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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 CO<sub>2</sub> 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<sup>1,2</sup>. Breakup also generates CO<sub>2</sub> from  
41 decompression melting and thermogenic CH<sub>4</sub> 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<sup>3,4</sup>). During the last stages of extension at some margins, seawater reaches and hydrates the  
44 mantle<sup>5,6</sup>, generating H<sub>2</sub> and CH<sub>4</sub>, which fuel chemosynthetic life in the deep seafloor, affecting the global  
45 carbon cycle<sup>7-9</sup>.

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 settings<sup>10</sup>. 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 deposits<sup>11-13</sup>, native hydrogen reservoirs<sup>7,14</sup>, CO<sub>2</sub> storage<sup>15</sup>, and as sites for generation of geothermal energy<sup>16,17</sup>.  
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 80s<sup>18-21</sup>.  
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<sup>22,23</sup>. 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<sup>24,25</sup>. 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<sup>26,27</sup>, so that the initial lithospheric thermal  
65 and compositional structure modulates rifting processes, resulting in a vast variety of margin architectures<sup>28</sup>.

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 temperature<sup>24</sup>, 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 margins<sup>29-32</sup> (Figures 1, 2a, 3a).

83 Another major discovery was sequences of seaward dipping reflections (SDRs [G]) observed in MCS  
84 images of several margins<sup>33,34</sup>. These were shown to be interbedded sequences of sediment and lava flows<sup>35-37</sup>.  
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 intrusions<sup>38</sup>, and typically significantly thicker than normal oceanic crust after  
87 breakup<sup>39,40</sup>. 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,  $\leq 2$  Myr<sup>41,42</sup> 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 margin<sup>43-46</sup>). 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<sup>47,48</sup>. 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<sup>49,50</sup>.  
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<sup>51,52</sup>, north-  
100 eastern Brazilian margins<sup>53-55</sup>, Gulf of California<sup>56</sup>, Ligurian<sup>57</sup>, Red Sea<sup>58</sup>, and the northeastern<sup>59</sup> and

101 northwestern South China Sea, SCS<sup>60</sup>, where drilling confirmed an abrupt COT<sup>61,62</sup>. 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 COT<sup>25</sup>. 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<sup>48,63,64</sup> (Figure 2a, D). WAS data suggest that the detachment is directly underlain by  
121 serpentinitised mantle<sup>6,65</sup> (Figure 3a), although a thin lower-crust layer underlying the detachment and overlying  
122 serpentinitised mantle might be possible<sup>66</sup>. 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<sup>48,64,67-69</sup>). In some places, however,  
124 the continental crust thins to breakup without the involvement of such structures (Tagus Abyssal Plain in West  
125 Iberia<sup>70,71</sup>).

126 **Asymmetry.** Magma-poor conjugate margins are often asymmetric (Figure 2a). This is the case of the  
127 Labrador – West Greenland margins<sup>72</sup>, 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<sup>73,74</sup> (Figure 2a). To the south, central sectors of the WIM-NF margins  
130 (Iberia Abyssal plain and central Newfoundland<sup>75,76</sup>) 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 crust<sup>77</sup>, indicating that asymmetry may emerge during terminal  
134 extension<sup>74,78,79</sup>.

135 **The Continent-Ocean Transition.** The COT of magma-poor margins contains a series of margin-  
136 parallel ridges<sup>76,80</sup>. Where drilled, in the WIM-NF, these ridges consist of serpentinised peridotite with a small  
137 proportion of magmatic products<sup>32</sup> (Figure 2a). The P-wave velocity of exhumed mantle reduces with degree  
138 of serpentinisation, from ~ 8 to 4.5 km/s<sup>81</sup> (Figure 3a, Box 1) and is anisotropic<sup>82</sup>, 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 domains<sup>83</sup> (Box 1).

