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Magmatic ocean-continent transitions

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Abstract :

Continental extension may, or may not, be coeval with significant mantle melting, leading to the formation of distinct types of passive margins (respectively magma-poor or volcanic), with distinct crustal structures. Especially in inter-cratonic mobile areas, magmatic breakup and development of volcanic passive margins (VPMs) may postdate the early development of a non-magmatic continental rift system. The time-span between the amagmatic and the magmatic systems may be relatively short or, conversely, long (tens of millions years). Such evolution is often associated with a significant apparent asymmetry in the wideness of conjugate VPMs. In this paper, we attempt to re-interpret the structure of three paired VPMs which developed close to, but separately from, a previous amagmatic aborted rift system. Due to opposite dips in major crustal detachment faults accommodating extension from sedimentary to volcanic stages, those composite margins tend to individualize a fault-dissected continental block (here designed as L-Block) along one of the conjugate margins. In addition to the amount and distribution of amagmatic extension, the time-span between amagmatic and magmatic extension exerts a major structural and rheological control on the final structure of the ocean-continent transitions, The finite margin geometry may be erroneously interpreted as resulting from a continuous process from hyper extension to the final magmatic breakup. However, the early syn-sedimentary extension appears, in many cases, to be low-rate or episodic. Consequently, the VPM final breakup of the lithosphere may shift away from the original amagmatic stretched area to the rifted margin.

Highlights

► Some Volcanic Passive Margins can be structurally complex at crustal scale. ► Tectonic complexity at Volcanic Passive Margins incomes from aborted rift systems. ► Magmatic breakup is away from thermally equilibrated lithosphere with thinned crust.

Keywords : Volcanic passive margins, SDRs, Polyphased lithospheric, Extension

1. Introduction

Studies of passive margins, whatever exposed or offshore located, have pointed out their bulk asymmetry (e.g. Manatschal, 2004; Brune et al., 2014). Those studies were mainly focused on passive margins for which lithospheric extension is not associated with significant volumes of melting ('non-volcanic' or 'amagmatic' or 'magma-poor') here designed as NVPMs. NVPMs usually show no (or very minor) syn-rift magmatism in the crust. Some melt could eventually be trapped within the lithospheric mantle but do not influence the crustal architecture dramatically (e.g. Müntener & Hermann, 2001). Following Wernicke's concepts (1985), NVPMs asymmetry is generally interpreted as the result of a simple shear model of plate breakup associated with the development of a trans-lithospheric detachment fault or shear zone (e.g. Lister et al., 1986; Lavier & Manatschal, 2006). At NVPMs, extreme thinning associated with rolling-hinge detachment development could be associated with the exhumation of serpentinised lithosphere mantle (Boillot et al., 1980).

By contrast, continental breakup associated with huge mantle melting does not clearly show such features (e.g. Skogseid et al., 2000; Menzies et al., 2002). VPMs (Fig. 1)have been described as relating to distinct deformation pattern and processes of continental breakup (e.g. Geoffroy, 2001, 2005; Franke et al., 2013; Quirk et al., 2014; Geoffroy et al., 2015). In deep seismic reflection profiles, those margins do not show any evidence of exhumed mantle (e.g. Zalán, 2013; Clerc et al., 2015; Paton et al., 2017). Sometimes, similar VPMs could develop at the edges of aborted sedimentary rift systems, i.e. domains which were already affected by moderate to significant amagmatic extension before the surface expression of intense magmatic activity. Assuming a NPVMs scenario, Peron-Pinvidic et al. (2013), followed by Peron-Pinvidic & Osmundsen (2016) and Peron-Pinvidic & Osmundsen (2018), proposed to integrate both pre-magmatic rift systems and breakup related volcanism within a conventional and unique hyperextended, and highly asymmetric rifted system controlled by detachment faulting. In this model, the breakup related magmatism is simply regarded as the natural ultimate result of a continuum crustal and lithospheric deformation.

Considering those views and recent controversies about the differences or similarities between NVPMs and VPMs (Gernigon et al., 2015; Osmundsen et al., 2016; Theissen-Krah et al., 2017; Zastrozhnov et al., 2018), we think that there is a need to clarify and redefine the nature and concepts of rifted margin and ocean-continent transition (OCT) when sedimentary rifts and basins are closely related in time or/and space with distal VPMs. In this paper, we first (1) recall the main characteristics of VPMs, (2) discuss both the symmetry and the definition of OCTs at the scale of thermal lithosphere, and (3) develop a different interpretation of polyphased rifted and distinct VPMs based on three case examples.

2. SDR (seaward dipping reflectors) wedges, faults and high velocity crust at VPMs

Interpretations of SDRs across VPMs (Fig. 1) have led to recent advances in passive margin studies (Planke et al., 2000; Stica et al., 2014; Geoffroy et al., 2015; Paton et al., 2017; Abdelmalak et al., 2018; McDermott et al., 2018). At VPMs, SDRs form thick wedges of seaward dipping subaerial volcanic flows emplaced at the ocean-continent transition (e.g. Hinz, 1981; Planke et al., 2000). One must distinguish between inner SDRs, which develop over the necked parts of VPMs associated with a strong Moho slope (e.g. Geoffroy, 2005; Fig. 1), and outer SDRs which develop over a gently dipping to flat-lying Moho, outward from the syn-magmatic necking zone (Planke et al., 2000; Franke et al., 2010; Geoffroy et al., 2015; Paton et al., 2017). Isostatic models were proposed for similar basaltic wedges of Mio-Pliocene age dipping towards the active rift zone in Iceland (e.g. Pálmason, 1980). Those models were extended to SDR wedges located at VPMs (e.g. Mutter, 1985; Buck, 2017; Hjartarson et al., 2017). However, some authors early pointed out the striking analogy in seismic reflection profiles, between SDRs and syn-sedimentary rollover that developed over large crustal-scale detachment faults (e.g. Gibson & Love, 1989; Tard et al., 1991). New insights on VPMs formation came from a detailed field investigation of onshore exposed SDRs and their underlying dyke injected crust (Geoffroy et al., 1998, 2001; Brooks,2011; Abdelmalak et al., 2015, 2017), followed by convergent interpretations of recent long offset seismic reflection data extending down to Moho depths (e.g. Rey et al., 2008; Stica et al., 2014; Quirk et al., 2014; Zalán, 2015; Pindell et al., 2014; Clerc et al., 2015; Geoffroy et al., 2015; Paton et al., 2017; McDermott et al., 2018). Those seismic data confirm that inner SDRs develop syn-tectonically within the continental thinned and stretched area of many VPMs (Fig. 1). This conveniently explains the local contamination of inner SDR mafic magma by continental crust (e.g. Larsen et al., 1998; Saunders et al., 1997, 1999; Meyer et al., 2007; Abdelmalak et al., 2016), ruling out the classic interpretation of SDRs as a diagnostic of oceanic crust domains (e.g. Mutter et al., 1982). Continentward dipping detachment faults bounding inner SDRs appear to accommodate significant upper and syn-crustal thinning and stretching in the magmatic necking zone. In addition, SDRs formation is contemporaneous with horizontal magma dilatation associated with huge dyking controlled by localized crustal magma reservoirs (Bromann-Klausen & Larsen, 2002; Callot & Geoffroy, 2004). Convergent observations suggest that sub-inner SDRs detachment faults root on top of a ductile and layered middle or lower continental crust (Geoffroy et al., 2015; Clerc et al., 2015). This lower crust displays high-Vp (P wave velocities), in possible connection with large volumes of intruded mafic magma, mainly injected as sills (White et al., 2008). A synthetic model for conjugate VPMs has already been proposed (Fig. 1; Geoffroy, 2010; Geoffroy et al., 2015), showing strong contrasts with the shear models proposed for NVPMs (Brun & Beslier, 1996; Reston et al., 1996; Lavier & Manatschal, 2006).

