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

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Abstract |

Rifted margins form due to extension and break-up of the continental lithosphere. They host large natural resources and will likely play a key role for the future transition to a carbon-neutral economy. Their potential for becoming reservoirs for hydrothermally-induced mineral deposits, native hydrogen, and as sites for CO_2 storage and generation of geothermal energy, rests on the tectonic, sedimentary and hydrothermal processes that shape them. The interactions between these processes are complex and lead to a wide variety of margins, yet specific well studied margins have led to a tendency to classify them as two end-members, magma-poor or magma-rich, with a reductionist view of their evolution. In this review, we first summarize our current knowledge of the structural elements and key processes shaping margins and their variability. We show how a process-based understanding combined with high-quality observations, can be used to decipher margin evolution, with a focus on specific examples. Finally, we discuss how a better understanding of the feedbacks between deformation, sedimentation, magmatism and fluid-rock interactions, will be crucial to assess the potential of margins to support the new carbon-neutral economy, and suggest new pathways to integrate observations and modelling to de-risk exploration of these environments.

1. Introduction

Rifting of continents is a fundamental process of plate tectonics whereby the continental lithosphere extends and thins, leading to breakup [G] and the formation of new oceans. Rifted margins are found surrounding the once contiguous continents and contain some of the largest sediment accumulations worldwide (Figure 1). Rifting has several wide-ranging consequences. For example, the opening of new oceanic gateways can change ocean circulation, climate and the distribution of life[1,2]. Breakup also generates CO_2 from decompression melting and thermogenic CH_4 during magma emplacement and cooling in sedimentary sections, which has consequences for understanding climate change during Earth history (e.g., Paleocene-Eocene thermal maximum[3,4]). During the last stages of extension at some margins, seawater reaches and hydrates the mantle[5,6], generating H_2 and CH_4, which fuel chemosynthetic life in the deep seafloor, affecting the global carbon cycle[7-9].

For the last 50 years rifted margins have been the target of hydrocarbon exploration, as 60% of world oil and gas reserves reside in these settings[10]. Today, as we shift towards a carbon neutral economy, the exploration of rifted margins will be driven by their potential for hosting hydrothermally-induced mineral deposits[11-13], native hydrogen reservoirs[7,14], CO_2 storage[15], and as sites for generation of geothermal energy[16,17]. These applications will require a significantly improved understanding of the tectonic, physio-chemical, hydrological and biological interactions occurring at margins. Bringing these processes, which occur at different temporal and spatial scales, into a common framework will be a key endeavour in future margin research.

The fundamental processes describing basin formation were developed in the 1970's and 80s[18-21]. Margins form due to extension of the continental lithosphere, which thins and subsides as the underlying asthenospheric mantle ascends, decompresses and melts. Extension culminates in the breakup of the former continent and the formation of new oceanic crust. Normal oceanic crust can be classed as 6-7 km thick, with an upper crust formed by extrusive basalts and dolerite dikes, overlying a gabbroic lower crust, and is known as Penrose-like oceanic crust[22,23]. Traditionally, the volume of magmatism during rifting was taken as a key observable – with margins being classified as either magma-rich to magma-poor[24,25]. This classification was made by analogy to that of oceanic spreading, where the competition between magmatic processes, which create new crust, and tectonic processes, which extend existing crust, shapes ridge architecture. The influence of tectonics and magmatism is itself controlled by spreading velocity, mantle temperature and composition. Continental margins, however, are further complicated by inheritance[26,27], so that the initial lithospheric thermal and compositional structure modulates rifting processes, resulting in a vast variety of margin architectures[28].

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

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

2. Anatomy of Rifted Margins

Rifted margins lie below sea-level such that drilling and seismic methods (wide-angle (WAS) [G] and multichannel (MCS) [G]) are key in determining their structure (Box1). The discovery that continental breakup does not always immediately produce Penrose-like oceanic crust, as had been expected for extension under normal mantle temperature[24], resulted in new paradigms to understand extension and breakup processes. Exhumed serpentinised mantle with scarce magmatic products was first drilled at the continent-ocean transition (COT [G]) west of Iberia, WIM, which together with its conjugate [G], the Newfoundland margin, NF, became the archetypical magma-poor margins[29-32] (Figures 1, 2a, 3a).

Another major discovery was sequences of seaward dipping reflections (SDRs [G]) observed in MCS images of several margins[33,34]. These were shown to be interbedded sequences of sediment and lava flows[35-37]. Margins with SDRs also have high p-wave seismic velocity lower crustal bodies, HVLC [G] (7.3-7.8 km/s) interpreted as magmatic intrusions[38], and typically significantly thicker than normal oceanic crust after breakup[39,40]. These margins are collectively known as magma-rich (Figures 1, 2b, 3b) and are frequently associated with onshore large igneous provinces (LIPs [G]) that form by the intrusion and extrusion of large volumes of magmatism in very short geological periods, ≤2 Myr[41,42] that are usually attributed to mantle plumes [G] (Figure 1a).

