bosworth et al. (2005)
Southern Red Sea	350–450	no or embryonic oceanic crust	Bosworth et al. (2005)
Gulf of Aden	<200	(embryonic?) oceanic crust	Bosworth et al. (2005)
Indian Ocean	as much as ~20,000 km	oceanic crust	
Pacific Ocean	as much as ~10,000 km	oceanic crust	

Dimensions and Maturity of Rift Systems

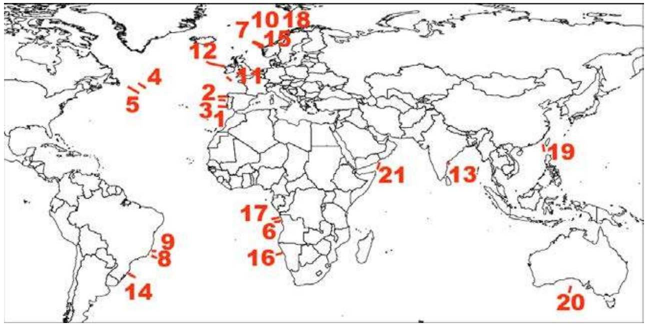
In order to assess the dimensions of rift systems and their distal subdomains, in addition to their maturity, we first compile the width and the lithology of basement rocks for several narrow and wide extensional systems around the world (Table 1). In this paper the width of an extensional system refers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

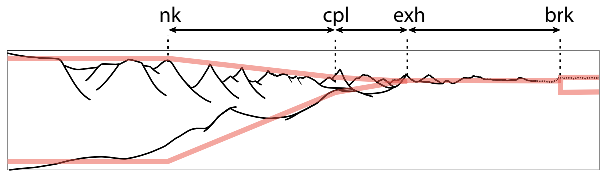
Second, we measure the width of the distal subdomains (namely the necking, hyperextended, and exhumation domains) for a selection of published seismic sections (Table 2). For this compilation, we only consider dip seismic lines imaging extensional systems formed in a single, unidirectional, and continuous rifting event (i.e., no significant time lag without extension). Note that the necking domain of the ISE1 seismic section (Zelt et al., 2003; see also interpretation by Sutra et al., 2013) seems anomalously wide compared to the other data points (see Fig. 2B). Its unique architecture displaying a double neck (see seismic section 2 in Table 2) suggests that it did not form in a single extensional event, since deformation tends to become more and more localized as extension progresses. Therefore, we remove this data point in the statistical diagrams of Figure 2 (C–G) but we include it in our discussion with an asterisk.

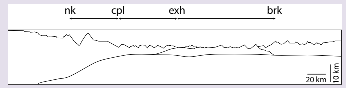
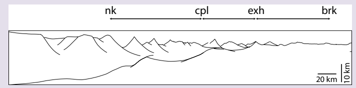
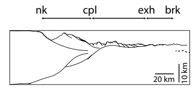
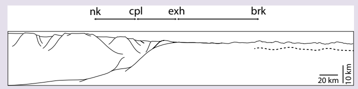
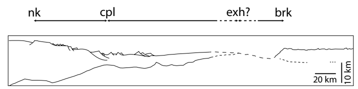
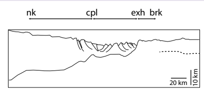
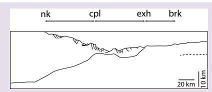
In addition, we restrict our measurements to domains with minor magmatic additions. Note that we did not include the width of the exhumation domain for rift basins that did not achieve lithospheric breakup, because in this case the full width of the exhumation domain is not realized.

Figure 2C shows that the width of the distal domain ranges from ~130 to 240 km (or to *350 km), with an average of 170 km (*185 km) and a median of 165 km (*165 km). Of all the distal subdomains, the necking and exhumation domains have the highest variability in width (see Figs. 2D, 2F). The width of the necking domain ranges from ~10 to 100 km (or to *210 km) with an average ~55 km (*60 km), a median ~50 km, and a standard deviation of 25 km (*40 km). The width of the exhumation domain varies between ~20 km and 110 km, with an average of 60 km, a median of ~70 km, and a standard deviation of ~30 km. In contrast, the width of the hyperextended domain seems more constant, varying between 20 and 70 km and with an average of 50 km (median ~60 km; standard deviation 15 km). In addition, Figure 2E indicates that no correlation exists between the width of any of the distal subdomains and the total width of the distal domain. This means that magma-poor hyperextended rifted margins, and hence each of their distal subdomains, have a specific range of width and can therefore be described with an average structural architecture (see the middle panel of Fig. 3A).

TABLE 2. WIDTH OF MARGINAL DOMAINS FOR SEVERAL HYPEREXTENED TO OCEANIC RIFT SYSTEMS

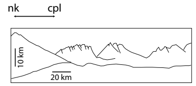
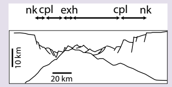
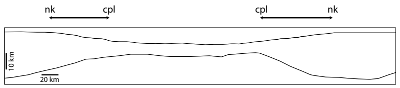
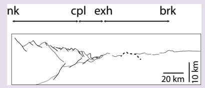
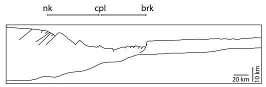
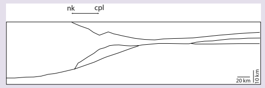
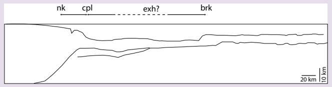
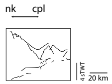
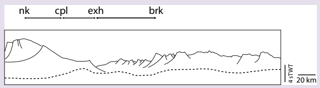
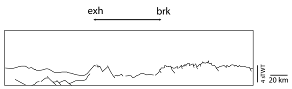
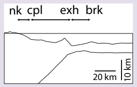