142 The width of exhumed mantle and its seismic velocity structure is variable (Nova Scotia<sup>84,85</sup>, Labrador  
143 sea<sup>72,86</sup>, Orphan basin<sup>87</sup>, Tyrrhenian<sup>88,89</sup>, Deep Galicia margin<sup>65</sup>, Tagus Abyssal Plain<sup>71</sup>, Flemish cap<sup>90</sup>, Goban  
144 Spur<sup>91</sup>, Bay of Biscay<sup>67</sup>), and may be asymmetric across conjugate margins (Figure 2a). For example, it is ~160  
145 km wide in the central WIM<sup>83</sup>, whereas in the central NF the COT is  $\leq 30$  km wide<sup>75</sup>, or may even be thin lower  
146 continental crust<sup>92</sup>.

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<sup>75,92</sup> (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 techniques<sup>85</sup>.

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 spreading<sup>83</sup> (Figure 3a). However, oceanwards  
160 of this location, oceanic crust is thinner than normal (~4 km thick<sup>71,83</sup>). 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 mantle<sup>73</sup> (Figure  
163 2a); while in the central sector of the Newfoundland margin, thin oceanic crust may appear after  $\leq 30$  km of  
164 mantle exhumation<sup>75</sup>. 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 extrusions<sup>93-95</sup>. 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<sup>96,97</sup> (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 interpretation<sup>98</sup>.

174           **From Proximal to Distal sectors.** The magmatic facies depend strongly on the inherited lithospheric  
175 structure<sup>27</sup>. 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<sup>96,99</sup> (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 faults<sup>100-102</sup> (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<sup>102,103</sup>. 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 depth<sup>104-107</sup>. In the North Atlantic the  
185 presence of injected sills within the lower crust has been directly imaged within the HVLC<sup>98</sup>. 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<sup>108,109</sup> (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 margin<sup>102</sup>; 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<sup>34,110</sup>. 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<sup>111,112</sup>. 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 crust<sup>100</sup>. 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 crust<sup>98</sup> (Figure 2b). On many margins, some extended ultra-thin  
199 continental crust cannot be excluded from the COT<sup>113</sup>. 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 anomalies<sup>114</sup>.

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 centre<sup>40</sup>. 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<sup>96,115</sup> (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 normal<sup>116</sup> (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 side<sup>117</sup>. 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<sup>118,119</sup>) whereas to the north there  
217 is little magmatism<sup>120,121</sup>. 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 opposite<sup>122</sup>. All  
219 these variabilities point to the interaction of the plume material with pre-existing lithospheric structure during  
220 the generation of magma-rich margins<sup>123-125</sup>.

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 margin<sup>126</sup>, and detachment structures, as in the SCS<sup>127,128</sup>, 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<sup>129,130</sup>. Therefore, the crustal structure, deformation style and the amount of the  
229 syn-rift magmatism varies greatly<sup>128</sup>. 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<sup>131,132</sup>.  
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<sup>60,133</sup>. 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<sup>127,134</sup> (D, Figure 2c). In WAS data<sup>135</sup>, 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<sup>127,134</sup>. Unlike magma-rich margins, SDRs are not  
242 observed and instead sills are interpreted to occur at syn-rift sedimentary levels<sup>136</sup>.

243 **The Continent-Ocean Transition.** The COT at intermediate margins is narrow, typically  $\leq 30$  km<sup>59-61</sup>  
244 (Figure 2c). In the SCS, increasing magmatic intrusions are associated with basement uplift and sills are  
245 emplaced in the overlying syn-rift sequences<sup>134</sup>. Syn-intrusive forced folds in the sediment sequences suggest  
246 dike intrusion occurred in late stage rifting and became younger oceanward<sup>136</sup>. Seawards, the HVLC diminishes  
247 in thickness and crustal breakup juxtaposes ultra-thin continental crust with normal Penrose-like oceanic  
248 crust<sup>135</sup>.

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 crust<sup>137</sup>. Seismic images from several segments support that magma instead of faults trigger the initial  
252 breakup in the SCS<sup>59,60</sup>. 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<sup>136,138</sup>. The early oceanic crust is 5-6 km thick, and thickens slightly to 7-8 km oceanward  
255 over 100 kms<sup>139,140</sup>.