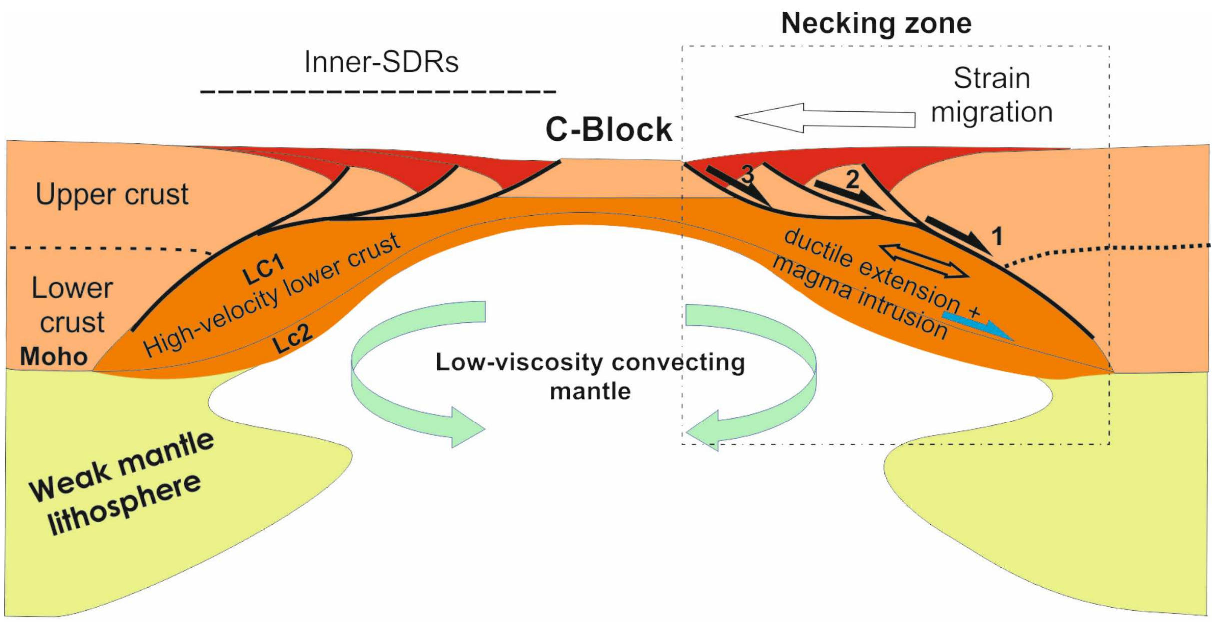

Figure 1. Conceptual model for conjugate VPMs at their initial stage of development (inner SDR stage; modified from Geoffroy, 2010, and Geoffroy et al., 2015). The outward dip of major detachment faults controlling inner SDRs development is explained by two possible mechanisms: mechanical coupling of the upper crust with oceanward flowing lower crust (Brun & Beslier, 1996;Gac & Geoffroy, 2004; Huismanns & Beaumont, 2011) and/or continentward gravity collapse of the upper continental crust away from the exhumed lower crust, in a way similar to post-collisional tectonics (Tirel et al., 2008; Geoffroy, 2010). The C-Block has firstly been defined by Geoffroy (2010), and mechanically reproduced by Geoffroy et al. (2015). It is the common footwall of the continentward dipping detachment faults from each conjugate volcanic passive margin.

3. Symmetry at conjugate VPMs

Seismic refraction profiles suggest that the wideness of the stretched and thinned continental crust at conjugate VPMs may be different than those illustrated in Figure 1. In the NE-Atlantic, such is the case for the SE-Greenland/Hatton (Fig. 2a and 2b; White & Smith, 2009; Funck et al., 2016) and the NE-Greenland/Norwegian conjugate VPMs (Voss & Jokat, 2007; Mjelde et al., 2005; Funck et al., 2016).

The degree and origin of asymmetry at VPMs depend on how the ocean-continent transition is defined. One should distinguish the continent-ocean boundary (COB) from the ocean-continent transition (OCT). The COB at VPMs is the location of the earliest MORB (mid-ocean ridge basalts)-type oceanic crust except if the MORB basalt overlaps the continental crust, provided it may be clearly identified (e.g. Eagles et al., 2015). Like NVPMs, the COB at VPMs should represent the true locus of continental lithosphere breakup. The OCT is interpreted as a transitional domain consisted of extremely thinned and stretched continental lithosphere bounding the COB inward. In deep seismic surveys, true oceanic crust is often marked by a clear Moho showing a constant crustal thickness (on average 6-7km) including sub-horizontal lava flows in the uppermost section (Karson et al., 2002). The symmetry of VPMs depends on both the true location of the COB and on the wideness of the OCT on both margins. In this context, the time and space occurrence of the lithospheric extension which led to continental breakup must be considered carefully. Particularly, one should consider if amagmatic continental extension predated the onset of syn-magmatic extension in a continuum of tectonic and lithospheric evolution (e.g. Gernigon et al., 2014; Theissen-Krah et al., 2017). As seen hereafter (section3.b), this crucial issue cannot always be resolved by the study of crust thickness variations alone.

a) Asymmetry and continent-ocean boundary location

In the NE-Atlantic (Fig. 2a), it is acknowledged that magnetic anomaly 24 marks the earliest oceanic accretion (Talwani & Eldholm, 1977; Gaina et al., 2009). Considering this to be exact, the conjugate VPMs display a moderate asymmetry, associated with different Moho slopes in the necking area of each paired margin (Fig. 2b, box 1).