Section, Source	x(cpl-nk) (km)	x(exh-cpl) (km)	x(brk-exh) (km)	x(brk) (km)	Moho and basement line drawing
1 IAM5 (Afilhado et al., 2008)	54	64	108	226	
2 ISE1 (Sutra et al., 2013)	208*	71	68	347	
3 TGS+LG12 (Sutra et al., 2013)	100	58	81	239	
4 SCREECH1 (Sutra et al., 2013)	54	57	30	141	
5 SCREECH2 (Sutra et al., 2013)	46	44	90	180	
6 Angola (Unternehr et al., 2010)	71			241	
7 Norway (Nirrengarten et al., 2014)	84				
8 Campos Basin	82	58	24	164	
9 Espírito Santo Basin	63	64	39	166	

(*continued*)

TABLE 2. WIDTH OF MARGINAL DOMAINS FOR SEVERAL HYPEREXTENED TO OCEANIC RIFT SYSTEMS (*continued*)

Section, Source	x(cpl-nk) (km)	x(exh-cpl) (km)	x(brk-exh) (km)	x(brk) (km)	Moho and basement line drawing
10 Norway (Osmundsen and Ebbing, 2008)	45				
11 East Porcupine (McDermott, 2014)	28	50			
11 West Porcupine (McDermott, 2014)	11	18			
12 East Rockall (Welford et al., 2010)	72				
12 West Rockall (Welford et al., 2010)	87				
13 India (Radhakrishna et al., 2012)	65	25	65	156	
14 South Pelotas (Stica et al., 2014)	71			155	
15 Norway (Kvarven, 2013)	43				
16 Namibia (Gladczenko et al., 1998)	51				
17 Angola (Aslanian 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	

Note: nk—necking point; cpl—coupling point; x(cpl-nk)—distance between cpl point and nk point; exh—exhumation point; brk—breakup point (see Fig. 1).

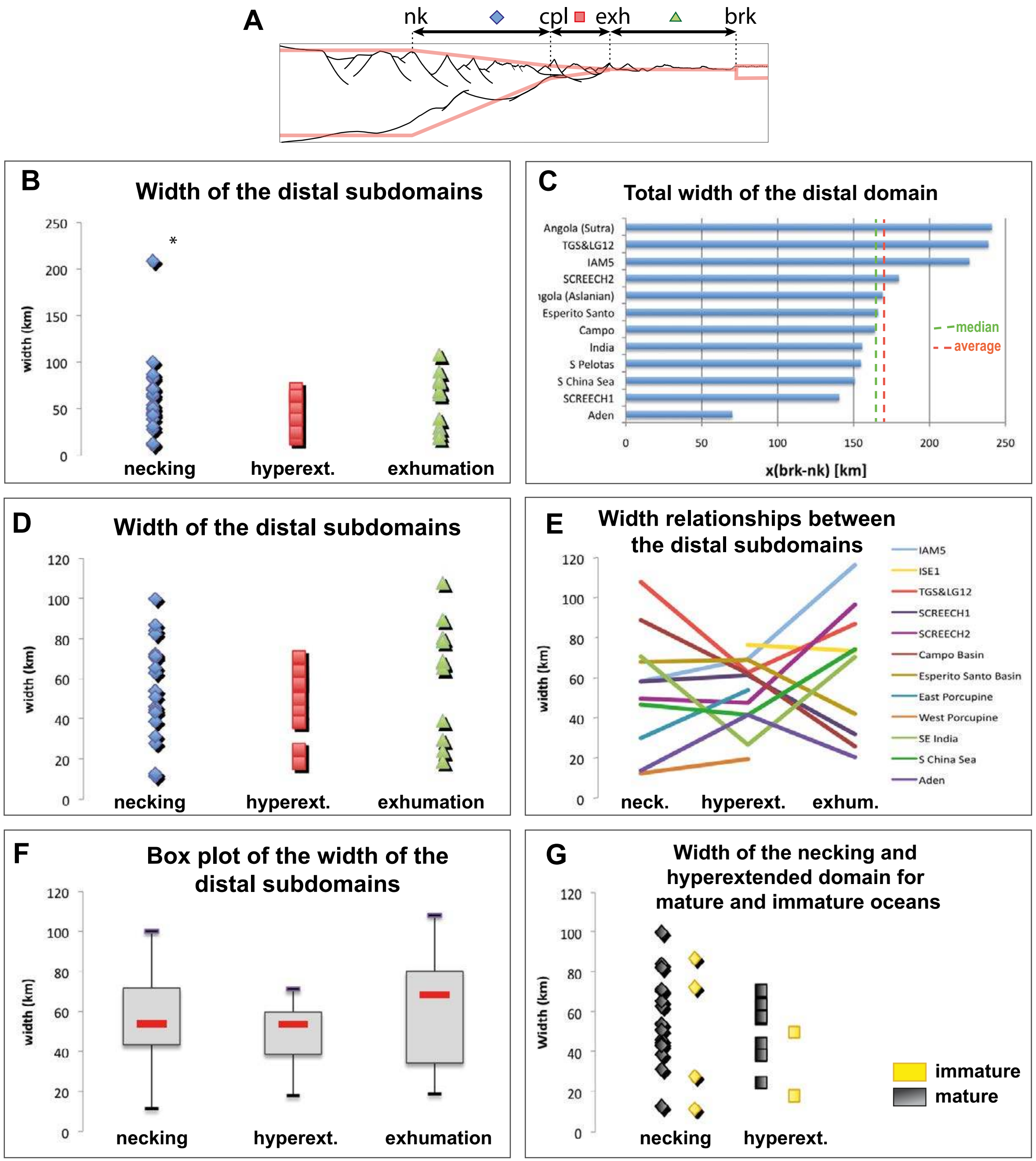


Figure 2. (A) Definition of the necking (blue diamond), hyperextended (red square), and exhumation (green triangle) domains; nk—necking point; cpl—coupling point; exh—exhumation point; brk—lithospheric breakup point. (B) Width of the necking, hyperextended, and exhumation domains for all selected margins. (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 (Zelt et al., 2003) 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.