Magma-rich margins are easily identified as SDR sequences stand out in MCS images (Figure 2). Margins that do not exhibit SDRs in MCS images were often classified as magma-poor and interpreted to have exhumed mantle at their COT (e.g. the South China sea, the central South Atlantic margins, the East Indian margin[43-46]). This interpretation resulted in a binary classification of margins into either magma-poor or rich, which was strongly influenced by the study of the magma-poor margin ophiolites in the Alps[47,48]. The identification of magma-poor margins from MCS data is however controversial as WAS is needed to clearly identify exhumed mantle at the COT (Box 1, Figure 3), and this ambiguity is becoming more recognised[49,50]. Furthermore, several margins lacking SDRs appear to contain oceanic crust with magnetic anomaly lineations [G] abutting continental crust, without intervening exhumed mantle, such as the Woodlark Basin[51,52], north-eastern Brazilian margins[53-55], Gulf of California[56], Ligurian[57], Red Sea[58], amd the northeastern[59] and northwestern South China Sea, SCS[60], where drilling confirmed an abrupt COT[61,62]. These examples display a wide variety of tectonic architectures and there is not a common structural template that fits them all.

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

2.1 Magma-poor margins. Rifting at these margins is characterised by large brittle faults, practically absent syn-rift magmatism and crustal break-up followed by exhumation of the mantle at the COT[25]. As there is much uncertainty in the basement composition of the COT, here we consider magma-poor margins only those whose COT has been characterized as exhumed mantle using integrated MCS and WAS data (see Box 1, Figure 1), while recognizing that some margins not classified here as such, may indeed be magma-poor.

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

Towards the outer margin, faults merge at depth into a sub-horizontal reflection often interpreted as a detachment fault[48,63,64] (Figure 2a, D). WAS data suggest that the detachment is directly underlain by serpentinised mantle[6,65] (Figure 3a), although a thin lower-crust layer underlying the detachment and overlying serpentinised mantle might be possible[66]. The lateral extent of these detachments varies from 30 -15 km (e.g., S and H reflectors, in the West Iberia, Armorican margins and Goban Spur[48,64,67-69]). In some places, however, the continental crust thins to breakup without the involvement of such structures (Tagus Abyssal Plain in West Iberia[70,71]).

Asymmetry. Magma-poor conjugate margins are often asymmetric (Figure 2a). This is the case of the Labrador – West Greenland margins[72], and of the northern sector of the WIM-NF margins, where the WIM is wider and has more rotated blocks, as described above, and the NF displays fewer fault blocks, no obvious detachment, and abrupter crustal thinning[73,74] (Figure 2a). To the south, central sectors of the WIM-NF margins (Iberia Abyssal plain and central Newfoundland[75,76]) shows the opposite trend. However, modern coincident MCS and WAS data on conjugate margin pairs are rare, thus it is not clear whether all magma-poor margins are asymmetric. For example, the Porcupine failed rift has flanks that appear symmetric although one fault vergence dominates at ultra-thin crust[77], indicating that asymmetry may emerge during terminal extension[74,78,79].

The Continent-Ocean Transition. The COT of magma-poor margins contains a series of margin-parallel ridges[76,80]. Where drilled, in the WIM-NF, these ridges consist of serpentinised peridotite with a small proportion of magmatic products[32] (Figure 2a). The P-wave velocity of exhumed mantle reduces with degree of serpentinisation, from ~ 8 to 4.5 km/s[81] (Figure 3a, Box 1) and is anisotropic[82], complicating interpretation of WAS profiles. This seismic velocity range overlaps with that of different types of metamorphic, intrusive magmatic or oceanic rocks. However, a shallow steep velocity gradient and lack of abrupt crust-to-mantle velocity boundary (Moho reflection) characterise exhumed mantle domains[83] (Box 1).

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

The degree of magmatism at the COT may also be asymmetric (Figure 2a). Comparison of WAS data in the WIM and NF suggest the NF may be more magmatic[75,92] (Figure 3a). Unfortunately, it is unclear whether some of the WAS velocity differences among the COTs reflect real variability in exhumation and serpentinisation processes, or are related to differences in velocity/depth models related to OBS spacing and modelling techniques (Box 1). Imaging small-scale structures, such as few-km-thick oceanic crust or discrete gabbroic bodies within serpentinised mantle, a narrower-than-typical OBS spacing of 10-15 km is required and modern WAS data inversion techniques[85].

Oceanic Crust. A Penrose-like, 6-7 km thick oceanic crust with a 2-gradient velocity structure and Moho reflection (Box 1) is not always observed oceanwards of exhumed mantle (Figure 2a), either because the seismic lines do not extend far enough, because of complexities in the 3D structure, or because the structure of oceanic crust in ultra-slow spreading crust differs from normal oceanic crust (Box 1). In West Iberia, WAS data show slightly thicker than normal oceanic crust abutting exhumed mantle at the location of the prominent J-magnetic-anomaly indicating an abrupt inception of seafloor spreading[83] (Figure 3a). However, oceanwards of this location, oceanic crust is thinner than normal (~4 km thick[71,83]). In the northern sector of the conjugate Newfoundland margin, WAS data indicate that continental crust transitions directly to anomalously thin, 1-3 km, igneous oceanic crust, with layer 3 absent in its thinnest portions, overlying serpentinised mantle[73] (Figure 2a); while in the central sector of the Newfoundland margin, thin oceanic crust may appear after ≤30 km of mantle exhumation[75]. These observations open different possible scenarios for the establishment of seafloor spreading (see Section 4).

