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1 **Towards a process-based approach to the study of rifted continental margins**

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19

20 **Abstract |**

21 Rifted margins form due to extension and break-up of the continental lithosphere. They host large
22 natural resources and will likely play a key role for the future transition to a carbon-neutral economy. Their
23 potential for becoming reservoirs for hydrothermally-induced mineral deposits, native hydrogen, and as sites
24 for CO₂ storage and generation of geothermal energy, rests on the tectonic, sedimentary and hydrothermal
25 processes that shape them. The interactions between these processes are complex and lead to a wide variety of
26 margins, yet specific well studied margins have led to a tendency to classify them as two end-members, magma-
27 poor or magma-rich, with a reductionist view of their evolution. In this review, we first summarize our current
28 knowledge of the structural elements and key processes shaping margins and their variability. We show how a
29 process-based understanding combined with high-quality observations, can be used to decipher margin
30 evolution, with a focus on specific examples. Finally, we discuss how a better understanding of the feedbacks
31 between deformation, sedimentation, magmatism and fluid-rock interactions, will be crucial to assess the
32 potential of margins to support the new carbon-neutral economy, and suggest new pathways to integrate
33 observations and modelling to de-risk exploration of these environments.

34

35 1. Introduction

36 Rifting of continents is a fundamental process of plate tectonics whereby the continental lithosphere
37 extends and thins, leading to breakup [G] and the formation of new oceans. Rifted margins are found
38 surrounding the once contiguous continents and contain some of the largest sediment accumulations worldwide
39 (Figure 1). Rifting has several wide-ranging consequences. For example, the opening of new oceanic gateways
40 can change ocean circulation, climate and the distribution of life^{1,2}. Breakup also generates CO₂ from
41 decompression melting and thermogenic CH₄ during magma emplacement and cooling in sedimentary sections,
42 which has consequences for understanding climate change during Earth history (e.g., Paleocene-Eocene
43 thermal maximum^{3,4}). During the last stages of extension at some margins, seawater reaches and hydrates the
44 mantle^{5,6}, generating H₂ and CH₄, which fuel chemosynthetic life in the deep seafloor, affecting the global
45 carbon cycle⁷⁻⁹.

46 For the last 50 years rifted margins have been the target of hydrocarbon exploration, as 60% of world
47 oil and gas reserves reside in these settings¹⁰. Today, as we shift towards a carbon neutral economy, the
48 exploration of rifted margins will be driven by their potential for hosting hydrothermally-induced mineral
49 deposits¹¹⁻¹³, native hydrogen reservoirs^{7,14}, CO₂ storage¹⁵, and as sites for generation of geothermal energy^{16,17}.
50 These applications will require a significantly improved understanding of the tectonic, physio-chemical,
51 hydrological and biological interactions occurring at margins. Bringing these processes, which occur at
52 different temporal and spatial scales, into a common framework will be a key endeavour in future margin
53 research.

54 The fundamental processes describing basin formation were developed in the 1970's and 80s¹⁸⁻²¹.
55 Margins form due to extension of the continental lithosphere, which thins and subsides as the underlying
56 asthenospheric mantle ascends, decompresses and melts. Extension culminates in the breakup of the former
57 continent and the formation of new oceanic crust. Normal oceanic crust can be classed as 6-7 km thick, with
58 an upper crust formed by extrusive basalts and dolerite dikes, overlying a gabbroic lower crust, and is known
59 as Penrose-like oceanic crust^{22,23}. Traditionally, the volume of magmatism during rifting was taken as a key
60 observable – with margins being classified as either magma-rich to magma-poor^{24,25}. This classification was
61 made by analogy to that of oceanic spreading, where the competition between magmatic processes, which
62 create new crust, and tectonic processes, which extend existing crust, shapes ridge architecture. The influence
63 of tectonics and magmatism is itself controlled by spreading velocity, mantle temperature and composition.
64 Continental margins, however, are further complicated by inheritance^{26,27}, so that the initial lithospheric thermal
65 and compositional structure modulates rifting processes, resulting in a vast variety of margin architectures²⁸.

66 In this Review we show how a process-based understanding of rifting, based on observations and
67 numerical modelling, illuminates the causes underlying the variety of extensional styles and magmatic budgets.

68 We first summarize our current knowledge of the structural elements (Section 2) and processes shaping margins
69 (Section 3). Subsequently we demonstrate how a process-based understanding, combined with high-quality
70 observations, can be used to decipher margin evolution across the global spectrum, with a focus on three
71 examples (Section 4). Our intention is to avoid evolutionary templates for margin architecture, but instead give
72 the tools to interpret their variability. Finally, we discuss how an understanding of the interactions between
73 deformation, sedimentation, magmatism, and fluid rock-interactions will be crucial to determine the potential
74 of margins for the new carbon-neutral economy (Section 5).

75 **2. Anatomy of Rifted Margins**

76 Rifted margins lie below sea-level such that drilling and seismic methods (wide-angle (WAS) [G] and
77 multichannel (MCS) [G]) are key in determining their structure (Box 1). The discovery that continental breakup
78 does not always immediately produce Penrose-like oceanic crust, as had been expected for extension under
79 normal mantle temperature²⁴, resulted in new paradigms to understand extension and breakup processes.
80 Exhumed serpentinised mantle with scarce magmatic products was first drilled at the continent-ocean transition
81 (COT [G]) west of Iberia, WIM, which together with its conjugate [G], the Newfoundland margin, NF, became
82 the archetypical magma-poor margins²⁹⁻³² (Figures 1, 2a, 3a).

83 Another major discovery was sequences of seaward dipping reflections (SDRs [G]) observed in MCS
84 images of several margins^{33,34}. These were shown to be interbedded sequences of sediment and lava flows³⁵⁻³⁷.
85 Margins with SDRs also have high p-wave seismic velocity lower crustal bodies, HVLC [G] (7.3-7.8 km/s)
86 interpreted as magmatic intrusions³⁸, and typically significantly thicker than normal oceanic crust after
87 breakup^{39,40}. These margins are collectively known as magma-rich (Figures 1, 2b, 3b) and are frequently
88 associated with onshore large igneous provinces (LIPs [G]) that form by the intrusion and extrusion of large
89 volumes of magmatism in very short geological periods, ≤ 2 Myr^{41,42} that are usually attributed to mantle plumes
90 [G] (Figure 1a).

91 Magma-rich margins are easily identified as SDR sequences stand out in MCS images (Figure 2).
92 Margins that do not exhibit SDRs in MCS images were often classified as magma-poor and interpreted to have
93 exhumed mantle at their COT (e.g. the South China sea, the central South Atlantic margins, the East Indian
94 margin⁴³⁻⁴⁶). This interpretation resulted in a binary classification of margins into either magma-poor or rich,
95 which was strongly influenced by the study of the magma-poor margin ophiolites in the Alps^{47,48}. The
96 identification of magma-poor margins from MCS data is however controversial as WAS is needed to clearly
97 identify exhumed mantle at the COT (Box 1, Figure 3), and this ambiguity is becoming more recognised^{49,50}.
98 Furthermore, several margins lacking SDRs appear to contain oceanic crust with magnetic anomaly lineations
99 [G] abutting continental crust, without intervening exhumed mantle, such as the Woodlark Basin^{51,52}, north-
100 eastern Brazilian margins⁵³⁻⁵⁵, Gulf of California⁵⁶, Ligurian⁵⁷, Red Sea⁵⁸, and the northeastern⁵⁹ and

101 northwestern South China Sea, SCS⁶⁰, where drilling confirmed an abrupt COT^{61,62}. These examples display a
102 wide variety of tectonic architectures and there is not a common structural template that fits them all.