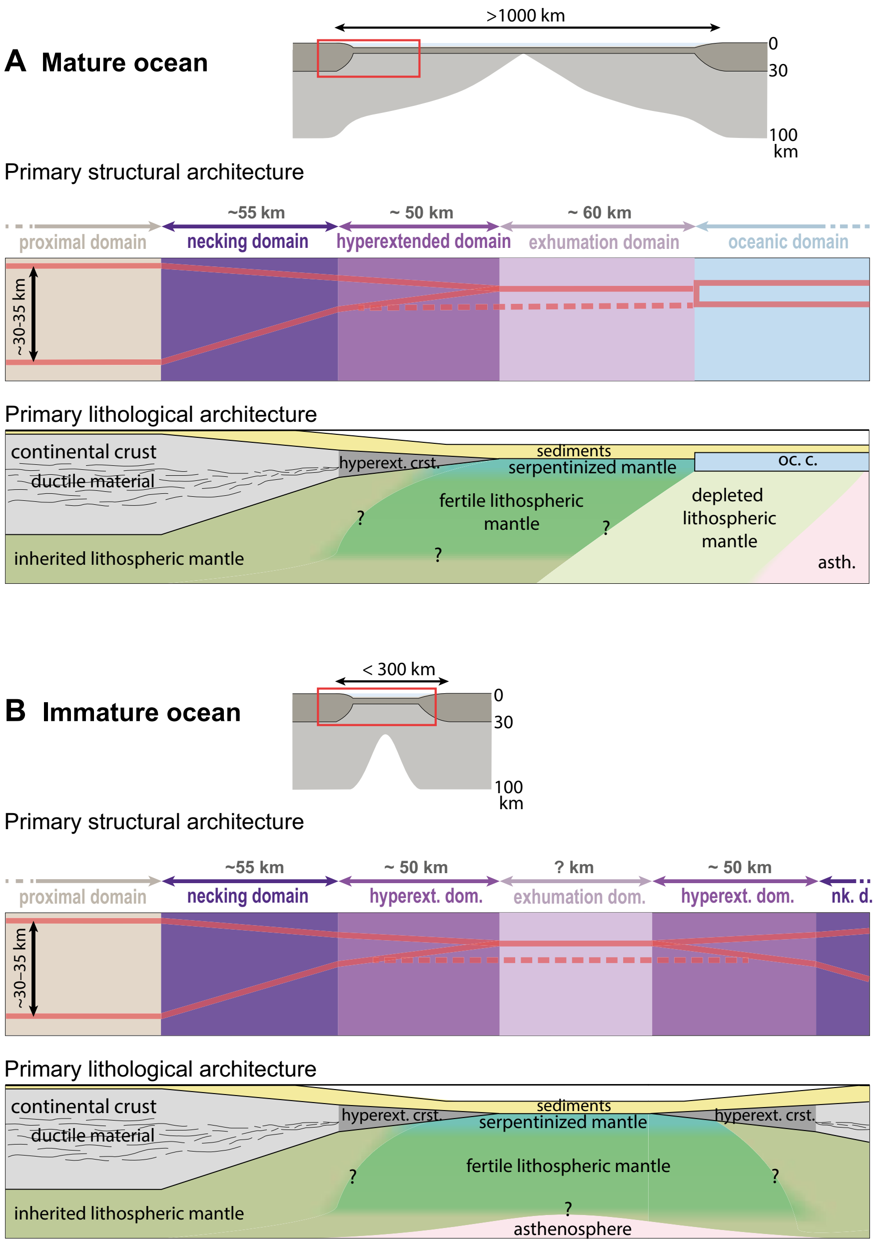


Figure 3. (A) Primary architecture of magma-poor hyperextended rifted margins from wide, mature oceans. (B) From narrow, immature oceans. Abbreviations: hyperext. crst.—hyperextended continental crust; oc. c.—oceanic crust; asth.—asthenospheric mantle; nk. d.—necking domain. See text for discussion.

Figure 2G suggests that the width of both the necking and hyperextended domains is unrelated to the maturity of the extensional system and thus to whether it is a narrow or a wide ocean. Therefore, the distal subdomains of narrow and/or immature oceans have a structural architecture similar to those of wide and/or mature oceans (cf. the middle panels of Figs. 3A and 3B).

Primary Lithological Architecture

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

When rifting is not triggered by magmatism (e.g., upwelling of a mantle plume), the lithology of the crust is not significantly modified by extension in the proximal domain and in the necking domain. Therefore, the continental crust of both domains can be approximated by a quartzofeldspathic material, as usually considered by numerical modelers. Note that, in the necking domain, the ductile layers are mechanically attenuated during extension, so that the crust is fully brittle in the hyperextended domain (Pérez-Gussinyé et al., 2003; Sutra et al., 2013; Manatschal et al., 2015; see the bottom panel of Figs. 3A, 3B). In addition, the subcontinental mantle underlying the proximal and necking domains is not significantly modified and can be approximated by an average inherited mantle consisting of peridotite with highly variable composition [~50%–70% olivine (ol), ~1%–30% clinopyroxene (cpx), ~20%–30% orthopyroxene (opx), and ~1%–4% spinel (sp); Müntener et al., 2010; Picazo et al., 2016], and pyroxenite (ol as low as 0%, cpx to 80%, opx to 60%, and 1%–2% spinel and/or garnet).