2.2 Magma-rich margins. These margins are characterized by voluminous magmatism during rifting. There is a complex interplay between tectonically induced crustal thinning and magmatism in the form of deep intrusions, diking, and extrusions[93-95]. In studies of magma-rich margins in the North Atlantic, distinct volcano-seismic facies and reflection characteristics were recognised that reflect different lava emplacement environments[96,97] (Figure 4). Key to this interpretation is to recognise if extrusions erupted into subaerial, shallow water lacustrine/marine or deep marine environments. Seismic interpretation below SDRs is challenging because the high acoustic impedance contrast between sediment and basalt. Magmatism that re-thickens the thinned crust may also mask tectonically induced structures which complicates interpretation[98].

From Proximal to Distal sectors. The magmatic facies depend strongly on the inherited lithospheric structure[27]. In the North Atlantic, pronounced crustal thinning beginning in the late Paleozoic and continuing throughout the Mesozoic occurred without significant magmatism, forming shallow marine basins. Initial volcanism erupted into shallow water resulting in hyaloclastite flows, tuff deposits, lava deltas, and landward flows that mark the proximal margin such as off Vøring and East Greenland[96,99] (Figure 4). In contrast, the first volcanism in the South Atlantic appears to have erupted into thicker continental crust undergoing necking associated with early extension. This resulted in subaerial lava flows filling half grabens along landward dipping faults[100-102] (Figures 2b and 4). In the literature these are termed type I SDR to distinguish them from the main type II SDR sequence over the COT, which reach thickness of up to 15 km, significantly thicker than those imaged in the North Atlantic[102,103]. The lower crust of the proximal margin and COT is typically marked by HVLC interpreted as evidence for intrusion of high MgO magmas at depth[104-107]. In the North Atlantic the presence of injected sills within the lower crust has been directly imaged within the HVLC[98]. In the South Atlantic, the mid- and lower-crust shows high-amplitude discordant and anastomosing reflectivity, suggesting either ductile deformation during extension and/or injected sills[108,109] (Figure 2b, DR).

The Continent-Ocean Transition. The COT is marked by the main SDR sequences. In the South Atlantic, the SDRs are spatially uniform and appear to form continuously (type II SDR, equivalent to 'inner SDR' on the Vøring margin[102]; Figure 2b, 4). The reflectivity pattern of type II SDR closely resembles the Cenozoic lava sequences in eastern Iceland, which are built up into laterally extensive flow units with arcuate shapes and consistent dips towards the neo-volcanic zone[34,110]. These lavas are believed to flow some kilometres away from the centres of eruption. Their seaward dip is attributed to the subsidence of the plate as it moves away from the axis[111,112]. The reflectivity pattern below most type II SDRs exhibits much smaller amplitudes than below type I SDRs, consisting of a shallower seismically transparent package, underlain by discordant high amplitude reflections in the lower crust[100]. This suggests that intrusion below the type II SDRs takes place in at most a very attenuated continental crust, so that most of the material is magmatic and exhibits less impedance contrast with the surrounding crust[98] (Figure 2b). On many margins, some extended ultra-thin continental crust cannot be excluded from the COT[113]. However, the pattern of reflectivity in combination with the lack of faulted contacts between the type II SDRs and basement suggests a COT formed by subaerial oceanic spreading that produces thicker igneous crust than normal. This interpretation is supported by the recognition of linear magnetic anomalies very similar to submarine seafloor spreading anomalies[114].

Oceanic crust. As the melt supply is reduced, volcanism evolves towards normal seafloor spreading. Along Greenland and Vøring, this reduction was gradual and resulted in progressive subsidence of the proto-spreading centre[40]. Here a transition from subaerial to shallow marine volcanism is marked by a rough top basement reflection or an outer volcanic high interpreted as hyaloclastite flows[96,115] (Figure 4). Eventually, the spreading system subsides to deeper water where the pressure prevents degassing and hyaloclastite formation (Figure 4). Large sheet flows form an outer set of SDR (Figure 3b). Well-developed linear magnetic anomalies can be interpreted in the crust that forms in these submarine environments. In contrast, in the South Atlantic, the peak magmatism coincided with the subaerial SDR stage, with the first submarine oceanic crust having a thickness of just 1 or 2 km more than normal[116] (Figure 4).

Asymmetry. The distribution of magmatic products may vary both between the conjugate pairs and along strike of the province. For example, in the South Atlantic WAS data shows 4 times larger HVLC zones in the African compared to the South American side[117]. This conjugate asymmetry is also observed in the SDRs which are longer and flatter in the African side (Figure 2a). Overall, southwards of the Parana-Etendeka LIP the magma budget becomes smaller (and the distribution more asymmetric[118,119]) whereas to the north there is little magmatism[120,121]. Similarly, in the North Atlantic south of Iceland the volume of magmatic rocks is much larger along the Greenland than European sides (Figure 3b), whereas to the north of it is opposite[122]. All these variabilities point to the interaction of the plume material with pre-existing lithospheric structure during the generation of magma-rich margins[123-125].