103 Below we summarize the current knowledge of structural elements across the spectrum of margins.
104 We distinguish a proximal domain under the continental shelf and neighboring necking zone [G]; the distal
105 domain, where the continental crust is less than 15 km thick, and the outer domain, where continental crust is
106 ultra-thin and there is breakup (Figure 1b). Within any margin type, the available observations document
107 variations in all aspects.

108 **2.1 Magma-poor margins.** Rifting at these margins is characterised by large brittle faults, practically
109 absent syn-rift magmatism and crustal break-up followed by exhumation of the mantle at the COT²⁵. As there
110 is much uncertainty in the basement composition of the COT, here we consider magma-poor margins only
111 those whose COT has been characterized as exhumed mantle using integrated MCS and WAS data (see Box 1,
112 Figure 1), while recognizing that some margins not classified here as such, may indeed be magma-poor.

113 **From proximal to distal sectors.** The continental platform is characterised by small offset faults
114 dipping land- and oceanward and small degree of crustal thinning (> 25 km thick crust) distributed over a wide
115 area. The necking zone, which records the thinning from the platform to the distal domain, is generally narrow
116 (Figure 2a). In the distal domain, large fault bound rotated blocks are connected at depth to sub-horizontal
117 reflectivity in the lower crust (Figure 2a, DR). As crust thins, rotated blocks become smaller and their bounding
118 faults listric [G] (Figure 2a, L).

119 Towards the outer margin, faults merge at depth into a sub-horizontal reflection often interpreted as a
120 detachment fault^{48,63,64} (Figure 2a, D). WAS data suggest that the detachment is directly underlain by
121 serpentinitised mantle^{6,65} (Figure 3a), although a thin lower-crust layer underlying the detachment and overlying
122 serpentinitised mantle might be possible⁶⁶. The lateral extent of these detachments varies from 30 -15 km (e.g.,
123 S and H reflectors, in the West Iberia, Armorican margins and Goban Spur^{48,64,67-69}). In some places, however,
124 the continental crust thins to breakup without the involvement of such structures (Tagus Abyssal Plain in West
125 Iberia^{70,71}).

126 **Asymmetry.** Magma-poor conjugate margins are often asymmetric (Figure 2a). This is the case of the
127 Labrador – West Greenland margins⁷², and of the northern sector of the WIM-NF margins, where the WIM is
128 wider and has more rotated blocks, as described above, and the NF displays fewer fault blocks, no obvious
129 detachment, and abrupter crustal thinning^{73,74} (Figure 2a). To the south, central sectors of the WIM-NF margins
130 (Iberia Abyssal plain and central Newfoundland^{75,76}) shows the opposite trend. However, modern coincident
131 MCS and WAS data on conjugate margin pairs are rare, thus it is not clear whether all magma-poor margins
132 are asymmetric. For example, the Porcupine failed rift has flanks that appear symmetric although one fault

133 vergence dominates at ultra-thin crust⁷⁷, indicating that asymmetry may emerge during terminal
134 extension^{74,78,79}.

135 **The Continent-Ocean Transition.** The COT of magma-poor margins contains a series of margin-
136 parallel ridges^{76,80}. Where drilled, in the WIM-NF, these ridges consist of serpentinised peridotite with a small
137 proportion of magmatic products³² (Figure 2a). The P-wave velocity of exhumed mantle reduces with degree
138 of serpentinisation, from ~ 8 to 4.5 km/s⁸¹ (Figure 3a, Box 1) and is anisotropic⁸², complicating interpretation
139 of WAS profiles. This seismic velocity range overlaps with that of different types of metamorphic, intrusive
140 magmatic or oceanic rocks. However, a shallow steep velocity gradient and lack of abrupt crust-to-mantle
141 velocity boundary (Moho reflection) characterise exhumed mantle domains⁸³ (Box 1).

142 The width of exhumed mantle and its seismic velocity structure is variable (Nova Scotia^{84,85}, Labrador
143 sea^{72,86}, Orphan basin⁸⁷, Tyrrhenian^{88,89}, Deep Galicia margin⁶⁵, Tagus Abyssal Plain⁷¹, Flemish cap⁹⁰, Goban
144 Spur⁹¹, Bay of Biscay⁶⁷), and may be asymmetric across conjugate margins (Figure 2a). For example, it is ~160
145 km wide in the central WIM⁸³, whereas in the central NF the COT is ≤ 30 km wide⁷⁵, or may even be thin lower
146 continental crust⁹².

147 The degree of magmatism at the COT may also be asymmetric (Figure 2a). Comparison of WAS data
148 in the WIM and NF suggest the NF may be more magmatic^{75,92} (Figure 3a). Unfortunately, it is unclear whether
149 some of the WAS velocity differences among the COTs reflect real variability in exhumation and
150 serpentinisation processes, or are related to differences in velocity/depth models related to OBS spacing and
151 modelling techniques (Box 1). Imaging small-scale structures, such as few-km-thick oceanic crust or discrete
152 gabbroic bodies within serpentinised mantle, a narrower-than-typical OBS spacing of 10-15 km is required and
153 modern WAS data inversion techniques⁸⁵.

154 **Oceanic Crust.** A Penrose-like, 6-7 km thick oceanic crust with a 2-gradient velocity structure and
155 Moho reflection (Box 1) is not always observed oceanwards of exhumed mantle (Figure 2a), either because the
156 seismic lines do not extend far enough, because of complexities in the 3D structure, or because the structure of
157 oceanic crust in ultra-slow spreading crust differs from normal oceanic crust (Box 1). In West Iberia, WAS
158 data show slightly thicker than normal oceanic crust abutting exhumed mantle at the location of the prominent
159 J-magnetic-anomaly indicating an abrupt inception of seafloor spreading⁸³ (Figure 3a). However, oceanwards
160 of this location, oceanic crust is thinner than normal (~4 km thick^{71,83}). In the northern sector of the conjugate
161 Newfoundland margin, WAS data indicate that continental crust transitions directly to anomalously thin, 1-3
162 km, igneous oceanic crust, with layer 3 absent in its thinnest portions, overlying serpentinised mantle⁷³ (Figure
163 2a); while in the central sector of the Newfoundland margin, thin oceanic crust may appear after ≤ 30 km of
164 mantle exhumation⁷⁵. These observations open different possible scenarios for the establishment of seafloor
165 spreading (see Section 4).

166 **2.2 Magma-rich margins.** These margins are characterized by voluminous magmatism during rifting.
167 There is a complex interplay between tectonically induced crustal thinning and magmatism in the form of deep
168 intrusions, diking, and extrusions⁹³⁻⁹⁵. In studies of magma-rich margins in the North Atlantic, distinct
169 volcano-seismic facies and reflection characteristics were recognised that reflect different lava emplacement
170 environments^{96,97} (Figure 4). Key to this interpretation is to recognise if extrusions erupted into subaerial,
171 shallow water lacustrine/marine or deep marine environments. Seismic interpretation below SDRs is
172 challenging because the high acoustic impedance contrast between sediment and basalt. Magmatism that re-
173 thickens the thinned crust may also mask tectonically induced structures which complicates interpretation⁹⁸.

174 **From Proximal to Distal sectors.** The magmatic facies depend strongly on the inherited lithospheric
175 structure²⁷. In the North Atlantic, pronounced crustal thinning beginning in the late Paleozoic and continuing
176 throughout the Mesozoic occurred without significant magmatism, forming shallow marine basins. Initial
177 volcanism erupted into shallow water resulting in hyaloclastite flows, tuff deposits, lava deltas, and landward
178 flows that mark the proximal margin such as off Vøring and East Greenland^{96,99} (Figure 4). In contrast, the first
179 volcanism in the South Atlantic appears to have erupted into thicker continental crust undergoing necking
180 associated with early extension. This resulted in subaerial lava flows filling half grabens along landward
181 dipping faults¹⁰⁰⁻¹⁰² (Figures 2b and 4). In the literature these are termed type I SDR to distinguish them from
182 the main type II SDR sequence over the COT, which reach thickness of up to 15 km, significantly thicker than
183 those imaged in the North Atlantic^{102,103}. The lower crust of the proximal margin and COT is typically marked
184 by HVLC interpreted as evidence for intrusion of high MgO magmas at depth¹⁰⁴⁻¹⁰⁷. In the North Atlantic the
185 presence of injected sills within the lower crust has been directly imaged within the HVLC⁹⁸. In the South
186 Atlantic, the mid- and lower-crust shows high-amplitude discordant and anastomosing reflectivity, suggesting
187 either ductile deformation during extension and/or injected sills^{108,109} (Figure 2b, DR).