In contrast, in hyperextended and exhumation domains, fluid-rock interactions form hydrous minerals, similar to the fluid-rock interactions observed in present-day mid-oceanic ridges (Mével, 2003; Bach et al., 2004; Boschi et al., 2006; Picazo et al., 2012, 2013). Hydrothermal circulation is responsible for the formation of hydrous minerals within hyperextended crust (e.g., sericite and illite; Pinto et al., 2015, and references therein), as well as in exhumed mantle (e.g., serpentinite, chlorite, talc; see Hess, 1955; Christensen, 1970; Früh-Green et al., 2004; Picazo et al., 2013). Therefore, the rheology of the hyperextended continental crust and the top 4–6 km of lithospheric mantle in the exhumation domain can be approximated by a phyllosilicate-type rheology.

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

In contrast, during lithospheric breakup, melt starts to be extracted and a sustainable magmatic system is established, marking the onset of steady-state seafloor spreading. The resulting oceanic crust is homogeneous in both composition and thickness (basaltic to gabbroic and ~6–7 km thick; Anonymous, 1972). The process of partial melting associated with the creation of oceanic crust depletes the underlying mantle in the most fusible elements, leaving an average composition of 57% ol, 13% cpx, 28% opx, and 2% sp (i.e., a depleted mid-oceanic ridge basalt, MORB, mantle; see Workman and Hart, 2005; Fig. 3A). Note that, even in the case of a so-called magma-poor rifting, onset of steady-state seafloor spreading is presumably triggered by a magmatic pulse (Bronner et al., 2011). Therefore, we expect the mantle underlying the most distal part of the margins of mature oceans to be depleted in fusible elements, converse to the mantle underlying immature oceans.

We calculated the density of fertile and depleted mantle between 700 °C and 1300 °C in steps of 100 °C, from 0.5 to 3 GPa (Table 3), based on their average modal composition (Müntener et al., 2010; McCarthy and Müntener, 2015), their physical properties (Hacker et al., 2003), and the geotherm of each mantle-type domain (Sclater et al., 1980). At 0.5 GPa, calculations show that fertile plagioclase peridotite is less dense than depleted harzburgite. From 1.0 to 2 GPa, depleted mantle is lighter by ~20–30 kg m^{-3} compared to fertile mantle. Note that the peridotites considered in this calculation are relatively rich in Fe, but if they were richer in Mg, the density contrast would be ~50–80 kg m^{-3}. At high pressures, because of the higher modal abundance of garnet in fertile rocks, the fertile peridotite is denser than the residual harzburgite by ~30–40 kg m^{-3}.

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

PRIMARY ARCHITECTURE OF COLLISIONAL OROGENS

Every ocean will eventually undergo subduction and be sutured after collision of its continental margins. A collisional orogen can be regarded primarily as a three-part system comprising (Figs. 4A, 4B) (1) two buttresses, (2) an accretionary wedge, and (3) a subducted part.

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

Oceanic lithosphere, due to its high density, tends to be efficiently subducted. According to Stern (2004), most of the ophiolites preserved within orogens correspond to obducted remnants of buoyant oceanic crust from small and young oceanic basins, usually former forearcs or backarcs, rather than mature oceanic crust. A significant proportion of sediments is also usually subducted, while the remaining part accumulates in the accretionary prism (Clift et al., 2004; Stern, 2011, and references therein). A significant proportion of the subducted material has to be integrated to the orogenic root in order to account for the isostasy of collisional orogens; however, the deep architecture of collisional orogens is very poorly constrained (Butler, 2013), as illustrated by the diversity in the interpretation of the deep part of the ECORS-CROP (Nicolas et al., 1990) seismic section (Fig. 5B).

Both the Alpine and the Pyrenean orogens display a similar architecture, where the external domain is made of little-deformed continental basement and the internal part of a complex stacking of material originating from the distal margin (De Graciansky et al., 2011; Bellahsen et al., 2014; Butler et al., 2006; Butler, 2013; Schmid et al., 2004; Beltrando et al., 2014; Casteras, 1933;

TABLE 3. DENSITY IN THE MANTLE AS A FUNCTION OF TEMPERATURE

T (°C)	700.000			800.000			900.000			1000.000			1100.000			1200.000			1300.000	
Mantle type	harz	fertile lherz		harz	fertile lherz		harz	fertile lherz		harz	fertile lherz		harz	fertile lherz		harz	fertile lherz		harzburgite	fertile lherz
Pressure (GPa)		plg			plg			plg			plg			plg			plg			
0.500	3.260	3.209		3.248	3.197	spl	3.235	3.185	spl	3.222	3.173	spl	3.209	3.161	spl	3.195	3.148	spl		
1.000	3.275	3.224	spl	3.263	3.213	3.276	3.250	3.201	3.264	3.237	3.189	3.251	3.225	3.177	3.239	3.212	3.165	3.226		
1.500	3.295		3.303	3.283		3.291	3.271		3.279	3.258		3.267	3.246		3.254	3.233		3.242		spl
2.000	3.309		3.317	3.297		3.306	3.300	grt	3.310	3.273		3.282	3.261	grt	3.270	3.249	grt	3.258	3.236	3.246
2.500	3.323	grt	3.331	3.311	grt	3.320	3.299	3.332	3.308	3.288	3.320	3.297	3.276	3.309	3.285	3.264	3.297	3.273	3.250	3.262
3.000	3.336	3.368		3.325	3.357		3.313	3.346		3.300	3.334		3.288	3.323		3.277	3.312		3.265	grt 3.300

Note: T—temperature. Pressure and composition calculated with the algorithms of Hacker (2003). Abbreviations: harz—harzburgite; lherz—lherzolite; plg—plagioclase; spl—spinel; grt—garnet; opx—orthopyroxene; cpx—clinopyroxene. Depleted harz: 80% olivine, 20% opx, both with a Mg# of 90–91; fertile plg peridotite (plg fertile, lherz): 9% plagioclase (An_{80}), 60% olivine (Fo_{90}), 24% opx (Mg# 90), 7% cpx (Mg# 90); fertile spl peridotite (spl fertile lherz): 2% spl (spl, hercynite, magnetite), 58% olivine (Fo_{90}), 26% opx (Mg# 90), 14% cpx (Mg# 90); fertile grt peridotite (grt fertile lherz): 9% garnet (pyrope 66%, almandine 24%, grossular 10%), 60% olivine (Fo_{89}), 24% opx (Mg# 89), 7% cpx (Mg# 90).