This slight asymmetry (e.g. Smith et al., 2005) appears to be correlated to a thinner continental crust east of Greenland before syn-magmatic Paleocene-Eocene extension. Although shallow sedimentary basins predating the Aptian are proposed to exist immediately along shore SE-Greenland and in the outer Hatton Basin (Gerlings et al., 2017), most of the inherited attenuated crust and main sedimentary basin developed in a broad area extending between Hatton Bank andBritish Islands (Fig. 2a). Crust thickness variations in the Hatton-Rockall area appear to be related to Mesozoic thinning and stretching (Fig. 2b; O'Reilly et al., 1996; Morewood et al., 2005). As elsewhere in the NE-Atlantic, the continental lithosphere was involved in a complex tectonic history including large-scale wrench tectonics during the Paleozoic (Ziegler, 1989; Chenin et al., 2015) followed, during the Mesozoic, by distinct stages of lithosphere stretching and thinning in E-W to NW-SE directions (e.g. Skogseid, 1994; O'Reilly et al., 1998; Doré et al., 1999; Kimbell et al., 2004). It is generally acknowledged that the latest stages of amagmatic extension, which occurred here at the very end of the Jurassic until the earliest Cretaceous, were also the most vigorous, in terms of lithosphere thinning (Brekke, 2000; Scheck-Wenderoth et al., 2007, Faleide et al., 2008, Osmundsen & Ebbing, 2008). This period was responsible for extreme crust reduction along a rift system extending from Rockall, to the south, to Lofoten to the north (Doré et al., 1999; Faleide et al., 2008, Osmundsen & Ebbing, 2008; Færseth, 2012; Lundin & Doré, 2011). Within the Hatton-Rockall area (Fig. 2) the estimated □□thinning factors reached a maximum of 6, probably associated, with serpentinization of the underlying and locally exhumed mantle (O'Reilly et al., 1998, Joppen & White, 1990). This extension lasted around

30Ma but never reached the breakup stage (i.e. oceanic crust accretion). In the Hatton-Rockall area (Fig. 2), this major event was likely followed by thermal subsidence throughout the Cretaceous until the Cenozoic syn-magmatic extension which led to continental breakup and accretion of oceanic lithosphere.

In the NE-Atlantic, the precise age of early oceanic crust is debatable, especially to the south of the Greenland-Iceland-Faroe-Ridge (GIFR, Fig. 2a). Erroneous interpretation of the earliest magnetic anomalies chrons may have huge consequences on the degree of asymmetry of conjugate VPMs (Fig. 2b). Outer SDRs are recognized off the SE-Greenland margin, extending to magnetic anomaly chron22 (e.g. Dahl-Jensen et al., 1997). However, the linear magnetic anomalies observed above the SDRs are not necessarily diagnostic of true oceanic crust, at least in the way seafloor oceanic crust and spreading are usually defined (e.g. Karson et al., 2002). It must be noted that SDRs have never been identified in active oceanic ridges, deep drillings of oceanic crust or ophiolitic complexes, whatever the rate of spreading. In oceanic domains, linear magnetic anomalies should solely be considered as a diagnostic of extrusions emplaced in a sub-symmetrical magma accretion context, the way VPMs also develop (Geoffroy et al., 2015). Seismic refraction data also show that Moho beneath the SE-Greenland outer SDRs is abnormal both in depth and slope (Fig. 2b; Holbrook et al., 2001; Hooper et al., 2003; Funck et al., 2016). Assuming that those outer SDRs emplaced alternatively in a continent-derived transitional area, this would not only extend the continental domain seaward but also increase the degree of asymmetry of the conjugate VPMs (Fig. 2b, box 2).

b) The general issue of passive margin (lithospheric) definition

Recent studies have claimed that the magmatism at the COB in the NE-Atlantic and elsewhere, was the simple result of a very long period of extreme thinning and stretching of the continental lithosphere, succeeding in final syn-magmatic breakup (e.g. Peron-Pinvidic et al., 2013; Peron-Pinvidic & Osmundsen, 2016, 2018). Those authors promote a classic model of continental lithosphere extension mostly derived from decades of geophysical and geological observations carried out along the Iberia-Newfoundl and NVPMs and/or derived from the inverted Jurassic and fossil passive margins from the Alps (e.g. Manatschal, 2004). According to Peron-Pinvidic et al. (2013), the thick NE-Atlantic Cretaceous basins, formerly considered as thermal post-rift, following the Late Jurassic-Early Cretaceous extension, should be interpreted as a syn-tectonic sag-type basin overlying an exhumed and serpentinized mantle. According to this model, a “magmatic pulse”, corresponding to our distinct VPMs would be responsible for the final lithosphere breakup. This evolutionary model for the NE-Atlantic was formerly established for the Vøring margin (Fig. 2a), which was previously considered by many authors as the archetype of “volcanic rifted margins”, including in their structure the nearby Mesozoic rift system (Skogseid et al., 2000; Zastrozhnov et al., 2018). However, Geoffroy (2001), Gernigon et al. (2014) and Theissen-Krah et al. (2017) pointed out the potential misconception in doing so. Off Norway, the major Mesozoic rift (with hyper-extended crust) is effectively very close (and parallel) to the Møre and Vøring VPMs. The buoyant volcanic plateau interpreted as a continental basement (Vøring High; Mjelde et al., 2005) divides the Mid-Mesozoic rift from the VPM. Further south, a similar rift system (e.g. Rockall Trough) is located 500km away from the Hatton VPM, and can hardly be considered as belonging to the same extensional system (Fig. 2a and 2b; e.g. White et al., 1987; White & McKenzie, 1988). Following the concepts of Peron-Pinvidic et al. (2013), this would make the conjugate margins unusually wide and highly asymmetric (Fig. 2b, box3).