188 **The Continent-Ocean Transition.** The COT is marked by the main SDR sequences. In the South
189 Atlantic, the SDRs are spatially uniform and appear to form continuously (type II SDR, equivalent to 'inner
190 SDR' on the Vøring margin¹⁰²; Figure 2b, 4). The reflectivity pattern of type II SDR closely resembles the
191 Cenozoic lava sequences in eastern Iceland, which are built up into laterally extensive flow units with arcuate
192 shapes and consistent dips towards the neo-volcanic zone^{34,110}. These lavas are believed to flow some kilometres
193 away from the centres of eruption. Their seaward dip is attributed to the subsidence of the plate as it moves
194 away from the axis^{111,112}. The reflectivity pattern below most type II SDRs exhibits much smaller amplitudes
195 than below type I SDRs, consisting of a shallower seismically transparent package, underlain by discordant
196 high amplitude reflections in the lower crust¹⁰⁰. This suggests that intrusion below the type II SDRs takes place
197 in at most a very attenuated continental crust, so that most of the material is magmatic and exhibits less
198 impedance contrast with the surrounding crust⁹⁸ (Figure 2b). On many margins, some extended ultra-thin
199 continental crust cannot be excluded from the COT¹¹³. However, the pattern of reflectivity in combination with

200 the lack of faulted contacts between the type II SDRs and basement suggests a COT formed by subaerial oceanic
201 spreading that produces thicker igneous crust than normal. This interpretation is supported by the recognition
202 of linear magnetic anomalies very similar to submarine seafloor spreading anomalies¹¹⁴.

203 **Oceanic crust.** As the melt supply is reduced, volcanism evolves towards normal seafloor spreading.
204 Along Greenland and Vøring, this reduction was gradual and resulted in progressive subsidence of the proto-
205 spreading centre⁴⁰. Here a transition from subaerial to shallow marine volcanism is marked by a rough top
206 basement reflection or an outer volcanic high interpreted as hyaloclastite flows^{96,115} (Figure 4). Eventually, the
207 spreading system subsides to deeper water where the pressure prevents degassing and hyaloclastite formation
208 (Figure 4). Large sheet flows form an outer set of SDR (Figure 3b). Well-developed linear magnetic anomalies
209 can be interpreted in the crust that forms in these submarine environments. In contrast, in the South Atlantic,
210 the peak magmatism coincided with the subaerial SDR stage, with the first submarine oceanic crust having a
211 thickness of just 1 or 2 km more than normal¹¹⁶ (Figure 4).

212 **Asymmetry.** The distribution of magmatic products may vary both between the conjugate pairs and
213 along strike of the province. For example, in the South Atlantic WAS data shows 4 times larger HVLC zones
214 in the African compared to the South American side¹¹⁷. This conjugate asymmetry is also observed in the SDRs
215 which are longer and flatter in the African side (Figure 2a). Overall, southwards of the Parana-Etendeka LIP
216 the magma budget becomes smaller (and the distribution more asymmetric^{118,119}) whereas to the north there
217 is little magmatism^{120,121}. Similarly, in the North Atlantic south of Iceland the volume of magmatic rocks is
218 much larger along the Greenland than European sides (Figure 3b), whereas to the north of it is opposite¹²². All
219 these variabilities point to the interaction of the plume material with pre-existing lithospheric structure during
220 the generation of magma-rich margins¹²³⁻¹²⁵.

221 **2.3 An example of an intermediate margin.** Intermediate margins display no SDRs, indicative of
222 excessive magmatism nor exhumed mantle at the COT, indicative of deficient magmatism. However, they may
223 display geophysical evidence for robust magmatism as lower crustal magmatic bodies, as in the Angola
224 margin¹²⁶, and detachment structures, as in the SCS^{127,128}, which resemble structures observed in both magma-
225 rich and poor margins. Here we use the SCS as an example of, but not a template for, an intermediate margin.
226 The South China Sea rifted a lithosphere formed by a Mesozoic subduction system. Bounded by the Yangjiang-
227 Yitongansha Transfer Fault, the eastern margin broke up in the fore-arc area and the western margin broke up
228 roughly along the volcanic arc^{129,130}. Therefore, the crustal structure, deformation style and the amount of the
229 syn-rift magmatism varies greatly¹²⁸. The most distinguishable feature is that the eastern margin has up to 12
230 km thick HVLC, while the western margin usually bears sporadic thin layer (~2 km thick) of HVLC^{131,132}.
231 Here, we focus on the north-eastern sub-basin where two IODP legs took place.

232 **From Proximal to Distal sectors.** Thinned continental crust in this segment extends asymmetrically
233 across a large width, ~450 km in the north (Figure 2c). Thinning is progressive, but a clear necking zone is
234 observed^{60,133}. The proximal margin extends over 200 km and has a laterally homogeneous crustal thickness of
235 ~25-30 km. In this area, upper crust is extended by normal high-angle faults forming wide horsts and narrow
236 and deep grabens (HG, Figure 2c). Seawards, the crust thins abruptly to ~7 km and subsequently the large-scale
237 trend of crustal thinning is punctuated by local thickening below what are interpreted as large detachment
238 faults^{127,134} (D, Figure 2c). In WAS data¹³⁵, this thickening is accommodated mainly by the addition to the lower
239 crust of high-velocity layers, HVLC, with P-wave velocities of 7.0-7.3 km/s. Above the HVLC, high-amplitude
240 discordant and anastomosing reflectivity in the lower crust is indicative of magmatic intrusion at higher crustal
241 levels accompanied by intense ductile lower crustal flow^{127,134}. Unlike magma-rich margins, SDRs are not
242 observed and instead sills are interpreted to occur at syn-rift sedimentary levels¹³⁶.

243 **The Continent-Ocean Transition.** The COT at intermediate margins is narrow, typically ≤ 30 km⁵⁹⁻⁶¹
244 (Figure 2c). In the SCS, increasing magmatic intrusions are associated with basement uplift and sills are
245 emplaced in the overlying syn-rift sequences¹³⁴. Syn-intrusive forced folds in the sediment sequences suggest
246 dike intrusion occurred in late stage rifting and became younger oceanward¹³⁶. Seawards, the HVLC diminishes
247 in thickness and crustal breakup juxtaposes ultra-thin continental crust with normal Penrose-like oceanic
248 crust¹³⁵.

249 **Oceanic crust.** The earliest mid-ocean ridge basalt that IODP drilled is from a volcanic edifice located
250 in the outer margin, underlain by intra-crustal reflections which are interpreted as dykes cutting through ultra-
251 extended crust¹³⁷. Seismic images from several segments support that magma instead of faults trigger the initial
252 breakup in the SCS^{59,60}. At the IODP location, seismic images suggest that the COT may be composed of highly
253 thinned continental crust overlain by erupted basalt and underplated by gabbroic lower crust that grade into and
254 the early oceanic crust^{136,138}. The early oceanic crust is 5-6 km thick, and thickens slightly to 7-8 km oceanward
255 over 100 kms^{139,140}.