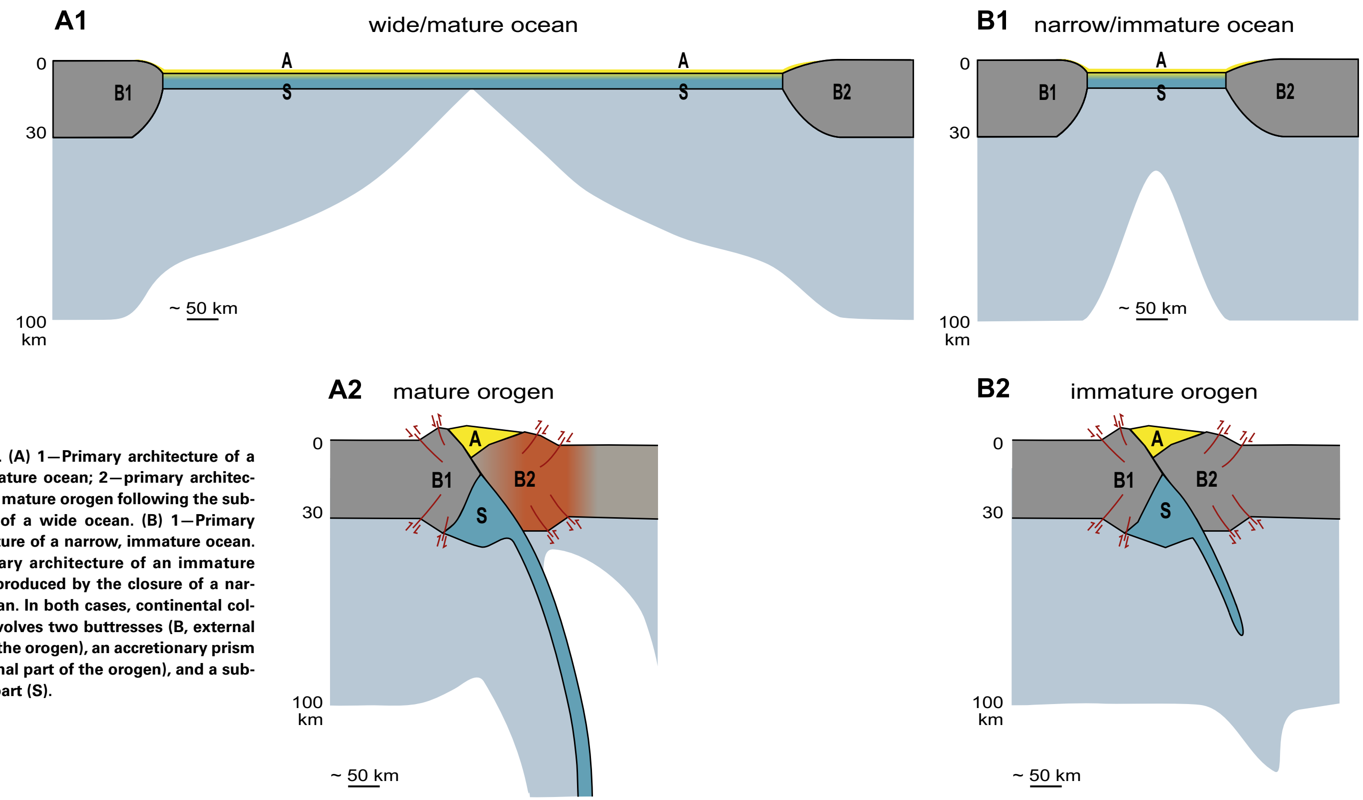


Figure 4. (A) 1—Primary architecture of a wide, mature ocean; 2—primary architecture of a mature orogen following the subduction of a wide ocean. (B) 1—Primary architecture of a narrow, immature ocean. 2—Primary architecture of an immature orogen produced by the closure of a narrow ocean. In both cases, 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).

Mattauer, 1968; Jammes et al., 2009; Muñoz, 1992). Recent studies (Mohn et al., 2014; Tugend et al., 2015) show consistently that the external parts of the orogen (the buttresses, B, in Fig. 4) correspond to the little deformed necking zones of the former continental margins. In contrast, the internal part corresponds to the accretionary prism (A in Fig. 4), which is composed of thinned continental basement remnants, ophiolites and/or exhumed mantle, and thick sequences of highly deformed sediments (see Beltrando et al., 2014, for a review). This accreted material is essentially derived from the sedimentary cover, but according to Andersen et al. (2012), some parts of the hyperextended and exhumation domains may also be integrated into the accretionary wedge, as evidenced by the remnants of exhumed mantle in the Caledonian orogen (see also Chew and Van Staal, 2014). This proposition is also verified in the Alps and in the Pyrenees, where most ophiolites are remnants from inherited or refertilized subcontinental mantle rather than from a steady-state spreading system (Lemoine et al., 1987; Müntener et al., 2010; Picazo et al., 2016).