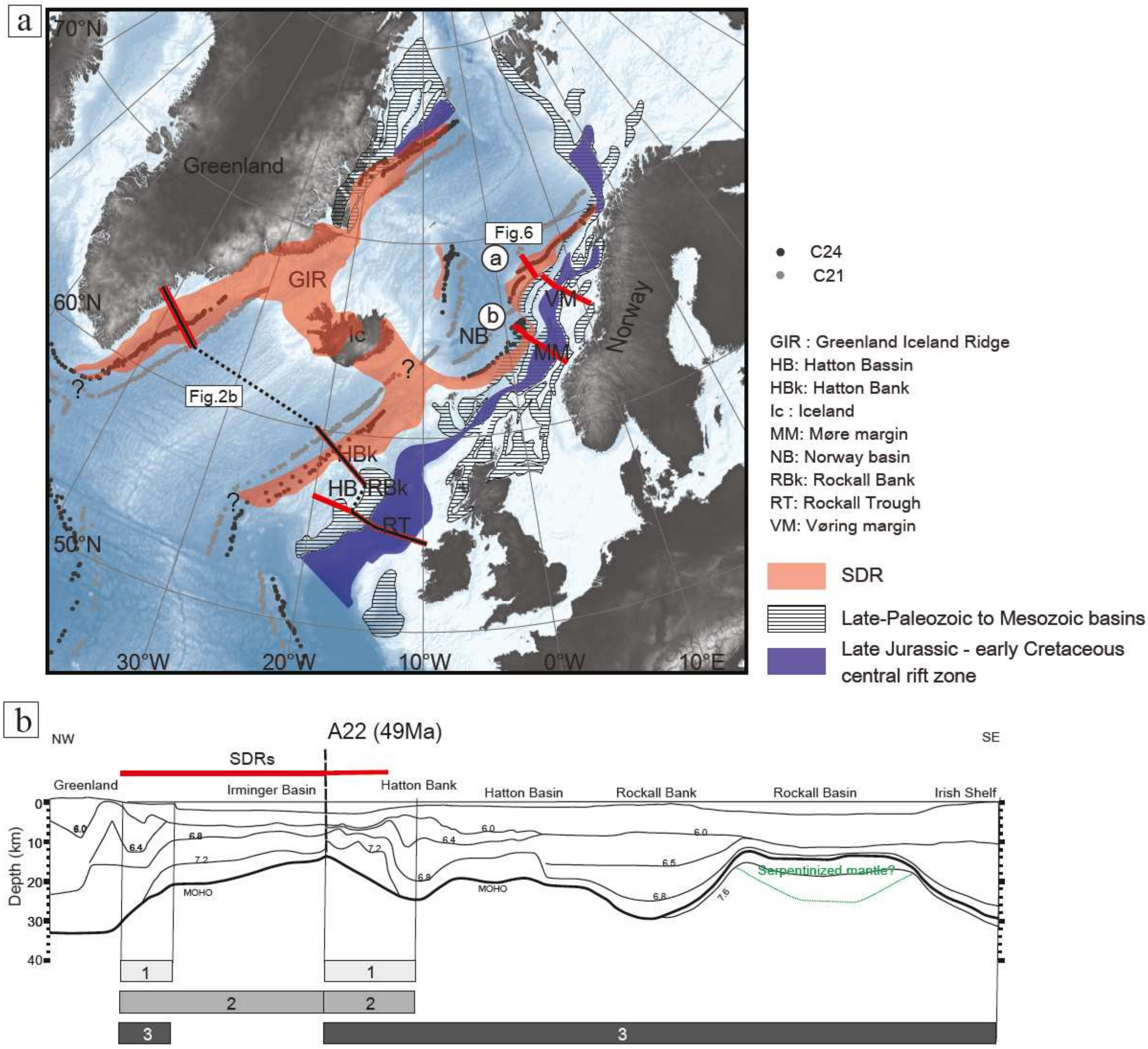

Figure 2. a. Bathymetric map of the NE-Atlantic with location of the studied profiles. Superposed in purple is the area of finite □□ thinning factor exceeding 4 (continental crust reduction down to 7 km, excluding magma addition) in the Late Jurassic/Early Cretaceous rift system, and in red the extension of SDRs (schematic). b. Crustal section and seismic velocities from SE-Greenland to Hatton at C21 (from Funck et al., 2016). Lower boxes 1, 2 and 3 refer to conjugate passive margins according to different hypotheses on margin definition (see text).

We performed a simple but robust thermal model of lithosphere thickness of the Hatton-Greenland paired VPMs with time, from 130 Ma to 63 Ma (earliest Paleocene) (Fig. 3). In this model, we considered that the major lithosphere thinning occurred during the Late Jurassic-Early Cretaceous extension and that most of the finite crust thinning before the Paleocene was related to this period. Although this extension lasted approximately 30 Myrs, we considered it as instantaneous, pure shear and depth-independent, in accordance with the McKenzie (1978) model. Note that a more realistic lower extension-rate or depth-dependent thinning would strengthen our conclusions by increasing the depth of the initial lithosphere-asthenosphere boundary (LAB). Therefore, the initial LAB isotherm was solely controlled by the □□ thinning ratio inferred from crust thickness, considering a reference crust thickness of 35 km.

The 2D heat diffusion equation was solved in a Cartesian frame (1599×300 km) with a finite difference method, using a resolution of 1.5 km in the horizontal and vertical directions. An explicit forward difference scheme was used for the time evolution, with a time-step of 10 kyrs. We used an isotropic thermal conductivity k=3 $W.m^{-1}$ $.K^{-1}$, a specific heat capacity cp=1200 $J.kg^{-1}.K^{-1}$ and a density ρ=3300 $kg.m^{-3}$. Decreasing the thermal conductivity to k=2 $W.m^{-1}$ $.K^{-1}$ or increasing the bottom temperature to 1400°C, would lower the depth of the 1300°C isotherm after 67 Myrs by a maximum of 22 km and 16 km respectively. Increasing the thermal conductivity, decreasing the bottom temperature or adding internal radioactive heat would all deepen the final isotherm.

We show that the NE-Atlantic lithosphere, at the time of the early magma emplacement, had a thickness exceeding 100km (Fig. 3). The main variation in thermal lithosphere thickness was between the cold and underformed Greenland craton and the lithosphere to the east, which was reactivated throughout the Phanerozoic. To the East, the plate shows more variations in crust thickness than in lithosphere thermal thickness.