256 **3. Key parameters and processes**

257 Rifted margins form by the interaction between tectonic, magmatic, hydrothermal, and surface
258 processes modulated by the initial lithospheric structure and extension rates. These interactions are complex
259 and lead to a variety of margin architectures and evolutions. In this section, lithospheric-scale numerical models
260 are used to depict some of these interactions. These models summarize previously published results that show
261 that the initial and evolving lithospheric strength determine the spatio-temporal distribution of deformation^{28,141},
262 fault geometry¹⁴², stratigraphy^{143,144} and magmatism¹⁴⁵⁻¹⁴⁷ (Figure 5, Suppl. Info and Movie S1-S4). Interactions
263 not covered by these models are referenced from the published literature.

264 **3.1 Lithospheric rheology.** Observations of rifted margins such as the North Sea¹⁴⁸, Gulf of Corinth¹⁴⁹
265 and Suez¹⁵⁰ have shown that early extension is spatially distributed across numerous disconnected, short faults,
266 which with time grow laterally and merge so that deformation becomes progressively localized in fewer, larger
267 faults. The predominance in time and space of each of these phases, distributed vs. localized, determines the
268 width and asymmetry of rifted margins, and is related to the importance of faulting vs. ductile thinning, which
269 is a function of the evolving lithospheric rheology^{28,141,151}. In turn, this depends on lithospheric composition,
270 geotherm, extension velocity and potential rift-plume interactions.

271 For strong lithospheres, initial deformation is coupled from the upper crust to the mantle, through
272 narrow shear zones in the lower crust, leading to relatively rapid crustal thinning and mantle uplift (Suppl.
273 Figure 2a, Movie S1). This makes the first phases of deformation already quite spatially localized and
274 narrow^{28,141}, as in the Baikal and East African rifts, which develop close to strong cratons or ancient
275 platforms^{28,152}.

276 When the lower crust is weak, initial deformation is characterized by viscous thinning of the lower
277 crust over a broad area, which decouples deformation between upper crust and mantle and leads to
278 comparatively slow crustal thinning and mantle uplift (Suppl. Figure 2b). This results in the formation of a
279 wide system of horst and grabens, HG, underlain by little extended crust (Movie S3), as observed in the Basin
280 and Range and the proximal margin of the South China sea, where the pre-rift lithosphere is thinner and weaker
281 than in the previous examples^{28,127}.

282 As extension continues and the crust thins to $< \sim 15$ km conductive cooling becomes increasingly
283 important^{153,154}. This cooling, which is more predominant for slower extension velocities, in addition to strain
284 weakening [G], which reduces the strength of faults and shear zones where extension was previously localised,
285 leads to progressive localisation of deformation, and its coupling from the upper crust to the mantle, through
286 faults and narrow shear zones. For the same extension velocity, rifts that started in strong lithosphere will break-
287 up quickly, forming two narrow symmetric conjugates^{28,141} (Figure 5a, Movie S1). Instead, those starting in
288 weak lithospheres, will only localize after a prolonged phase of distributed extension, followed by rapid
289 localization and break-up. Their architecture is dominated by horsts and grabens, HG, formed during the initial
290 distributed phase (Figure 5c, Movie S3), resulting in wide, symmetric margins^{28,141}.

291 Symmetric margin development is disrupted when lateral migration of the deformation from one rift
292 side towards the conjugate becomes dominant¹⁵⁵. This occurs when the lower crust becomes strong enough to
293 couple deformation from upper crust to mantle, but is still ductile enough to prevent crustal breakup by
294 faulting^{145,155,156} (Suppl. Figure 2c, Movie S2). In this case, conjugate margins evolve asymmetrically, via a
295 system of oceanward younging, deep-penetrating fault/shear zones that work sequentially in time and nucleate
296 in the hangingwall [G] of the previous deep penetrating structure. This process generates a wide margin where
297 the sequential fault array [G], SF is active, and a narrow one which constitutes the hangingwall to the SF array⁷⁴

298 (Figure 5b, Movie S2). Lateral rift migration becomes more dominant with increasing extension velocity, as it
299 allows focused localisation at high temperatures which keep the lower crust and mantle ductile, thus preventing
300 break-up¹⁵⁵.

301 Conjugate margin asymmetry can also develop after a long phase of slow, distributed deformation,
302 which results in cooling and strengthening of the extending areas and localisation of rifting into the side of the
303 basin, where the crust is thicker and lithosphere weaker^{157,158} (Figure 5d, Movie S4).

304 **3.2 Fault distribution and geometry.** The evolution of fault systems and their relationship to ductile
305 deformation is important for assessing the amount of extension accommodated by brittle vs. ductile processes,
306 which impacts heat-flow. Fault system evolution strongly depends on lower crustal rheology^{143,159} (Movies S1-
307 S4).

308 During early rifting in weak lower crust, deformation is accommodated by many, simultaneously
309 active, small-offset faults, forming a system of horst and grabens, HG (Figure 5c-d, Movies S3-S4). With
310 increasing extension, new generations of faults cut the previous ones, a process known as polyphase faulting,
311 PF (Figure 5c and 5d), which may prevent fault recognition in MCS sections^{160,161}. For strong lower crust, early
312 faults exhibit larger offsets and faulting quickly migrate basinward, where no previous large faults were active,
313 and the PF is less intense (Movie S1, Figure 5a).

314 With increasing extension, deformation may migrate laterally, and fault geometry will depend on the
315 relative strength of the lower crust. When it is relatively strong, the wide margin exhibits a series of oceanward
316 rotated fault blocks, bounded by sequentially active faults, SF, that become progressively listric (SF, Figure 5b,
317 Movie S2) and eventually merge at depth to form sub-horizontal structures resembling detachment faults^{74,142}.
318 For relatively weaker crust, faults can attain longer offsets, as the flexural work needed to elastically
319 accommodate fault offset is small, and acquire concave-downward geometries (CCD in Figure 5d, Movie S4),
320 similar to those observed in oceanic core complexes¹⁶²⁻¹⁶⁵ and interpreted in the distal margins of the SCS
321 (Figure 2c).

322 **3.3 Sedimentation.** In addition to rheology, numerical models have shown that syn-tectonic
323 sedimentation influences deformation by increasing fault offset and thermally insulating the lithosphere, thus
324 keeping it warmer for longer. The first effect is most important in the proximal margin sectors, while the second
325 will decrease cooling in the distal areas where the crust is thin, increasing the ductility of the lower crust and
326 mantle in heavily sedimented margins^{144,166-168}.

327 In terms of stratigraphic architecture, early models of extension assumed that faulting occurred
328 synchronously with sediment deposition. Thus, syn-rift [G] sediments were classified as those exhibiting
329 wedge-shape geometries against their bounding faults, whereas post-rift [G] sediments draped over previous
330 topography, indicative for post-tectonic [G] deposition. Today, observational studies^{74,148,169,170} and numerical

331 models^{143,144} show that during rifting deformation migrates oceanward. This results in the syn-tectonic sediment
332 [G] younging oceanward, and the upper syn-rift sedimentary section drapes over previous topography along
333 the more proximal margin sections, i.e. has a post-tectonic geometry^{74,143,171} (Figure 5).

334 When the margin is wide, post-tectonic sequences can form large sag basins of syn-rift age, as observed
335 in the central South Atlantic (e.g. Angola margin^{143,172}, Santos basin¹⁷³, Figure 5d). As a result of rift migration,
336 unconformities [G] separating syn- and post-tectonic packages across a margin do not date continental breakup,
337 as earlier assumed, but the age at which deformation terminated in a given area and migrated basinwards. Only
338 unconformities across the outer margin should match breakup age¹⁴³. This is an important finding as such
339 unconformities had been typically used to date breakup age.