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

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

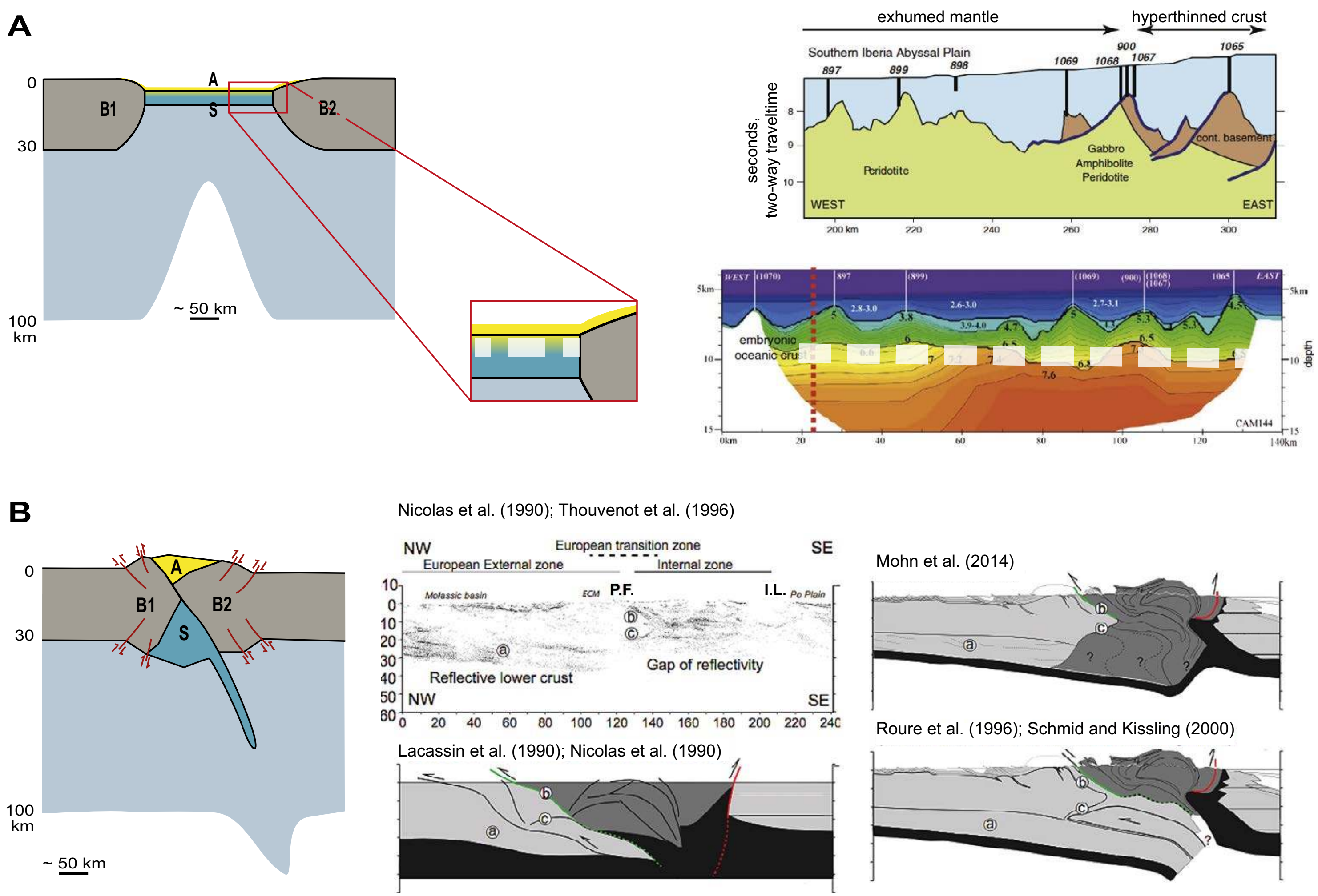


Figure 5. (A) The decoupling level between the subducted and accreted material during subduction corresponds possibly to the hydration front based on the geological interpretation of the reflection Lusigal 12 seismic section and the velocity model of the adjacent CAM 144 seismic section by Beltrando (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). Circled a–c mark the position of specific reflectors; they facilitate comparison of the different interpretations. A—accretionary prism; B—buttresses; S—subducted; ECM—external crystalline massifs; P.F.—Penninic Front; I.L.—Insubric Line.

In contrast, although high-grade metamorphism is recorded in the Alps (e.g., in the Tauern window and the Lepontine dome), heat transfer from the mantle is limited. The metamorphic belts of Alpine-type systems are created by stacking upper continental crustal slivers and internal heat production by radioactive decay (Burg and Gerya, 2005). The resulting metamorphism may have some impact on the rheology of the buried crust (Bellahsen et al., 2014), but is not significant enough to affect the overall architecture of the orogen (Fig. 4B). Therefore, in our study, we neglect the effects of low-grade metamorphism in immature orogens.