2.3 An example of an intermediate margin. Intermediate margins display no SDRs, indicative of excessive magmatism nor exhumed mantle at the COT, indicative of deficient magmatism. However, they may display geophysical evidence for robust magmatism as lower crustal magmatic bodies, as in the Angola margin[126], and detachment structures, as in the SCS[127,128], which resemble structures observed in both magma-rich and poor margins. Here we use the SCS as an example of, but not a template for, an intermediate margin. The South China Sea rifted a lithosphere formed by a Mesozoic subduction system. Bounded by the Yangjiang-Yitongansha Transfer Fault, the eastern margin broke up in the fore-arc area and the western margin broke up roughly along the volcanic arc[129,130]. Therefore, the crustal structure, deformation style and the amount of the syn-rift magmatism varies greatly[128]. The most distinguishable feature is that the eastern margin has up to 12 km thick HVLC, while the western margin usually bears sporadic thin layer (~2 km thick) of HVLC[131,132]. Here, we focus on the north-eastern sub-basin where two IODP legs took place.

From Proximal to Distal sectors. Thinned continental crust in this segment extends asymmetrically across a large width, ~450 km in the north (Figure 2c). Thinning is progressive, but a clear necking zone is observed[60,133]. The proximal margin extends over 200 km and has a laterally homogeneous crustal thickness of ~25-30 km. In this area, upper crust is extended by normal high-angle faults forming wide horsts and narrow and deep grabens (HG**,** Figure 2c). Seawards, the crust thins abruptly to ~7 km and subsequently the large-scale trend of crustal thinning is punctuated by local thickening below what are interpreted as large detachment faults[127,134] (D, Figure 2c). In WAS data[135], this thickening is accommodated mainly by the addition to the lower crust of high-velocity layers, HVLC, with P-wave velocities of 7.0-7.3 km/s. Above the HVLC, high-amplitude discordant and anastomosing reflectivity in the lower crust is indicative of magmatic intrusion at higher crustal levels accompanied by intense ductile lower crustal flow[127,134]. Unlike magma-rich margins, SDRs are not observed and instead sills are interpreted to occur at syn-rift sedimentary levels[136].

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

Oceanic crust. The earliest mid-ocean ridge basalt that IODP drilled is from a volcanic edifice located in the outer margin, underlain by intra-crustal reflections which are interpreted as dykes cutting through ultra-extended crust[137]. Seismic images from several segments support that magma instead of faults trigger the initial breakup in the SCS[59,60]. At the IODP location, seismic images suggest that the COT may be composed of highly thinned continental crust overlain by erupted basalt and underplated by gabbroic lower crust that grade into and the early oceanic crust[136,138]. The early oceanic crust is 5-6 km thick, and thickens slightly to 7-8 km oceanward over 100 kms[139,140].

3. Key parameters and processes

Rifted margins form by the interaction between tectonic, magmatic, hydrothermal, and surface processes modulated by the initial lithospheric structure and extension rates. These interactions are complex and lead to a variety of margin architectures and evolutions. In this section, lithospheric-scale numerical models are used to depict some of these interactions. These models summarize previously published results that show that the initial and evolving lithospheric strength determine the spatio-temporal distribution of deformation[28,141], fault geometry[142], stratigraphy[143,144] and magmatism[145-147] (Figure 5, Suppl. Info and Movie S1-S4). Interactions not covered by these models are referenced from the published literature.

3.1 Lithospheric rheology. Observations of rifted margins such as the North Sea[148], Gulf of Corinth[149] and Suez[150] have shown that early extension is spatially distributed across numerous disconnected, short faults, which with time grow laterally and merge so that deformation becomes progressively localized in fewer, larger faults. The predominance in time and space of each of these phases, distributed vs. localized, determines the width and asymmetry of rifted margins, and is related to the importance of faulting vs. ductile thinning, which is a function of the evolving lithospheric rheology[28,141,151]. In turn, this depends on lithospheric composition, geotherm, extension velocity and potential rift-plume interactions.

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

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

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

Symmetric margin development is disrupted when lateral migration of the deformation from one rift side towards the conjugate becomes dominant[155]. This occurs when the lower crust becomes strong enough to couple deformation from upper crust to mantle, but is still ductile enough to prevent crustal breakup by faulting[145,155,156] (Suppl. Figure 2c, Movie S2). In this case, conjugate margins evolve asymmetrically, via a system of oceanward younging, deep-penetrating fault/shear zones that work sequentially in time and nucleate in the hangingwall [G] of the previous deep penetrating structure. This process generates a wide margin where the sequential fault array [G], SF is active, and a narrow one which constitutes the hangingwall to the SF array[74]

(Figure 5b, Movie S2). Lateral rift migration becomes more dominant with increasing extension velocity, as it allows focused localisation at high temperatures which keep the lower crust and mantle ductile, thus preventing break-up[155].

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

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

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

With increasing extension, deformation may migrate laterally, and fault geometry will depend on the relative strength of the lower crust. When it is relatively strong, the wide margin exhibits a series of oceanward rotated fault blocks, bounded by sequentially active faults, SF, that become progressively listric (SF, Figure 5b, Movie S2) and eventually merge at depth to form sub-horizontal structures resembling detachment faults[74,142]. For relatively weaker crust, faults can attain longer offsets, as the flexural work needed to elastically accommodate fault offset is small, and acquire concave-downward geometries (CCD in Figure 5d, Movie S4), similar to those observed in oceanic core complexes[162-165] and interpreted in the distal margins of the SCS (Figure 2c).