340 **3.4 Melting.** During thinning of the continental crust, lava is extruded at magma-rich margins forming
341 flows imaged as SDRs in MCS images⁹⁶. These extrusions are accompanied by intrusions of melt into the lower
342 crust⁹⁸ (Figure 2b). Intrusions will change how extension is accommodated. For example, in Afar extension
343 was initially accommodated by slip at the border faults, but presently extension is significantly accommodated
344 by melt intrusion at the rift centre¹⁷⁴, which facilitates rifting¹⁷⁵. Repeated dike and sill intrusions that can
345 generate the typical thickness of ~10 km of HVLC in periods of roughly 5 Myr, as observed magma-rich
346 margins e.g. Hatton Bank^{27,98}, would increase crustal temperatures by around 800°C¹⁷⁶, potentially weakening
347 the crust and promoting significant ductile deformation¹¹⁹.

348 On the other hand, not only will melting influence crustal strength, but the rheology of the crust will
349 also impact melt generation during extension. Thus, unlike at mid-ocean ridges melting is not only dependent
350 on extension velocity, mantle temperature, and composition, but also on lower crustal rheology^{145,177}. Spatially
351 distributed extension tends to delay the onset of decompression melting as mantle uplift is slow, while localised
352 extension tend to focus melting at the rift centre, thus resulting in a different spatial distribution of melt intrusion
353 along the margin (Figure 5, Movies S1-S4).

354 In addition, the composition of lavas and the surface area of flows in Iceland suggests that magmatism
355 is sensitive to surface processes, in this case the change in load due to deglaciation¹⁷⁸. Thus, there is a
356 connectivity between surface processes and deep mantle processes and the response times are rapid¹⁷⁹. When
357 the lower crust is weak, sediment loading can likewise impact the degree of decompression melting¹⁸⁰.

358 **3.5 Hydrothermal circulation.** The influence of hydrothermal circulation on the temperature field
359 during extension, and on shaping fault geometry and margin architecture, is not yet understood. Models
360 coupling deformation to hydrothermal circulation to our knowledge do not exist, mainly because of the complex
361 non-linear problem of coupling both fluid flow for highly viscous fluids (water and brine) with the visco-elasto-
362 plastic deformation of the crust and lithosphere. However, models have been used to look at the impact of
363 hydrothermal circulation on crustal accretion at mid-ocean ridges¹⁸¹ and it has a profound effect on crustal
364 temperatures and the depth of the melt lens. During margin formation hydrothermal circulation will likewise

365 favour mantle serpentinization as it will decrease the geothermal gradient, particularly at distal margins. In
366 addition, understanding the evolution of hydrothermal system will be important in predicting the location of
367 lithium rich brines and sediment-hosted metals¹¹.

368 **4. Towards a process-based understanding of rifting**

369 Having described how various geological attributes and processes influence margin evolution, we now
370 focus on how they shape the structural elements observed across the wide spectrum of margins, which is crucial
371 to assess their potential to host resources for the new green economy.

372 **4.1 Rifting at magma-poor margins.** Magma-poor margins form where extension velocities are on the
373 ultra-slow range¹⁸² (< 20 mm/yr full spreading) or the mantle is originally cold¹⁸³ or depleted¹⁸⁴. These factors
374 result in cold geothermal conditions which have several consequences.

375 **Absence of magmatism and mode of extension.** Under the above cited circumstances numerical
376 models predict very little or no melt production and mantle exhumation at the COT¹⁸⁵. Initial rifting conditions
377 suggest that extension in these margins probably started with a decoupled upper and lower crust^{74,186}. Due to
378 the ultra-slow extension, conductive cooling became important with ongoing extension, resulting in an
379 increased localisation and coupling of deformation throughout the crust⁵. This led to the formation of large
380 faults, which continued as ductile shear zones at depth, and pulled the ductile lower crust towards their
381 hangingwall as observed in WAS data⁷⁹. In numerical models⁶⁶ this pattern of extension leads to deformation
382 fabrics that resemble the lower crustal reflectivity patterns, DR, often observed in magma-poor margins^{43,79}
383 (Figure 2a) indicating the global predominance of coupling processes at these margins.

384 **Faulting, serpentinitisation and mantle exhumation.** The history of faulting, serpentinitisation and
385 mantle exhumation influences the potential of magma-poor margins to host natural H₂ reservoirs, as H₂ is a by-
386 product of serpentinitisation. Early, simple 1D models of the rheological evolution of these margins predicted
387 that cooling leads to embrittlement of the whole crust before detachments observed in the distal margin were
388 formed⁵ (D in Figure 2a). This was thought to be a key stage during margin evolution, (also known as "coupling
389 point"⁴³), as it allows faults to reach the mantle and bring enough water to serpentinitise it⁵. Serpentinitisation
390 induced weakening would have facilitated the formation of low-angle detachments in the brittle field⁶³.
391 However, new 2D dynamic simulation techniques allowing the reconstruction of seismic sections at fault-block
392 scale (Figure 6), suggest that these detachments may form in the ductile field⁶⁶. During extension of the ultra-
393 thin crust (< 10 km thick), the hangingwall of deep penetrating faults becomes hotter and ductile, so that the
394 overlying faults propagate at depth into ductile shear zones active at low angles⁶⁶ as interpreted from 3D MCS
395 data⁶³. Instead, the footwall [G] cools and becomes brittle allowing mantle serpentinitisation in this area⁶⁶.

396 **Crustal break-up and mantle exhumation.** Final breakup occurs through either a pair of conjugate
397 normal faults, forming two distal highs on each side of the new ocean floor, formed by either serpentinitised

398 mantle, as drilled in the West Iberia margin²⁹ or continental crust, as interpreted in Newfoundland⁷³.
399 Alternatively, interpretation from margin ophiolites in the Alps, suggests it occurs through concave-downward
400 detachments exposing serpentinised mantle and generating and asymmetric COT¹⁶². Numerical models indicate
401 that both situations are possible with normal faults forming in slightly colder conditions than concave-
402 downward detachments¹⁶³ (Figure 5c,d, Movies S2, and S4). Mantle exhumation may occur through a series of
403 faults that cut each other's footwall, known as flip-flop detachments, inferred to form the system of ridges
404 observed in the COT of magma-poor margins¹⁸⁷.

405 **Transition to seafloor spreading.** The abrupt lateral change from exhumed mantle to oceanic crust
406 west of Iberia indicates rapid initiation of seafloor spreading, perhaps supporting that the first oceanic crust
407 was formed by mid-ocean ridge propagation⁷¹. In contrast a slow transition from exhumed mantle to thin
408 oceanic crust as suggested for offshore central Newfoundland, might support the gradual establishment of a
409 spreading centre by progressive thinning of the lithosphere^{73,75,92} and a gradual change from tectonic to
410 magmatic processes¹⁸⁸. Such a progressive increase in magmatism would be promoted if rifting velocities
411 increase during extension¹⁸⁹. In addition, the asymmetric distribution of exhumed mantle suggests that the
412 location of deformation may have jumped, during the establishment of the steady-state oceanic spreading
413 centre^{71,73}. The ridge propagation hypothesis would indicate that magma-poor margins form where the mantle
414 is cold and only transitions to spreading upon ridge propagation and thus indicates strong lateral heterogeneity
415 of the sub-lithospheric mantle. Instead, the progressive establishment of a mid-ocean ridge would suggest that
416 mantle exposure is produced under ultra-slow velocities which increase with time as a result of extension
417 dynamics¹⁸⁹.

418 **4.2 Rifting at magma-rich margins.** Excess magmatism at rifted margins is principally a function of
419 the temperature of the mantle, but can be significantly modified by many secondary inherited attributes, such
420 as lithosphere structure, asthenosphere composition, and processes such as small-scale convection¹⁹⁰.