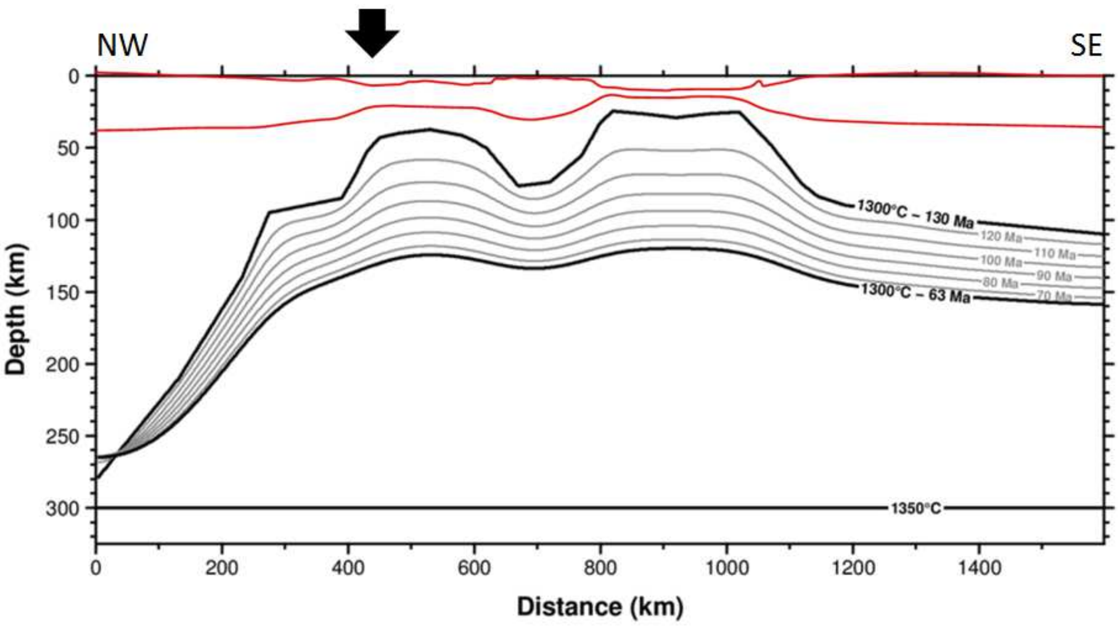

Figure 3. NE-Atlantic thermal lithosphere thickening with time from Greenland to Hatton, starting from 130 Ma to immediately before the onset of Tertiary mantle melting (63 Ma). The initial 1300°C isotherm takes into consideration instantaneous thinning due to extension (McKenzie, 1978) at 130 Ma, which is considered as a convenient upper-bound for the results. The thickness of the post-Caledonian continental crust is considered to be 32 km (Cloetingh et al., 2010). Taking into account an initial 25 km in thickness would only shift the thermal boundary layer downward, thus making the lithosphere still stronger. The thermal lithosphere thickness is inverted from a simple 1D diffusion model excluding any lateral heat diffusion. Black arrow: area of Eocene breakup. Red line: continental crust.

Many factors control the strength of continental lithosphere (e.g. Burov, 2011). A thermally equilibrated continental lithosphere with thin crust is much more difficult to extend due to a lower quartz/olivine ratio (e.g. Kusznir & Park, 1987). Although not fully equilibrated, the Paleocene NW-Europe lithosphere thus had strongly integrated strength, due to extreme silicic crust reduction in places (Fig. 2b). No "whole lithosphere failure" could be expected for such thermally re-equilibrated lithosphere with thin continental crust. The Greenland craton, albeit with a thicker crust, was too old and thus too cold to be extended. It appears from observation and modelling that renewed stretching and thinning of the plate focused preferentially within the lithosphere bearing the (relatively) thicker crust and the most significant lateral variation in lithosphere thickness, i.e. along Greenland (black arrow; Fig. 3). This is in favor of mechanisms involving 3D edge-convection (e.g. King & Anderson, 1998) as independently suggested by 3D model of VPMs formation (Callot et al., 2002).

This robust rheological evidence suggests that the NE-Atlantic volcanic margin asymmetry is limited to the area with syn-rift magmatism (excluding box3 from Fig. 2) and that the degree of VPM asymmetry solely depends on the consideration, or not, of outer SDRs as overlying continent-derived lower crust (Fig. 2b, boxes1 and 2).

4. Crustal structure of polyphased VPMs

A similar disposition of continentward dipping faults exists at both facing VPMs (Fig. 1). No evidence is met from crust sections down to the Moho of any trans-lithospheric decoupling, coeval with magmatic continental breakup. However, like in the NE-Atlantic (see above) continental breakup and VPM edification may occur within suture lines of inter-cratonic areas which experienced long-term unsuccessful stages of amagmatic extension before syn-magmatic breakup (Fig. 2; Chenin et al. 2015). Including those rifts, the crustal structure of those "polyphased margins" should theoretically be more complex than the conceptual representations for each type presented so far. We hereafter present three case studies illustrating this complexity.

4.1 The W-Greenland margin

The W-Greenland margin (Figs. 4a and 4b) is another case example of polyphased extension being first amagmatic and becoming fully magmatic. From NW-Disko Island to Svartenhuk, clear inner SDRs developed from Paleocene to Eocene and are presently largely exposed onshore (Geoffroy et al., 1998, 2001; Chauvet et al., submitted). The amagmatic stage corresponds to the development of a Cretaceous continental rift within Baffin Bay, from Disko Bay to Melville Bay (Fig. 4a; Whittaker et al., 1997; Schenk, 2011) associated with moderate lithosphere thinning and stretching (Gregersen et al., 2013). Though the kinematic of this margin is complex and could be influenced by the Ungava transfer system, the stretching direction in Svartenhuk segment is clearly orthogonal to the extensional axis and shows no obliquity (Abdelmalak et al., 2018). South of the Ungava transfer system, significant lithosphere thinning and stretching occurred in the Labrador Sea with possibly mantle exhumation (e.g. Chian & Louden, 1994). Further north, the Baffin Bay Mesozoic sedimentary basin is bounded everywhere over the Greenland basement by a major dislocation dipping seaward, often adopting a bayonet-shape, and dominantly trending NNW-SSE (Figs. 4a and 4b). Because no extension is evidenced in its footwall, this fault is best interpreted as the break-away of the Mesozoic extensional system (Fig. 4c).