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

DISCUSSION

Since the advent of plate tectonics and the understanding of primary subduction processes, collisional orogens have been usually regarded as the result of the telescoping of continental margins after long-lasting subduction of a wide oceanic domain (Uyeda, 1981; Willett et al., 1993; Ernst, 2005; Handy et al., 2010). Such orogens are characterized by paired metamorphic belts, namely a low-temperature–high-pressure belt corresponding to the accretionary wedge and a high-temperature–low-pressure belt related to arc metamorphism and/or magmatism (Miyashiro, 1961; Dewey and Horsfield, 1970; Brown, 2009, and references therein). However, several collisional orogens such as the Alps and the Pyrenees lack voluminous magmatism contemporaneous to subduction; that is, they are devoid of remnants of arcs, forearcs, and backarcs and of high-temperature–low-pressure metamorphic assemblages in the upper plate. Yet both orogens result from the closure of relatively narrow (<400–600 km for the Alpine Tethys, <200 km for the Pyrenean rift system), hyperextended (Pyrenean rift system), or possibly embryonic (Alpine Tethys) oceans (Rosenbaum et al., 2002; Rosenbaum and Lister, 2005; Mohn et al., 2010; Lemoine et al., 1987). This may be the main cause for the lack of significant magmatic products. Because significant dehydration of the basaltic crust and serpentinized mantle starts from a depth of only 100–200 km (Peacock et al., 1994; Rüpke et al., 2004), a magmatic arc is unlikely to develop before the slab is subducted to this depth (e.g., Jarrard, 1986; England et al., 2004). Therefore, there must be a critical width for rift systems, below which their subduction is devoid of significant magmatic activity. In such cases, the subsequent orogenies may be essentially controlled by mechanical processes, in which the structural and lithologic architecture of the intervening margins may be the dominant factor in controlling the architecture of the orogen.

Narrow Oceans and Magma-Poor Subduction

Because slabs subduct with an average dip of 50°–60° in the upper mantle (Stevenson and Turner, 1977; Tovish et al., 1978; Billen, 2008), the slab must be at least 130 km long to reach a depth of 100 km. Adding to this twice the 55 km average length of each necking domain (see Primary Architecture discussion), which is usually not subducted (see Primary Architecture of Collisional Orogens discussion), magma generation seems very unlikely during closure of rift systems narrower than 240 km. Furthermore, subduction must last long enough for a sufficient amount of volatiles to induce hydrous partial melting (Peacock, 1991; Gaetani and Grove, 1998). This is consistent with the compilation by Jarrard (1986), which shows that the length of the subducted slab associated with the youngest arc (the Philippines arc, 6 m.y. old) is at least 170 km long. As a consequence, we expect rift basins narrower than 300 km to be devoid of significant magmatism expressed at the surface. In the following, we refer to these as narrow oceans, as opposed to wide oceans larger than ~1000 km.

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

Characteristics of Narrow versus Wide Oceans

As highlighted in Table 1, extensional systems narrower than 300 km are usually devoid of a mature, self-sustaining spreading system, thus of normal oceanic crust. As a result, their seafloor is composed of thinned continental crust, exhumed mantle, and/or embryonic oceanic crust. On the contrary, mature, steady-state oceanic systems are usually characterized by a homogeneous, Penrose-type oceanic crust. There is a wide range of oceanic crust types between these embryonic and Penrose-type end members. For example, in ultraslow-spreading systems, a thin (2–5 km thick) oceanic crust composed of both magmatic and amagmatic segments may be steadily emplaced for millions of years (Dick et al., 2003). However, like normal oceanic crust, oceanic crust formed at ultraslow-spreading ridges is most likely to be efficiently subducted during the closure of wide oceans (Stern, 2004), thus it will a priori not influence the subsequent orogeny.

As we showed herein (see Primary Architecture discussion), there is statistically no relationship between the size or maturity of a magma-poor hyperextended rift system and the architecture of its margins. Therefore, the main difference between narrow and wide oceans is the existence of a significant amount of oceanic crust and underlying depleted mantle (cf. Figs. 3A and 3B). In contrast, the mantle underlying narrow embryonic oceans is likely to retain its fertile composition resulting from the impregnation by asthenospheric melts during hyperextension (Müntener et al., 2010).

Subduction of Narrow versus Wide Oceans and Subsequent Orogenies

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

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

In contrast, protracted subduction associated with the closure of wide oceans is likely to become self-sustained, in particular due to the eclogitization of the subducting slab, which makes it denser than the encompassing asthenospheric mantle (Doin and Henry, 2001; Aoki and Takahashi, 2004). Long-lasting subduction develops usually vigorous convection in the mantle wedge that efficiently transports volatiles derived from the dehydration of the slab to great depth (Peacock et al., 1994). The resulting partial melting creates thickened sialic crust at the surface (magmatic arcs), induces high-temperature metamorphism in the encompassing upper plate, and depletes the underlying mantle in fusible elements (Uyeda, 1981).

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

Impact on Subsequent Collapse or Rifting Magmatic Budget

The difference in mantle composition resulting from the closure of narrow and/or embryonic oceans versus wide oceans may dictate the magmatic budget of subsequent extensional events such as postorogenic collapse or rifting. The depleted mantle beneath orogens related to mature subduction systems may not allow for voluminous magma production, in contrast to the fertile