421 **Plume-lithosphere interaction.** Observations show that magmatism is not distributed symmetrically
422 around the LIP or assumed plume impact location, as classic models for magma-rich margins would
423 suggest^{24,191}. These variabilities point to the importance of the interaction of the plume material with pre-
424 existing lithospheric structure in the generation of magma-rich margins¹²²⁻¹²⁴. Since buoyant plume material
425 will flow towards the thinnest lithosphere¹⁹², a key factor for the formation of magma-rich margins is the timing
426 of the arrival of the thermal plume relative to the thinning of the continental lithosphere. In particular, if there
427 is no prior extension, the mantle will likely cool significantly prior to eventual breakup leading to limited
428 decompression melting of a depleted mantle as seen in the Seychelles-Laxmi Ridge margin²⁷. In contrast,
429 significant pre-existing thin spots that drain plume material favours forming magma-rich margins^{123,124}.

430 **Brittle vs. ductile deformation and emplacement of SDRs.** How the continental crust extends during
431 rifting at magma-rich margins is not clear, as seismic images are blurred beneath the thick SDR piles. In the

432 South Atlantic¹⁰², the first phases of extension appear to be accompanied by faulting that bounds the Type I
433 SDRs (Figure 2b). Instead, in the North Atlantic⁹⁶, SDRs are all akin to Type II and appear to be deposited
434 during increasing subsidence from the magma injection site, without the intervention of faulting (Figure 4).
435 Structural issues such as these will become critical if these volcanic sequences are required for future carbon
436 sequestration.

437 In the South Atlantic, close to the Parana-Etendeka flood basalts, the thickness of the SDR packages is
438 largest, conjugate margins are symmetric and the crust appears to have undergone large ductile
439 deformation^{102,119}, probably related to crustal heating resulting from magmatic intrusion. With distance from
440 the onshore flood basalts, the amount of offshore magmatism decreases¹¹⁶ and the margins become asymmetric,
441 suggesting relatively stronger (but still ductile) rheologies, as the smaller magma input provided less crustal
442 heating.

443 **Transition to seafloor spreading.** In the South Atlantic the COT appears to be formed by sub-aerial
444 spreading under large magmatic input, generating the long Type II SDRs sections. Seafloor spreading probably
445 started when the plume material beneath the extending axis waned as the ocean basin opened and the plume
446 had more space to laterally distribute. Decreasing magmatic input would generate increasing subsidence below
447 a water depth where extrusives would not laterally flow and form SDRs but form pillow lavas. The difference
448 in style of the SDRs formed in the South and North Atlantic show that elevation during SDR emplacement may
449 vary significantly depending on the pre-rift conditions (Figure 4).

450 **4.3 Rifting at intermediate margins.** Intermediate margins are characterised by an abrupt transition
451 from thinned continental to classic Penrose-like oceanic crust. These margins differ greatly in crustal
452 architecture and we focus here only in one example, the South China Sea, where full extension velocity of the
453 first oceanic crust is 50 mm/yr, much larger than at magma-poor margins, thus more magmatism is expected.
454 Extension in the South China Sea probably started in a wide rift mode, promoted by a hydrated and weak lower
455 crust and mantle resulting from its setting in a former volcanic arc^{59,129,130}. Wide rifting would have promoted
456 the formation of horst and grabens over ~200 km in the proximal margin (Figure 2c and HG in Figure 5d) until
457 localisation due to ongoing cooling took place in the distal margin. Here extension started to migrate towards
458 the basin centre, through a series of faults accompanied by ductile lower crustal deformation, which, in
459 conjunction with increasing heat released by the intruded magmatism may have promoted the formation of the
460 interpreted core complexes under ductile conditions in this area (D in Figure 2c, CCD in Figure 5d). As the
461 crust and lithosphere thinned, magma reached shallower levels and it intruded in the form of dykes in addition
462 to underplated gabbro. With increasing extension, the continental crust is dissected by several faults and dykes
463 giving way to the formation of classic Penrose-like oceanic crust.

464 **5. Importance for the energy transition**

465 Rifted margins are massive repositories of sedimentary, igneous and ultra-mafic rocks, are globally
466 distributed and lie adjacent to large, coastal populations. Thus, they are likely to hold a pivotal role in the future
467 transition to the new green economy. In the offshore environment, where there is the legacy of drilling by the
468 hydrocarbon industry, there is a potential for a greater acceptance of the level of future exploitation that will
469 be needed for the energy transition¹⁹³.

470 Whilst current Carbon Capture and Storage (CCS) projects have demonstrated the feasibility of storing
471 CO₂ underground, to meet IPCC model targets injection operations will need to be increased by several orders
472 of magnitude¹⁹⁴. This will require exploration for new reservoirs and careful consideration of long-term storage
473 security, especially for saline aquifers where the natural rates of mineralisation are sluggish. In this respect,
474 magma-rich margins may prove especially promising¹⁹⁵ as they contain abundant, reactive rocks such as basalts,
475 where injected CO₂ rapidly mineralizes to form stable carbonates, as has been proven for the onshore CarbFix
476 (Iceland) and Wallula (USA) projects¹⁹⁶.

477 Similarly, large amounts of H₂ are expected to be produced at the COTs of magma-poor margins due
478 to hydration of the exposed mantle rocks. Molecular hydrogen may become an important new fuel for internal
479 combustion engines. Promising reserves have been detected onshore where sections of these COTs are now
480 exposed in ophiolites, such as in the Mauleon Basin in the Pyrenees¹⁴. This setting portrays key ingredients for
481 long-term trapping of H₂, which are: a seal, e.g. impermeable clay or salt; a kitchen, e.g. iron-rich rocks; an
482 hydraulic system that favours water-rock interaction; and reservoir temperatures of ~100-200° C, where H₂ is
483 relatively inert¹⁴. In the offshore environment, present-day H₂ fluxes at the COT and whether it can be trapped
484 and form extractable reservoirs is still unknown.

485 Mineral deposits in the sedimentary sections of rifted margins form by hydrothermal water-rock
486 interaction. For example, base metals accumulate where there is a transition from thick to thin lithosphere¹¹.
487 This suggests that margins that form adjacent to cratonic lithosphere might be future reservoirs of base metals.
488 In rifts, such as the Rhine Graben, the startup Lithium de France will start geothermal exploration to extract
489 both heat and Lithium from reservoirs¹⁹⁷.

490 In all these examples, new understanding is needed of the fault-block scale interactions between fluid
491 flow at short time scales and the long-time scale lithosphere deformation, melt transport, and sedimentation
492 that create the conditions favourable for storage and extraction. 3D seismic imaging, which proved so critical
493 for hydrocarbon exploration, will need to be applied to new targets. Process-based models can help reduce risk
494 through fitting numerical models of coupled lithosphere and basin formation to observations to map the most
495 likely reservoir system at a large scale. In addition, the goal of understanding a given margin at the scales
496 needed for quantitative resource prediction can be pursued by fusing simulations of rift dynamics with
497 geological and geophysical data; commonly known as inversion and data assimilation (DA). A model with
498 adequate dynamics could, for example, assimilate MCS images. A first step in this direction is the heuristic

499 nudging approach used by⁶⁶ (Figure 6), which sequentially assimilates interpretations of MCS data, and shows
500 the potential of DA towards understanding tectonics at fault-block scale. In the future, advanced DA approaches
501 should be developed for geodynamic modelling of rifted margins.

502 **6. Summary and future Perspectives**

503 Rifted margins form by the interaction between tectonic, sedimentary, magmatic, hydrothermal and
504 surface processes modulated by the initial lithospheric structure and extension velocity. This leads to a variety
505 of architectures where magma-rich and -poor margins are only the end-members of a as yet poorly-defined
506 spectrum. In this review we documented that even within end-member types there is a rich variety in syn-rift
507 magmatism, margin width, conjugate asymmetry and COT nature. Drilling at the South China sea has shown
508 that conceptual templates of how margins evolve have skewed the interpretation of margins towards end-
509 member models. Similarly, the contrasting patterns of magmatism at the North and South Atlantic magma-rich
510 margins, shows how established conceptual models can be incomplete.