In Baffin Bay, the extension was discontinuous with time during Cretaceous (Gregersen et al., 2013). The most important pulse in stretching probably occurred during the Albian until, possibly, the lower Campanian (e.g. Chalmers et al., 1999; Dam et al., 2009). The syn-sedimentary pattern of normal faults and associated tilted blocks show that this early extension trended close to N-S in the Nuussuaq-Svartenhuk area (Abdelmalak, 2010), in possible connection with a ridge-ridge-ridge (RRR) triple junction. It was followed by a stage of thermal subsidence and sag basin formation. A later pulse of stretching occurred during the Maastrichtian in the NE-SW trend, predated the onset of regional volcanism during the Paleocene. To the east of the W-Greenland VPM, the mid-Paleocene volcanic traps (could be considered as inner SDRs, Fig. 4b) overlying the exposed W-Greenland Mesozoic basin were not affected by any significant tectonic stretching or thinning (Fig. 4b). Considering the nearby Greenland crustal thickness (~40km in average; Steffen et al., 2017) and the thinnest crust at the most distal part of the sedimentary basin (Chalmers et al., 1999), the crustal thinning factor □ associated with finite Mesozoic extension did not exceed 1.7 before the onset of magmatism. This estimate does not take into account possible dilatation of the basement lower crust by mafic sills (Geoffroy et al., 2015). However, it is comparable to the maximum ~2.3 □□crustal thinning deduced further north in the amagmatic Melville Bay (Fig. 4a) combining the EIGEN-6C4 gravity model (Steffen et al., 2017) to seismic refraction data (Altenbernd et al., 2014). The conspicuous W-Greenland (Disko-Svartenhuk) inner SDRs developed during the Paleocene-Eocene period. It represents the most proximal part of the VPM system which certainly extends far offshore (Fig. 4a). Both sedimentary and volcanic development of this VPM is controlled by a major syn-magmatic crustal scale continentward dipping detachment fault, the Arfertuarssuk Fault (Figure 4c and 4d; Geoffroy et al., 2001). The inner SDR is dissected by a number of syn-constructional faults and is clearly associated with important tectonic stretching and thinning, in addition to magma dilatation through dyking (Geoffroy et al., 2001; Abdelmalak et al., 2012).

Geologically, the W-Greenland innermost SDR developed significantly to the west of the exposed sedimentary basin and the associated border faults (Figs. 4b and 4c). The syn-magmatic Paleocene-Eocene extensional strain field has thus shifted seaward of the main Late-Jurassic/Early-Cretaceous basin. The true extent of the sedimentary basin beneath the inner SDRs is unknown but the pattern reflects at least a narrowing of the lithosphere necking zone with time (Figure 4c).In addition, the pattern of major faults and hanging wall structure differed dramatically from the sedimentary to the magmatic stage. The sedimentary basin border fault dipping seawards, is steep (e.g. Abdelmalak, 2010), and its footwall is cross-cut by seaward dipping synthetic faults bounding rotated rigid blocks (Fig. 4c; Abdelmalak, 2010). The SDR bounding fault is continentward, listric, and the tectonized hangingwall is flexed as a rollover anticline over the fault (Fig. 4c). The distance between the sedimentary basin border fault and the SDR-bounding detachment fault varies along strike, from approximately 100 km to the south, to 50 km to the north, isolating a clear upper crustal block (Fig. 4c). Due to seaward rigid rotation, the dip of Cretaceous faults beneath the SDR is thought to be almost vertical.