### 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<sup>28,141</sup>,  
262 fault geometry<sup>142</sup>, stratigraphy<sup>143,144</sup> and magmatism<sup>145-147</sup> (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 Sea<sup>148</sup>, Gulf of Corinth<sup>149</sup>  
265 and Suez<sup>150</sup> 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<sup>28,141,151</sup>. 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<sup>28,141</sup>, as in the Baikal and East African rifts, which develop close to strong cratons or ancient  
275 platforms<sup>28,152</sup>.

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<sup>28,127</sup>.

282           As extension continues and the crust thins to  $< \sim 15$  km conductive cooling becomes increasingly  
283 important<sup>153,154</sup>. 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<sup>28,141</sup> (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<sup>28,141</sup>.

291           Symmetric margin development is disrupted when lateral migration of the deformation from one rift  
292 side towards the conjugate becomes dominant<sup>155</sup>. 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<sup>145,155,156</sup> (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 array<sup>74</sup>

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-up<sup>155</sup>.

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<sup>157,158</sup> (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<sup>143,159</sup> (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<sup>160,161</sup>. 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<sup>74,142</sup>.  
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 complexes<sup>162-165</sup> 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<sup>144,166-168</sup>.

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<sup>74,148,169,170</sup> and numerical

331 models<sup>143,144</sup> 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<sup>74,143,171</sup> (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<sup>143,172</sup>, Santos basin<sup>173</sup>, 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 age<sup>143</sup>. 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 images<sup>96</sup>. These extrusions are accompanied by intrusions of melt into the lower  
342 crust<sup>98</sup> (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 centre<sup>174</sup>, which facilitates rifting<sup>175</sup>. 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<sup>27,98</sup>, would increase crustal temperatures by around 800°C<sup>176</sup>, potentially weakening  
347 the crust and promoting significant ductile deformation<sup>119</sup>.

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<sup>145,177</sup>. 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 deglaciation<sup>178</sup>. Thus, there is a  
356 connectivity between surface processes and deep mantle processes and the response times are rapid<sup>179</sup>. When  
357 the lower crust is weak, sediment loading can likewise impact the degree of decompression melting<sup>180</sup>.

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 ridges<sup>181</sup> 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 metals<sup>11</sup>.

#### 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 range<sup>182</sup> (< 20 mm/yr full spreading) or the mantle is originally cold<sup>183</sup> or depleted<sup>184</sup>. 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 COT<sup>185</sup>. Initial rifting conditions  
377 suggest that extension in these margins probably started with a decoupled upper and lower crust<sup>74,186</sup>. 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 crust<sup>5</sup>. 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 data<sup>79</sup>. In numerical models<sup>66</sup> this pattern of extension leads to deformation  
382 fabrics that resemble the lower crustal reflectivity patterns, DR, often observed in magma-poor margins<sup>43,79</sup>  
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 H<sub>2</sub> reservoirs, as H<sub>2</sub> 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 formed<sup>5</sup> (D in Figure 2a). This was thought to be a key stage during margin evolution, (also known as "coupling  
389 point"<sup>43</sup>), as it allows faults to reach the mantle and bring enough water to serpentinitise it<sup>5</sup>. Serpentinitisation  
390 induced weakening would have facilitated the formation of low-angle detachments in the brittle field<sup>63</sup>.  
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 field<sup>66</sup>. 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 angles<sup>66</sup> as interpreted from 3D MCS  
395 data<sup>63</sup>. Instead, the footwall [G] cools and becomes brittle allowing mantle serpentinitisation in this area<sup>66</sup>.

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 margin<sup>29</sup> or continental crust, as interpreted in Newfoundland<sup>73</sup>.  
399 Alternatively, interpretation from margin ophiolites in the Alps, suggests it occurs through concave-downward  
400 detachments exposing serpentinised mantle and generating and asymmetric COT<sup>162</sup>. Numerical models indicate  
401 that both situations are possible with normal faults forming in slightly colder conditions than concave-  
402 downward detachments<sup>163</sup> (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 margins<sup>187</sup>.