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

In terms of stratigraphic architecture, early models of extension assumed that faulting occurred synchronously with sediment deposition. Thus, syn-rift [G] sediments were classified as those exhibiting wedge-shape geometries against their bounding faults, whereas post-rift [G] sediments draped over previous topography, indicative for post-tectonic [G] deposition. Today, observational studies[74,148,169,170] and numerical models[143,144] show that during rifting deformation migrates oceanward. This results in the syn-tectonic sediment [G] younging oceanward, and the upper syn-rift sedimentary section drapes over previous topography along the more proximal margin sections, i.e. has a post-tectonic geometry[74,143,171] (Figure 5).

When the margin is wide, post-tectonic sequences can form large sag basins of syn-rift age, as observed in the central South Atlantic (e.g. Angola margin[143,172], Santos basin[173], Figure 5d). As a result of rift migration, unconformities [G] separating syn- and post-tectonic packages across a margin do not date continental breakup, as earlier assumed, but the age at which deformation terminated in a given area and migrated basinwards. Only unconformities across the outer margin should match breakup age[143]. This is an important finding as such unconformities had been typically used to date breakup age.

3.4 Melting. During thinning of the continental crust, lava is extruded at magma-rich margins forming flows imaged as SDRs in MCS images[96]. These extrusions are accompanied by intrusions of melt into the lower crust[98] (Figure 2b). Intrusions will change how extension is accommodated. For example, in Afar extension was initially accommodated by slip at the border faults, but presently extension is significantly accommodated by melt intrusion at the rift centre[174], which facilitates rifting[175]. Repeated dike and sill intrusions that can generate the typical thickness of ~10 km of HVLC in periods of roughly 5 Myr, as observed magma-rich margins e.g. Hatton Bank[27,98], would increase crustal temperatures by around 800°C[176], potentially weakening the crust and promoting significant ductile deformation[119].

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

In addition, the composition of lavas and the surface area of flows in Iceland suggests that magmatism is sensitive to surface processes, in this case the change in load due to deglaciation[178]. Thus, there is a connectivity between surface processes and deep mantle processes and the response times are rapid[179]. When the lower crust is weak, sediment loading can likewise impact the degree of decompression melting[180].

3.5 Hydrothermal circulation. The influence of hydrothermal circulation on the temperature field during extension, and on shaping fault geometry and margin architecture, is not yet understood. Models coupling deformation to hydrothermal circulation to our knowledge do not exist, mainly because of the complex non-linear problem of coupling both fluid flow for highly viscous fluids (water and brine) with the visco-elasto-plastic deformation of the crust and lithosphere. However, models have been used to look at the impact of hydrothermal circulation on crustal accretion at mid-ocean ridges[181] and it has a profound effect on crustal temperatures and the depth of the melt lens. During margin formation hydrothermal circulation will likewise favour mantle serpentinization as it will decrease the geothermal gradient, particularly at distal margins. In addition, understanding the evolution of hydrothermal system will be important in predicting the location of lithium rich brines and sediment-hosted metals[11] .

4. Towards a process-based understanding of rifting

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

4.1 Rifting at magma-poor margins. Magma-poor margins form where extension velocities are on the ultra-slow range[182] (< 20 mm/yr full spreading) or the mantle is originally cold[183] or depleted[184]. These factors result in cold geothermal conditions which have several consequences.

Absence of magmatism and mode of extension. Under the above cited circumstances numerical models predict very little or no melt production and mantle exhumation at the COT[185]. Initial rifting conditions suggest that extension in these margins probably started with a decoupled upper and lower crust[74,186]. Due to the ultra-slow extension, conductive cooling became important with ongoing extension, resulting in an increased localisation and coupling of deformation throughout the crust[5]. This led to the formation of large faults, which continued as ductile shear zones at depth, and pulled the ductile lower crust towards their hangingwall as observed in WAS data[79]. In numerical models[66] this pattern of extension leads to deformation fabrics that resemble the lower crustal reflectivity patterns, DR, often observed in magma-poor margins[43,79] (Figure 2a) indicating the global predominance of coupling processes at these margins.

Faulting, serpentinisation and mantle exhumation. The history of faulting, serpentinisation and mantle exhumation influences the potential of magma-poor margins to host natural H_2 reservoirs, as H_2 is a by-product of serpentinisation. Early, simple 1D models of the rheological evolution of these margins predicted that cooling leads to embrittlement of the whole crust before detachments observed in the distal margin were formed[5] (D in Figure 2a). This was thought to be a key stage during margin evolution, (also known as "coupling point"[43]), as it allows faults to reach the mantle and bring enough water to serpentinise it[5]. Serpentinisation induced weakening would have facilitated the formation of low-angle detachments in the brittle field[63]. However, new 2D dynamic simulation techniques allowing the reconstruction of seismic sections at fault-block scale (Figure 6), suggest that these detachments may form in the ductile field[66]. During extension of the ultra-thin crust (< 10 km thick), the hangingwall of deep penetrating faults becomes hotter and ductile, so that the overlying faults propagate at depth into ductile shear zones active at low angles[66] as interpreted from 3D MCS data[63]. Instead, the footwall [G] cools and becomes brittle allowing mantle serpentinisation in this area[66].