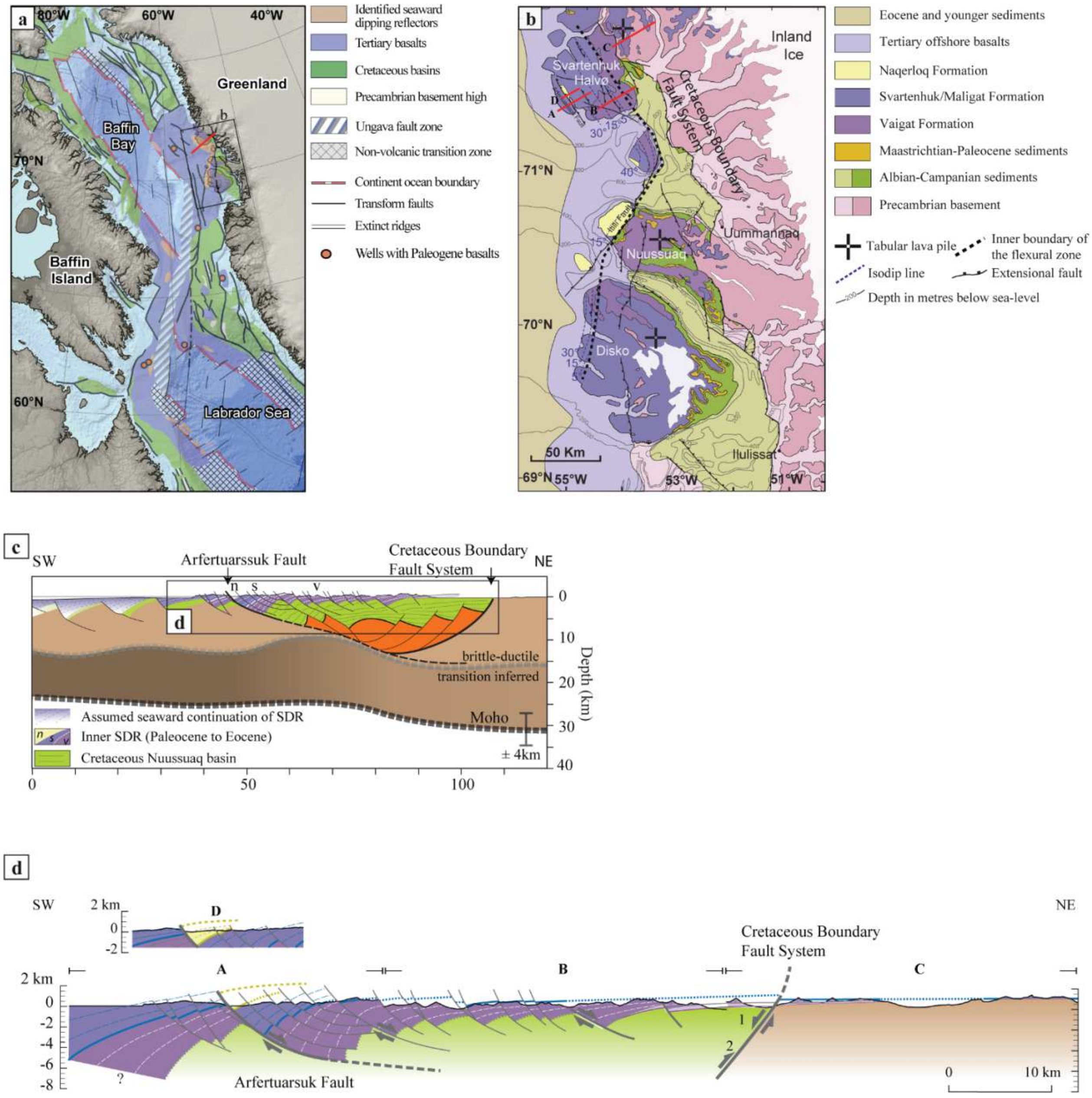

Figure 4.a. Bathymetric map of the Labrador-Baffin area with overprint of Cretaceous sedimentary basins and Palaeocene/Eocene basalts. b. Geological map of the Svartenhuk-Disko area (Abdelmalak et al., 2012). c. Cross-section

(composite section including profiles A, B, and C in figure 4b) across the Svartenhuk peninsula. d. Enlargement of the Eocene-dated formation located along the Arfertuarssuk fault.

4.2 The Laxmi system

In the west of India (Figs. 5a and 5b), the aborted Laxmi volcanic rift (e.g. Calvès et al., 2011; Misra et al., 2015; Nemčok et al., 2016) predated the post-Deccan breakup of the Arabian Sea in-between the Seychelles and the Laxmi ridges (Figs. 5a and 5b). Earliest ocean floor accretion in the Arabian Sea is thought to have occurred during Paleocene (Chron 28n), postdating the Deccan Traps (Fig. 5b; Naini & Talwani, 1982; Devey & Stephens, 1991; Collier et al., 2008). East of the Arabian Sea, the Laxmi basin clearly shows a pair of conjugate VPMs showing both inner and outer SDRs and a C-Block (Figure 1; e.g. Krishna et al., 2006; Guan et al., 2016; Nemčok et al., 2016). We reevaluated the crustal structure of this basin using the ION Geophysical IndiaSPAN™ long-offset seismic reflection database (Guan et al., 2016).

The Laxmi VPM developed at the top of a continental crust which was apparently previously thinned and stretched before the onset of SDR-related volcanism (Roberts, 2008). On both sides of the Laxmi basin, respectively the inner part of the Laxmi Ridge (Fig. 5b; Guan et al., 2016; Nemčok et al., 2016) and the sub-Deccan traps crust on the western Indian margin (31.5 km; Kaila et al., 1981), the Moho is clearly imaged and no significant syn-magmatic stretching/thinning appears to exist. Considering the continental crustal thickness of both the two sides and the uncertainties on Moho depths, we tentatively infer a pre-magmatic □crustal thinning factor □□of 2.5 +/-1.