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 propagation<sup>71</sup>. 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<sup>73,75,92</sup> and a gradual change from tectonic to  
410 magmatic processes<sup>188</sup>. Such a progressive increase in magmatism would be promoted if rifting velocities  
411 increase during extension<sup>189</sup>. 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<sup>71,73</sup>. 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 dynamics<sup>189</sup>.

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 convection<sup>190</sup>.

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<sup>24,191</sup>. 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 margins<sup>122-124</sup>. Since buoyant plume material  
425 will flow towards the thinnest lithosphere<sup>192</sup>, 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 margin<sup>27</sup>. In contrast,  
429 significant pre-existing thin spots that drain plume material favours forming magma-rich margins<sup>123,124</sup>.

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 Atlantic<sup>102</sup>, the first phases of extension appear to be accompanied by faulting that bounds the Type I  
433 SDRs (Figure 2b). Instead, in the North Atlantic<sup>96</sup>, 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<sup>102,119</sup>, probably related to crustal heating resulting from magmatic intrusion. With distance from  
440 the onshore flood basalts, the amount of offshore magmatism decreases<sup>116</sup> 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<sup>59,129,130</sup>. 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 transition<sup>193</sup>.

470 Whilst current Carbon Capture and Storage (CCS) projects have demonstrated the feasibility of storing  
471 CO<sub>2</sub> underground, to meet IPCC model targets injection operations will need to be increased by several orders  
472 of magnitude<sup>194</sup>. 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 promising<sup>195</sup> as they contain abundant, reactive rocks such as basalts,  
475 where injected CO<sub>2</sub> rapidly mineralizes to form stable carbonates, as has been proven for the onshore CarbFix  
476 (Iceland) and Wallula (USA) projects<sup>196</sup>.

477 Similarly, large amounts of H<sub>2</sub> 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 Pyrenees<sup>14</sup>. This setting portrays key ingredients for  
481 long-term trapping of H<sub>2</sub>, 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 H<sub>2</sub> is  
483 relatively inert<sup>14</sup>. In the offshore environment, present-day H<sub>2</sub> 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 lithosphere<sup>11</sup>.  
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 reservoirs<sup>197</sup>.

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 by<sup>66</sup> (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 WAS<sup>71</sup> and/or undertake downward continuation of both  
519 MCS and OBS data<sup>198</sup>, and when instrument spacing is sufficiently close use higher resolution full waveform  
520 inversion<sup>85</sup>.

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 investigated<sup>199</sup>.  
526 3D modelling approaches are necessary to understand how rift development is linked to a broader tectonic plate  
527 perspective<sup>200-202</sup>.

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 CO<sub>2</sub> 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 from<sup>203</sup>. 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 shown<sup>204</sup>. 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 uncertain<sup>205</sup>); 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<sup>206,207</sup>. 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-poor<sup>208-210</sup> 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<sup>83,92</sup> and b) the magma-rich of SE Greenland-**  
563 **Hatton Bank in the North Atlantic<sup>113,115</sup> (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 anomaly<sup>115</sup>.

574  
575 **Figure 4. Formation of Seaward-Dipping Reflectors (SDRs) in the South and North Atlantic** (adapted  
576 from<sup>96,102</sup>. **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 Atlantic<sup>177</sup>. 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 reverse<sup>177</sup>.

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.

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604 **Figure 6. Sequential nudging approach to simulate rift dynamics at fault-block scale (adapted from<sup>66</sup>).**

605 Comparison of **a)** two conjugate MCS lines along the West Iberia-Newfoundland margins, with, **b)** results of  
606 the modelling technique, Kinedyn<sup>66,211</sup>. 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 CO<sub>2</sub> 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 H<sub>2</sub> 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 CO<sub>2</sub> storage in their SDRs sequences.

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#### 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 length<sup>212</sup>. 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 Moho<sup>55</sup>. 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 depth<sup>71</sup>.

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<sup>98,213,214</sup>. 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<sup>215,216</sup>. 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<sup>217,218</sup>. Care is therefore needed when interpreting lithology in such settings and across the COT of  
658 magma-poor margins.

659  
660

#### 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 (580<sup>0</sup>C) 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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