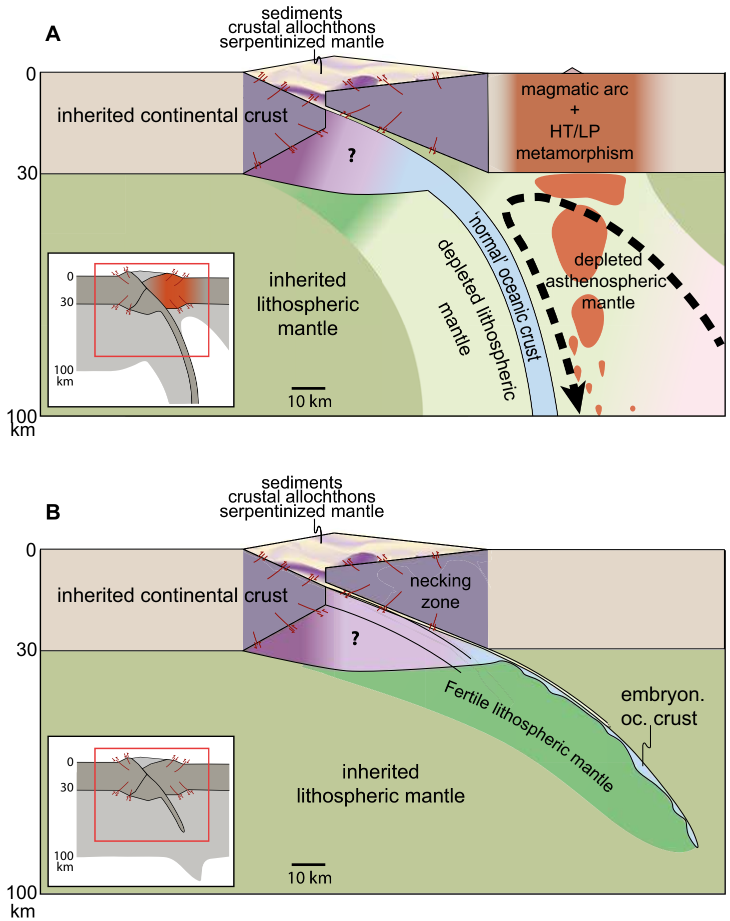


Figure 6. Architecture of orogens. (A) Following the closure of a wide and mature ocean. (B) Following the closure of a narrow and immature ocean. The insets highlight the scale of the zones represented. Abbreviations: embryon. oc.—embryonic oceanic crust; HT/LP—high temperature–low pressure.

mantle underlying orogens produced by the closure of narrow oceans. This hypothesis may account for both the amagmatic collapse of the Scandinavian Caledonides and the large amount of magmatism during the Variscan orogenic collapse.

The Variscides of western Europe result from the closure of several narrow oceans (McKerrow et al., 2000a; Franke, 2006), in addition to the suturing of the wide (Torsvik, 1998; Nance and Linnemann, 2008) Rheic Ocean (for reviews, see Matte, 2001; Kröner and Romer, 2013). Only the closure of the

Rheic Ocean presumably formed a significant magmatic arc (Franke, 2006). The orogenic collapse of the Variscan topography was followed by significant magmatic activity, which resulted in widespread, more or less acidic intrusions within the crust, and formed a thick mafic crustal underplating across most of the orogenic area (Bois et al., 1989; Rey, 1993; Costa and Rey, 1995; Schaltegger, 1997; Petri, 2014).

In contrast, the Scandinavian Caledonides between Norway and Greenland resulted essentially from the closure of the wide (>2000 km; van Staal et al., 2012) Iapetus Ocean (McKerrow et al., 2000b). At this latitude, the two-sided subduction of this ocean formed at least one major volcanic arc now exposed in Norway (Mac Niocaill et al., 1997; McKerrow et al., 2000b). Note that, further south, the British and Appalachian Caledonides involved many more microcontinents, narrow oceanic tracks, and volcanic arcs (see Roberts, 2003; Chew and Van Staal, 2014; van Staal et al., 2012; Fig. 1 therein), comparable to the Variscides of western Europe (see Franke, 2006; Fig. 2 therein). Thus we restrict our consideration of the Scandinavian Caledonides to the norther part between Norway and Greenland, where accretion of terranes was extremely limited. The Caledonian topography underwent a phase of orogenic collapse, which was essentially achieved through mechanical deformation without significant magmatic activity north of the Elbe lineament (McClay et al., 1986; Andersen, 1998; Meissner, 1999; Fossen et al., 2014).

An alternative hypothesis to account for the low magmatic budget of the Caledonian orogenic collapse relies on the depleted composition of the mantle underlying the two continents involved in the orogeny, Laurentia and Baltica. Both are composed of Archean cratonic cores (the North American and East European craton, respectively), which are characterized by a thick, cold, and depleted lithospheric mantle (Bernstein et al., 1998; Griffin et al., 2003; Beyer et al., 2004).

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

CONCLUSIONS

In this paper we show that each distal rift domain, as defined by Sutra et al. (2013), has a specific range of width. In contrast to the necking and exhumation domains, whose range of width is large, the width of the hyperextended domain is relatively consistent among rifted margins (~50 km). The widths of the necking domain and the hyperextended domain are independent of the total width of the margin (i.e., the distance from the coupling point to the lithospheric breakup point) and the maturity of the rift system (i.e., whether steady-state, self-sustaining seafloor spreading is achieved).

As narrow oceans are usually devoid of mature spreading systems (therefore called immature) in contrast to wide, mature oceans, the main difference between these end members is whether their margins are separated by a wide domain of normal oceanic crust. Furthermore, narrow and immature oceans are likely to be underlain by fertile mantle resulting from melt impregnation

TABLE 4. SUMMARY OF THE CHARACTERISTICS OF NARROW VERSUS WIDE EXTENSIONAL SYSTEMS, SUBDUCTION PROCESSES, AND OROGENS

	Narrow ocean	Wide ocean
Spatial extent	~350–400 km	>1000 km
Oceanic crust	Unsteady seafloor spreading ⇒ rough, heterogeneous in thickness and composition; no mantle depletion	Steady-state seafloor spreading ⇒ smooth, homogeneous in thickness and composition; depleted underlying mantle
Magmatic activity	None or very little ⇒ no mantle depletion	Moderate to large ⇒ 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 overthickened crust creation and associated high-temperature 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
Postorogenic collapse	Possibly magmatic	Purely extensional

during the phase of hyperextension, converse to wide and mature oceans, which have a typical mid-oceanic ridge-type depleted mantle.

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

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

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

Because of the lack of magmatic activity during the closure of narrow oceans, the mantle underlying the resulting orogens is likely more fertile than the mantle underlying orogens due to the closure of wide oceans. This difference in fertility may dictate the magmatic budget of a subsequent extensional event, such as a postorogenic collapse or an episode of rifting.

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