Crustal break-up and mantle exhumation. Final breakup occurs through either a pair of conjugate normal faults, forming two distal highs on each side of the new ocean floor, formed by either serpentinised mantle, as drilled in the West Iberia margin[29] or continental crust, as interpreted in Newfoundland[73]. Alternatively, interpretation from margin ophiolites in the Alps, suggests it occurs through concave-downward detachments exposing serpentinised mantle and generating and asymmetric COT[162]. Numerical models indicate that both situations are possible with normal faults forming in slightly colder conditions than concave-downward detachments[163] (Figure 5c,d, Movies S2, and S4). Mantle exhumation may occur through a series of faults that cut each other's footwall, known as flip-flop detachments, inferred to form the system of ridges observed in the COT of magma-poor margins[187].

Transition to seafloor spreading. The abrupt lateral change from exhumed mantle to oceanic crust west of Iberia indicates rapid initiation of seafloor spreading, perhaps supporting that the first oceanic crust was formed by mid-ocean ridge propagation[71]. In contrast a slow transition from exhumed mantle to thin oceanic crust as suggested for offshore central Newfoundland, might support the gradual establishment of a spreading centre by progressive thinning of the lithosphere[73,75,92] and a gradual change from tectonic to magmatic processes[188]. Such a progressive increase in magmatism would be promoted if rifting velocities increase during extension[189]. In addition, the asymmetric distribution of exhumed mantle suggests that the location of deformation may have jumped, during the establishment of the steady-state oceanic spreading centre[71,73]. The ridge propagation hypothesis would indicate that magma-poor margins form where the mantle is cold and only transitions to spreading upon ridge propagation and thus indicates strong lateral heterogeneity of the sub-lithospheric mantle. Instead, the progressive establishment of a mid-ocean ridge would suggest that mantle exposure is produced under ultra-slow velocities which increase with time as a result of extension dynamics[189].

4.2 Rifting at magma-rich margins. Excess magmatism at rifted margins is principally a function of the temperature of the mantle, but can be significantly modified by many secondary inherited attributes, such as lithosphere structure, asthenosphere composition, and processes such as small-scale convection[190].

Plume-lithosphere interaction. Observations show that magmatism is not distributed symmetrically around the LIP or assumed plume impact location, as classic models for magma-rich margins would suggest[24,191]. These variabilities point to the importance of the interaction of the plume material with pre-existing lithospheric structure in the generation of magma-rich margins[122-124]. Since buoyant plume material will flow towards the thinnest lithosphere[192], a key factor for the formation of magma-rich margins is the timing of the arrival of the thermal plume relative to the thinning of the continental lithosphere. In particular, if there is no prior extension, the mantle will likely cool significantly prior to eventual breakup leading to limited decompression melting of a depleted mantle as seen in the Seychelles-Laxmi Ridge margin[27]. In contrast, significant pre-existing thin spots that drain plume material favours forming magma-rich margins[123,124].

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

South Atlantic[102], the first phases of extension appear to be accompanied by faulting that bounds the Type I SDRs (Figure 2b). Instead, in the North Atlantic[96], SDRs are all akin to Type II and appear to be deposited during increasing subsidence from the magma injection site, without the intervention of faulting (Figure 4). Structural issues such as these will become critical if these volcanic sequences are required for future carbon sequestration.

In the South Atlantic, close to the Parana-Etendeka flood basalts, the thickness of the SDR packages is largest, conjugate margins are symmetric and the crust appears to have undergone large ductile deformation[102,119], probably related to crustal heating resulting from magmatic intrusion. With distance from the onshore flood basalts, the amount of offshore magmatism decreases[116] and the margins become asymmetric, suggesting relatively stronger (but still ductile) rheologies, as the smaller magma input provided less crustal heating.

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

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

5. Importance for the energy transition

Rifted margins are massive repositories of sedimentary, igneous and ultra-mafic rocks, are globally distributed and lie adjacent to large, coastal populations. Thus, they are likely to hold a pivotal role in the future transition to the new green economy. In the offshore environment, where there is the legacy of drilling by the hydrocarbon industry, there is a potential for a greater acceptance of the level of future exploitation that will be needed for the energy transition[193].

Whilst current Carbon Capture and Storage (CCS) projects have demonstrated the feasibility of storing CO_2 underground, to meet IPCC model targets injection operations will need to be increased by several orders of magnitude[194]. This will require exploration for new reservoirs and careful consideration of long-term storage security, especially for saline aquifers where the natural rates of mineralisation are sluggish. In this respect, magma-rich margins may prove especially promising[195] as they contain abundant, reactive rocks such as basalts, where injected CO_2 rapidly mineralizes to form stable carbonates, as has been proven for the onshore CarbFix (Iceland) and Wallula (USA) projects[196].