511 Margin studies require reference to all geophysical and geological data available, and an understanding
512 of how they complement each other. Geophysical surveys, in particular WAS surveys, which are not typically
513 carried out by industrial partners, need to be more common and, when existent, considered by the rest of the
514 community. For this, it is important that the different lateral and depth resolutions of MCS and WAS data and
515 the different information they convey is properly understood (Box 1). In addition, in sectors where the basement
516 is laterally heterogenous, such as the COTs, instrument spacing needs to be sufficiently small, and modelling
517 approaches exploit that resolution. One way to increase resolution is by using MCS data in between OBS
518 instrument location, and jointly invert both MCS and WAS⁷¹ and/or undertake downward continuation of both
519 MCS and OBS data¹⁹⁸, and when instrument spacing is sufficiently close use higher resolution full waveform
520 inversion⁸⁵.

521 Geodynamic models have been pivotal in documenting how different processes may explain margin
522 architecture. Initial lithospheric strength is key in determining rift width and asymmetry, and the spatio-
523 temporal distribution of magmatism, faulting and sedimentation. In numerical models, however, the link
524 between the original lithospheric structure and rift geometries is established in a general way. How the structure
525 and age of the initial fold-belt lithosphere affects rift architecture has only recently begun to be investigated¹⁹⁹.
526 3D modelling approaches are necessary to understand how rift development is linked to a broader tectonic plate
527 perspective²⁰⁰⁻²⁰².

528 Petroleum geoscience matured over many decades, meanwhile we are only at the beginning of
529 exploration for hydrogen and an expanding need for base and rare-earth metals and CO₂ storage. The source
530 rocks, fluid flow migration pathways, and reservoirs for these resources will require a renewed focus on the
531 interactions between sediments, deformation, fluid flow and magmatism. The fusion of data and models that

532 couple sedimentation, deformation, fluid flow and magmatism at sufficiently small scales will be pivotal to
533 make future accurate predictions for these new storage and energy needs.

534

535 **Figure captions**

536 **Figure 1. Global distribution of magma-rich, intermediate or uncertain and magma-poor margins.**

537 Oceanic crust age from²⁰³. Seaward dipping reflectors, SDRs, and onshore large-igneous provinces, with ages
538 <200 Ma, along with the ages of the onset of the main large-igneous province phase are shown²⁰⁴. Magmatic
539 margins are distinguished by their SDRs and comprise: NA: North Atlantic margins, associated to the North
540 Atlantic Igneous province, NAIP; the Eastern North American margin and its conjugate in Africa, associated
541 with the Central Atlantic Magmatic province, CAMP (with the African side classified as uncertain²⁰⁵); the
542 South Atlantic magmatic margins, SA, associated with the Paraná-Etendeka; the India-Seychelles margins
543 associated with the Deccan magmatic province, and the NW Australian margin. The Gulf of Mexico is
544 magmatic in the East but not in the West^{206,207}. Magma poor margins exhibit exhumed mantle at their continent-
545 ocean transition. Margins where exhumed mantle at the COT has been interpreted from wide-angle data are
546 shown. These include the West Iberia, WIM, and Newfoundland, NF, the Bay of Biscay, BB, the Goban Spur,
547 GB, the Porcupine Basin, PB, Rockall Trough, RT, Orphan Basin, OB, Labrador, L, the southern West
548 Greenland, WG, and the Thyrreanean, THY. The South Australian and its Antarctic conjugate have been
549 typically considered as magma-poor²⁰⁸⁻²¹⁰ but wide-angle data has not confirmed exhumed mantle yet, so a
550 question mark is included. Also shown are ODP drillings mentioned in the text, as black circles. Red lines are
551 wide-angle data shown in Figure 3.

552

553 **Figure 2. Conceptual models for a) magma-poor, b) magma-rich and c) intermediate margins based on**
554 **the West Iberia-Newfoundland margins, the South Atlantic magma-rich margins and the South China**
555 **sea.** As mentioned in the main text, within any of these margin types there is structural variability in terms of
556 amount of width and amount of magmatism within the COT, and margin width, along strike as well as between
557 conjugate margins. SDRs: seaward dipping reflectors, HVLC: high-velocity lower crustal body, DR: deep
558 reflectivity in the lower crust, L: listric faults, HG: Horst and grabens, PR: peridotite ridge. Note that the
559 horizontal scale differs in each panel.

560

561 **Figure 3. Wide-angle p-wave seismic velocity models across two contrasting conjugate continental**
562 **margins. a) the magma-poor Newfoundland-West Iberia margins^{83,92} and b) the magma-rich of SE Greenland-**
563 **Hatton Bank in the North Atlantic^{113,115} (location in Figure 1).** Models were generated by inverting wide-angle
564 seismic and seismic reflection data, the latter were used to constrain the sediment and basement properties.
565 Velocities at the continent-ocean transitions, COT, of magma-poor margins are interpreted as exhumed and
566 serpentinised mantle, with various degrees of magmatic contributions. Typical, 2-layered magmatic oceanic

567 crust is found in the seaward end of West Iberia but not in the Newfoundland section. Magma-rich margins
568 exhibit a high-velocity lower crustal body, HVLC, at their base. The thickness of the HVLC and its structural
569 position is asymmetric in this example, indicating asymmetry in melting and extension. Towards the oceanic
570 crust the HVLC thins. The distance over which this thinning occurs before typical oceanic crust is observed,
571 can be quite variable on different margins. In coincident MCS data it is shown that the HVLC is overlain by
572 seaward dipping reflections, SDRs. In the SE Greenland-Hatton Bank profiles, the oceanic crust inception is
573 defined by the C24r magnetic anomaly¹¹⁵.

574
575 **Figure 4. Formation of Seaward-Dipping Reflectors (SDRs) in the South and North Atlantic** (adapted
576 from^{96,102}. **a**) The geometries of the lava flows (colored blue, orange and green) depends on the tectonics
577 (formation of large-offset faults or not), magma supply rate and interaction with standing water (lakes or sea-
578 water – shown by the wavy line). The characteristic tilt results from either faulting or subsidence. In the North
579 Atlantic, as the lava sequence erupted, the elevation fell from above to below sea-level. This caused two distinct
580 lava flow accumulations separated by a hyaloclastic outer high (colored purple). In the South Atlantic, the
581 whole sequence appears to have been erupted while above sea-level with two packages (fault bounded colored
582 orange and non-fault bounded colored green) stacked on-top of each other. **b**) The difference in land elevation
583 may be explained by differences in crustal thickness (dotted blue line) at the time of plume arrival, with the
584 amount of pre-existing thinning larger in the North Atlantic¹⁷⁷. The melt production, shown by red line,
585 experienced a peak at different times during rifting, such that in the North Atlantic, the oceanic crust is thicker
586 than the SDRs, whereas in the South Atlantic it is reverse¹⁷⁷.

587
588 **Figure 5. Effect of decreasing initial lithospheric strength in tectonic, sedimentary architecture and**
589 **melting.** Extension velocity is 10 mm/yr, i.e. within the ultra-slow domain for oceanic spreading, and the
590 mantle temperature is 1300°C, thus models apply for magma-poor to intermediate margins (see Suppl. text for
591 model setup). The lithospheric strength decreases from **a**) to **d**) (see Figure S1 and Table S1). Models 1 to 3,
592 share the same crustal and lithospheric thickness (32 km and 120 km, respectively) and rheological properties
593 for upper (wet quartzite), lower crust (wet anorthite) and mantle (dry and wet olivine in the lithospheric and
594 asthenospheric mantle, respectively). The Moho temperature is 540°C in **a**), 570°C in **b**) and 770°C in **c**).
595 Model shown in **d**) is as model shown in **c**) but has a thicker crust of 35 km. Red shade is plastic/brittle strain
596 rate and blue is ductile strain rate. Grey shading shows accumulated plastic/brittle strain and represents areas
597 where faulting occurred. Sediments are color coded by age since the start of rifting. Dark orange areas are a
598 prediction of the thickness and location of underplated magmatic products. In the model the magma is only
599 underplated, but in nature it will intrude the crust and extrude it as dykes. With decreasing initial rheological
600 strength, margins become wider and the style of faulting, sediment architecture and the amount and distribution

601 of underplated/intruded magma changes. DH: distal highs, L: Listric faults, CCD: Concave-downward
602 detachments.

603

604 **Figure 6. Sequential nudging approach to simulate rift dynamics at fault-block scale (adapted from⁶⁶).**

605 Comparison of **a)** two conjugate MCS lines along the West Iberia-Newfoundland margins, with, **b)** results of
606 the modelling technique, Kinedyn^{66,211}. The interpretation of fault kinematics in the MCS line is incorporated
607 during the dynamic model run, so that the model is nudged towards the MCS data observations. Modelling at
608 fault-block scale yields the sediment and basement temperature field during the syn and post-rift margin
609 evolution. In this case, the model was used to understand how brittle and ductile deformation evolve with time,
610 and their influence on detachment, D, formation, and the spatial distribution of serpentinisation. In the future,
611 such models can be used to predict the pattern of hydrothermal circulation at margins and their consequences
612 for element exchange and deposition of minerals within the sedimentary sections. F1 to F7 are faults, B1 to B6
613 are fault-blocks. Red-blue-white stripes show the deformation within the crust, blue thick line is the Moho from
614 the model, DR are deep reflections. Shown are also dredges and ODP wells. Age of sediment is shown in colors.
615 Model is shown after 115 Myr of evolution.

616

617 **Figure 7. Conceptual margin cross section showing potential for economic resources and processes**

618 During rifting and perhaps thereafter higher than normal geothermal gradients will lead to hydrothermal
619 circulation, releasing lithium rich brines and base metals in the sedimentary sections (A and B). In addition,
620 sediments may host oil and gas reservoirs and may be used also as sites for CO₂ storage (C). Turbidites may
621 pose a hazard for marine infrastructure and at the same time host unconventional oil and gas reservoirs (D).
622 Finally, native H₂ produced during mantle hydration may be found in reservoirs which could be potentially
623 used as energy sources (E). 1) Proximal margins and necking zone, 2) distal margin, 3) outer margin, 4)
624 exhumed mantle, 5) oceanic crust. Note that magma-rich margins will look different and will host potential for
625 CO₂ storage in their SDRs sequences.

626

627

628 **Box 1 | Interpreting seismic data**

629 Seismic data typically falls into two categories – “multi-channel” seismic (MCS) where both the seismic source
630 and receiver are surface towed, and mainly records near-normal incidence reflected waves, and “wide-angle”
631 seismic (WAS) where the source is surface towed and the receiver placed on the seabed, and records wide-
632 angle refracted waves and reflected waves. WAS is currently the only technology to provide velocity-depth
633 information across the full depth extent needed for continental margin studies.

634 In recent years MCS data imaging has been revolutionized by using longer hydrophone streamers and
635 improved air gun technology, increasing our ability to suppress multiples and so image reflectivity at crustal

636 scale. However, MCS images are often shown in two-way-travel times which gives a highly distorted
637 impression due to highly variable low-velocity water and sediment layers across a margin. Pre-stack depth
638 migration (PSDM) was developed to address this issue, but it lacks sensitivity at depths larger than the streamer
639 length²¹². This means that coincident PSDM and wide-angle sections, may exhibit up to some kms of difference
640 in the depth of deep reflectors such as the Moho⁵⁵. In addition, wide-angle seismic models sample bulk
641 properties, thus complex reflectivity seen on co-incident MCS lines which do not present a bulk property
642 change, will not be observed in wide-angle data.

643 Although, absolute compressional wave velocities, V_p , of rocks, provided by wide-angle data do not
644 intrinsically differentiate between rock type as their values overlap in certain velocity ranges, their V_p gradients
645 allow to distinguish basement nature (Figure B1 and Figure B2). Over the years, mapping of these V_p gradients
646 has continuously improved with increasing quality of field data and modelling technologies. Closely spaced
647 OBS (<10 km) in conjunction with joint inversion of coincident MCS and wide-angle data provides to date,
648 the most accurate depiction of velocity-depth structure and Moho depth⁷¹.

649 Continental domains are readily distinguished from oceanic domains based on velocity-depth structure
650 (Figure B2). At magma-rich margins HVLC bodies are readily detected as they have absolute velocities higher
651 than the surrounding lower continental rocks ($7.0 < V_p < 7.8$). Analysis of their velocities have been shown to
652 be consistent with their volumes and is controlled by the depth and temperature of melting^{98,213,214}. Oceanic
653 crust generally displays a two-layer structure, although at slow-spreading the contrast between the upper steep
654 gradient layer and the lower gentle velocity gradient layer can be softened by the inclusion of a proportion of
655 serpentinitised mantle rocks^{215,216}. At ultra-slow oceanic spreading environments (< ~12 mm/yr spreading
656 velocity) the oceanic crust may disappear altogether so that the mantle is exhumed on the seafloor and a Moho
657 is missing^{217,218}. Care is therefore needed when interpreting lithology in such settings and across the COT of
658 magma-poor margins.

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661 **Glossary**

662 **Breakup:** The point (in time and space) where the thinning continents physically separate from each other.

663 **COT:** Continent-ocean transition. The area seawards of the thinned continental crust, which does not show
664 seismic velocity-depth nor tectonic structures which are typical of thinned continental or oceanic crust.

665 **Conjugate margins:** Two sides of an ocean basin that before rifting were joined.

666 **Footwall:** The block that doesn't experience subsidence on one side of an active fault.

667 **Hangingwall:** The block that experiences subsidence on one side of an active fault.

668 **HVLC:** High velocity lower crustal bodies (V_p of 7.3-7.8 km/s) observed in wide-angle seismic data along
669 some margins.

670 **LIP:** Large igneous provinces formed by the injection of large volumes of magmatism in very short
671 geological periods, 1-2 Myr or less. Traditionally interpreted as formed by the impact of a new mantle plume.

672 **Listric:** A fault that has high-angle in its shallowest segments (~60-30°) and has much lower angle at depth.

673 **MCS:** Multichannel seismic data, also known as reflection seismics. Distance between the source and
674 receiver is fixed. Provides high-resolution image but limited velocity information.

675 **Magnetic anomaly lineations:** Commonly interpreted to indicate seafloor spreading (oceanic crust). Formed
676 when mafic magma cools below the Curie Point (580⁰C) and takes the polarity of the Earth's field at that
677 time.

678 **Necking zone:** Where the crust thins from the continental platform (crustal thickness 28-25 km) towards the
679 distal domain (crustal thickness of ~20-15 km or less). Necking zones can be spatially abrupt or occur over a
680 large distance.

681 **Post-rift:** Period of time after breakup.

682 **Post-tectonic sediment:** sediment deposited after the activity of the underlying basement faults.

683 **SDRs:** Seaward-dipping reflectors. Formed of tilted, stacked, lava and sediment interbeds.

684 **Sequential faulting:** A system of faults that young oceanward and cut through the hangingwall of the
685 previous ones.

686 **Strain weakening:** A reduction in the strength of the lithosphere due to mechanical damage. Strain
687 weakening may result from the presence of fluids at faults, mineralization, reduction in grain size and the
688 formation of crystallographic preferred orientations.

689 **Syn-rift:** Period of time before breakup.

690 **Syn-tectonic sediment:** sediment deposited during the activity of the underlying basement faults.

691 **Unconformity:** A boundary between sedimentary rocks caused by a period of erosion or a pause in sediment
692 accumulation.

693 **Underplating** – igneous material added to the base of the crust.

694 **WAS:** Wide-angle seismic data, also known as refraction seismics. Distance between the source and receiver
695 varies. Provides low-resolution image but important velocity information.

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