Similarly, large amounts of H_2 are expected to be produced at the COTs of magma-poor margins due to hydration of the exposed mantle rocks. Molecular hydrogen may become an important new fuel for internal combustion engines. Promising reserves have been detected onshore where sections of these COTs are now exposed in ophiolites, such as in the Mauleon Basin in the Pyrenees[14]. This setting portrays key ingredients for long-term trapping of H_2, which are: a seal, e.g. impermeable clay or salt; a kitchen, e.g. iron-rich rocks; an hydraulic system that favours water-rock interaction; and reservoir temperatures of ~100-200° C, where H_2 is relatively inert[14]. In the offshore environment, present-day H_2 fluxes at the COT and whether it can be trapped and form extractable reservoirs is still unknown.

Mineral deposits in the sedimentary sections of rifted margins form by hydrothermal water-rock interaction. For example, base metals accumulate where there is a transition from thick to thin lithosphere[11]. This suggests that margins that form adjacent to cratonic lithosphere might be future reservoirs of base metals. In rifts, such as the Rhine Graben, the startup Lithium de France will start geothermal exploration to extract both heat and Lithium from reservoirs[197].

In all these examples, new understanding is needed of the fault-block scale interactions between fluid flow at short time scales and the long-time scale lithosphere deformation, melt transport, and sedimentation that create the conditions favourable for storage and extraction. 3D seismic imaging, which proved so critical for hydrocarbon exploration, will need to be applied to new targets. Process-based models can help reduce risk through fitting numerical models of coupled lithosphere and basin formation to observations to map the most likely reservoir system at a large scale. In addition, the goal of understanding a given margin at the scales needed for quantitative resource prediction can be pursued by fusing simulations of rift dynamics with geological and geophysical data; commonly known as inversion and data assimilation (DA). A model with adequate dynamics could, for example, assimilate MCS images. A first step in this direction is the heuristic nudging approach used by[66] (Figure 6), which sequentially assimilates interpretations of MCS data, and shows the potential of DA towards understanding tectonics at fault-block scale. In the future, advanced DA approaches should be developed for geodynamic modelling of rifted margins.

6. Summary and future Perspectives

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

Margin studies require reference to all geophysical and geological data available, and an understanding of how they complement each other. Geophysical surveys, in particular WAS surveys, which are not typically carried out by industrial partners, need to be more common and, when existent, considered by the rest of the community. For this, it is important that the different lateral and depth resolutions of MCS and WAS data and the different information they convey is properly understood (Box 1). In addition, in sectors where the basement is laterally heterogenous, such as the COTs, instrument spacing needs to be sufficiently small, and modelling approaches exploit that resolution. One way to increase resolution is by using MCS data in between OBS instrument location, and jointly invert both MCS and WAS[71] and/or undertake downward continuation of both MCS and OBS data[198], and when instrument spacing is sufficiently close use higher resolution full waveform inversion[85].

Geodynamic models have been pivotal in documenting how different processes may explain margin architecture. Initial lithospheric strength is key in determining rift width and asymmetry, and the spatio-temporal distribution of magmatism, faulting and sedimentation. In numerical models, however, the link between the original lithospheric structure and rift geometries is established in a general way. How the structure and age of the initial fold-belt lithosphere affects rift architecture has only recently begun to be investigated[199]. 3D modelling approaches are necessary to understand how rift development is linked to a broader tectonic plate perspective[200-202].

Petroleum geoscience matured over many decades, meanwhile we are only at the beginning of exploration for hydrogen and an expanding need for base and rare-earth metals and CO_2 storage. The source rocks, fluid flow migration pathways, and reservoirs for these resources will require a renewed focus on the interactions between sediments, deformation, fluid flow and magmatism. The fusion of data and models that couple sedimentation, deformation, fluid flow and magmatism at sufficiently small scales will be pivotal to make future accurate predictions for these new storage and energy needs.

Figure captions

Figure 1. Global distribution of magma-rich, intermediate or uncertain and magma-poor margins. Oceanic crust age from[203]. Seaward dipping reflectors, SDRs, and onshore large-igneous provinces, with ages <200 Ma, along with the ages of the onset of the main large-igneous province phase are shown[204]. Magmatic margins are distinguished by their SDRs and comprise: NA: North Atlantic margins, associated to the North Atlantic Igneous province, NAIP; the Eastern North American margin and its conjugate in Africa, associated with the Central Atlantic Magmatic province, CAMP (with the African side classified as uncertain[205]); the South Atlantic magmatic margins, SA, associated with the Paraná-Etendeka; the India-Seychelles margins associated with the Deccan magmatic province, and the NW Australian margin. The Gulf of Mexico is magmatic in the East but not in the West[206,207]. Magma poor margins exhibit exhumed mantle at their continent-ocean transition. Margins where exhumed mantle at the COT has been interpreted from wide-angle data are shown. These include the West Iberia, WIM, and Newfoundland, NF, the Bay of Biscay, BB, the Goban Spur, GB, the Porcupine Basin, PB, Rockall Trough, RT, Orphan Basin, OB, Labrador, L, the southern West Greenland, WG, and the Thyrrenean, THY. The South Australian and its Antarctic conjugate have been typically considered as magma-poor[208-210] but wide-angle data has not confirmed exhumed mantle yet, so a question mark is included. Also shown are ODP drillings mentioned in the text, as black circles. Red lines are wide-angle data shown in Figure 3.

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

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

crust is found in the seaward end of West Iberia but not in the Newfoundland section. Magma-rich margins exhibit a high-velocity lower crustal body, HVLC, at their base. The thickness of the HVLC and its structural position is asymmetric in this example, indicating asymmetry in melting and extension. Towards the oceanic crust the HVLC thins. The distance over which this thinning occurs before typical oceanic crust is observed, can be quite variable on different margins. In coincident MCS data it is shown that the HVLC is overlain by seaward dipping reflections, SDRs. In the SE Greenland-Hatton Bank profiles, the oceanic crust inception is defined by the C24r magnetic anomaly[115].

Figure 4. Formation of Seaward-Dipping Reflectors (SDRs) in the South and North Atlantic (adapted from[96,102]**. a)** The geometries of the lava flows (colored blue, orange and green) depends on the tectonics (formation of large-offset faults or not), magma supply rate and interaction with standing water (lakes or sea-water – shown by the wavy line). The characteristic tilt results from either faulting or subsidence. In the North Atlantic, as the lava sequence erupted, the elevation fell from above to below sea-level. This caused two distinct lava flow accumulations separated by a hyaloclastic outer high (colored purple). In the South Atlantic, the whole sequence appears to have been erupted while above sea-level with two packages (fault bounded colored orange and non-fault bounded colored green) stacked on-top of each other. **b)** The difference in land elevation may be explained by differences in crustal thickness (dotted blue line) at the time of plume arrival, with the amount of pre-existing thinning larger in the North Atlantic[177]. The melt production, shown by red line, experienced a peak at different times during rifting, such that in the North Atlantic, the oceanic crust is thicker than the SDRs, whereas in the South Atlantic it is reverse[177].

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

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

Figure 6. Sequential nudging approach to simulate rift dynamics at fault-block scale (adapted from[66]). Comparison of **a)** two conjugate MCS lines along the West Iberia-Newfoundland margins, with, **b)** results of the modelling technique, Kinedyn[66,211]. The interpretation of fault kinematics in the MCS line is incorporated during the dynamic model run, so that the model is nudged towards the MCS data observations. Modelling at fault-block scale yields the sediment and basement temperature field during the syn and post-rift margin evolution. In this case, the model was used to understand how brittle and ductile deformation evolve with time, and their influence on detachment, D, formation, and the spatial distribution of serpentinisation. In the future, such models can be used to predict the pattern of hydrothermal circulation at margins and their consequences for element exchange and deposition of minerals within the sedimentary sections. F1 to F7 are faults, B1 to B6 are fault-blocks. Red-blue-white stripes show the deformation within the crust, blue thick line is the Moho from the model, DR are deep reflections. Shown are also dredges and ODP wells. Age of sediment is shown in colors. Model is shown after 115 Myr of evolution.

Figure 7. Conceptual margin cross section showing potential for economic resources and processes During rifting and perhaps thereafter higher than normal geothermal gradients will lead to hydrothermal circulation, releasing lithium rich brines and base metals in the sedimentary sections (A and B). In addition, sediments may host oil and gas reservoirs and may be used also as sites for CO_2 storage (C). Turbidites may pose a hazard for marine infrastructure and at the same time host unconventional oil and gas reservoirs (D). Finally, native H_2 produced during mantle hydration may be found in reservoirs which could be potentially used as energy sources (E). 1) Proximal margins and necking zone, 2) distal margin, 3) outer margin, 4) exhumed mantle, 5) oceanic crust. Note that magma-rich margins will look different and will host potential for CO2 storage in their SDRs sequences.

Box 1 | Interpreting seismic data

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

In recent years MCS data imaging has been revolutionized by using longer hydrophone streamers and improved air gun technology, increasing our ability to suppress multiples and so image reflectivity at crustal scale. However, MCS images are often shown in two-way-travel times which gives a highly distorted impression due to highly variable low-velocity water and sediment layers across a margin. Pre-stack depth migration (PSDM) was developed to address this issue, but it lacks sensitivity at depths larger than the streamer length[212]. This means that coincident PSDM and wide-angle sections, may exhibit up to some kms of difference in the depth of deep reflectors such as the Moho[55]. In addition, wide-angle seismic models sample bulk properties, thus complex reflectivity seen on co-incident MCS lines which do not present a bulk property change, will not be observed in wide-angle data.

Although, absolute compressional wave velocities, V_p, of rocks, provided by wide-angle data do not intrinsically differentiate between rock type as their values overlap in certain velocity ranges, their V_p gradients allow to distinguish basement nature (Figure B1 and Figure B2). Over the years, mapping of these V_p gradients has continuously improved with increasing quality of field data and modelling technologies. Closely spaced OBS (<10 km) in conjunction with joint inversion of coincident MCS and wide-angle data provides to date, the most accurate depiction of velocity-depth structure and Moho depth[71].

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

Glossary

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

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

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

Footwall: The block that doesn’t experiences subsidence on one side of an active fault.

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

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

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

Listric: A fault that has high-angle in its shallowest segments (~$60\text{-}30^0$) and has much lower angle at depth.

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

Magnetic anomaly lineations: Commonly interpreted to indicate seafloor spreading (oceanic crust). Formed when mafic magma cools below the Curie Point (580^0C) and takes the polarity of the Earth's field at that time.

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

Post-rift: Period of time after breakup.

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

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

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

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

Syn-rift: Period of time before breakup.

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

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

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

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

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