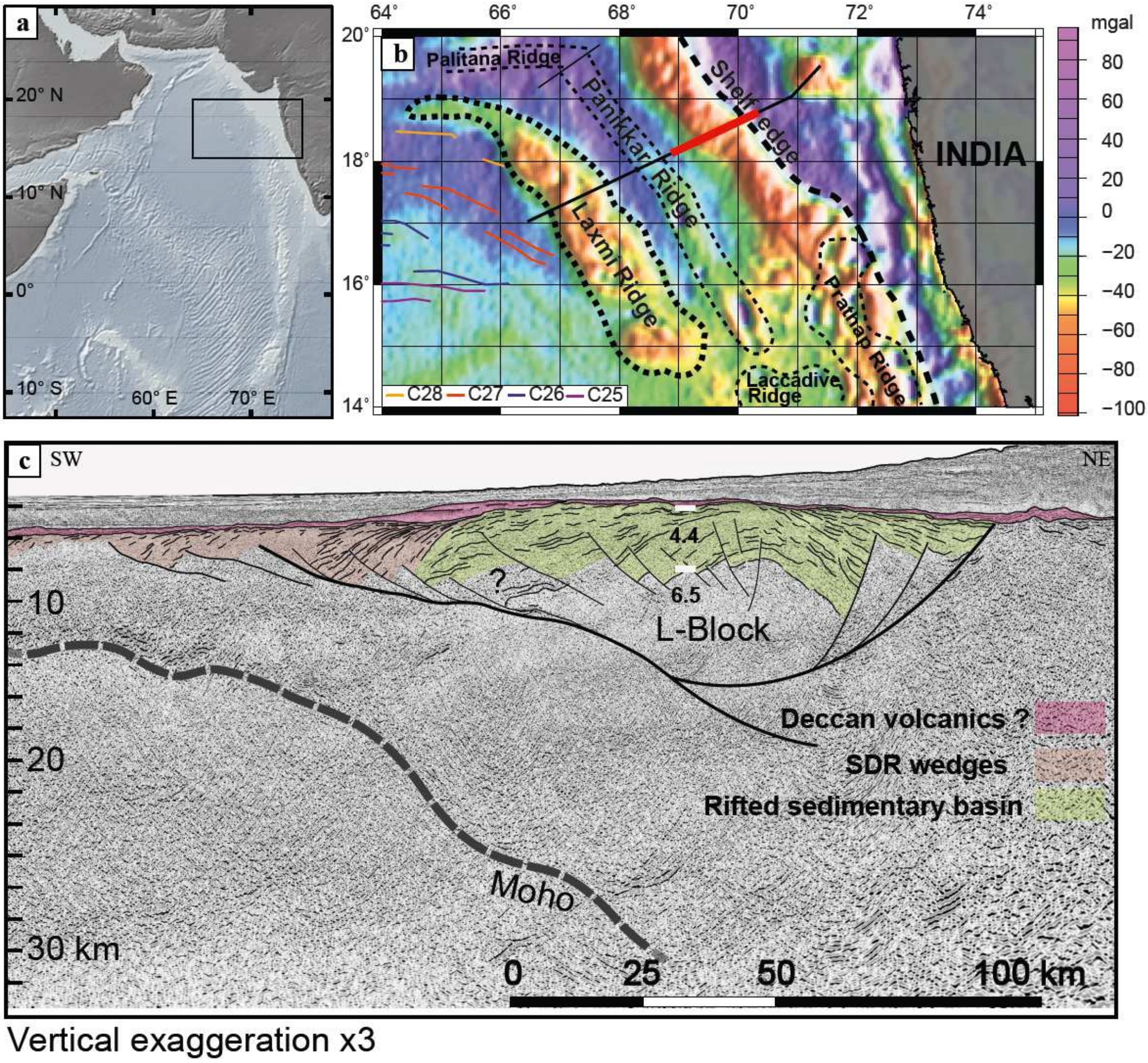

Figure 5. a. The Laxmi basin in the framework of the Indian Ocean; b. Physiography of the Laxmi basin from free-air gravity (WGM2012, http://bgi.omp.obs-mip.fr/) and location of Fig. 5c line drawing; c. Interpretation of the eastern part of the ION Geophysical IndiaSPAN™4000, enlarged. In green: postulated Cretaceous sediments, brown: inner SDR wedges in the Laxmi basin, pink: post-SDRs basalts (mainly), possibly corresponding to Deccan traps. Available Vp values are shown in the basin (from Naini & Talwani, 1982).

Of particular interest is the identification in the eastern Laxmi basin (ION Geophysical IndiaSPAN™ lines 4000 and 5000), of a rollover anticline structure with strong disrupted reflections, developed over a prominent westward dipping (i.e. seaward dipping) detachment fault (Fig. 5c). This structure has been previously interpreted as SDRs wedges (Misra et al., 2015). However, basing ourselves on existing P-wave velocities not exceeding 4.4 km/s (Fig. 5c;Naini&Talwani, 1982), this wedge is probably constituted of sill-intruded sediments. A similar interpretation arose from distinct seismic data to the south of the studied profiles (Roberts, 2008). The age of this half-graben basin is unknown but clearly predates the inner SDRs emplacement. Together with the inner SDRs, the basin is also covered by post-SDR reflective layers of possible volcanic nature, passing laterally seawards to transparent seismic facies (clays? volcanoclastics?) (Fig. 5c). Roberts et al. (2008) interpreted those highly reflective layers as possibly representing basalts from the Cretaceous-Tertiary transition Deccan traps. Incidentally, this interpretation would suggest that the Laxmi paired VPMs are related to a mantle melting event occurring before the onset of the Deccan-Seychelles LIP (see also Calvès et al., 2011). This would strongly challenge the timing of the Deccan plume hypothesis. In any case, our postulated sedimentary basin appears to be pre-Tertiary. During the Mesozoic, the nearby Indian basement is actually known to have experienced continental rifting along or across inherited Precambrian trends (e.g. Biswas, 1999; Sheth, 1999). For example, rifting along the NNW-SSE Dharwar trend (i.e. parallel to the western coast of India) resulted in sedimentation in the Cambay rift system from the Early Cretaceous (Biswas, 1982) or from the Late Cretaceous (Tewari et al., 1995).

The East Laxmi margin's geometry is thus the combination of two successive (possibly pre-Deccan) extensional events associated with major faults with opposite dips (Fig. 5c). Like for the W-Greenland margin, the upper plate of the syn-magmatic continentward dipping fault system associated with inner SDRs is also the upper plate of the seaward dipping syn-sedimentary detachment fault.

4.3 The Vøring and Møre Margins

Like for the Hatton VPM, the Norwegian VPMs developed away from, but much closer to the hyperextended NE-Atlantic Mesozoic proto-rift (Fig. 2a). As elsewhere in the NE-Atlantic and Baffin Bay system, syn-magmatic (syn-SDRs) extension was Paleocene to Eocene in age (e.g. Skogseid et al., 1992; Mjelde et al., 2001; Gernigon et al., 2015; Schiffer et al., 2018). Like elsewhere in the NE-Atlantic (section 2.a), this breakup occurred within a previously stretched and thinned continental crust which experienced major wrench and extensional tectonic events (Ziegler, 1989; Doré et al., 1999; Skogseid et al., 2000; Mosar et al., 2002). However, to the north of the GIFR (Fig. 2a), a distinct characteristic of North American-Eurasia plate separation is that rifting occurred along the inherited

Caledonian lithosphere. This lithosphere is intrinsically weaker, as suggested by lower Te (effective elastic thickness) values compared to Greenland-Baltica Archean/Proterozoic lithospheres (e.g. Pérez-Guissinyé & Watts, 2005). Gravitational continental crust thinning affected a broad area during the end of the Silurian and early Devonian (Voss & Jokat, 2009; Osmundsen & Andersen, 2001). Onshore and offshore observations from NE-Greenland and Norway suggest that some lithosphere stretching and thinning phases occurred during the Permo-Triassic and Jurassic within a diffuse rift system between Norway and Greenland (Brekke et al. 2001; Osmundsen et al. 2002; Müller et al. 2005; Hamann et al. 2005 ; Guarnieri et al., 2017). Like for the Rockall Trough (Fig. 2a) the most significant stretching and thinning in Vøring and Møre rifts (Fig. 6a) occurred at the very end of the Jurassic and Early Cretaceous (Ziegler, 1988; Lundin & Doré 1997; Roberts et al. 1999; Shannon et al., 1999; Zastrozhnov et al., 2018), with extreme crust reduction locally (Osmundsen et al., 2002; Funck et al., 2016; Maystrenko et al., 2017). However, in both of those rifts, Moho depth and P-waves velocities with depth (Funck et al., 2016; Fig. 6b) strongly question the generalized exhumation of the lithospheric mantle as recently suggested (Peron-Pinvidic et al., 2013; Osmundsen et al., 2016; Peron-Pinvidic & Osmundsen, 2016, 2018). A 6.2 to 6.4 km.s-1 layer is indeed observable beneath the oldest sediments (including Devonian-Carboniferous?) suggesting the existence of a continental basement, non-serpentinized mantle (Fig. 6b). The following Cretaceous period was mainly associated with thermal subsidence with, in most places, a clear unconformity of the Lower Cretaceous over the syn-rift Late Jurassic/Earliest Cretaceous sediments (Brekke, 2000; Osmundsen et al., 2002). However, some renewed minor extension locally occurred during the Cretaceous, notably during the Aptian-Albian (Lundin & Doré, 1997; Zastrozhnov et al., 2018). A clear exception is the Gjallar Ridge, a NNE-SSW trending Late Cretaceous doming structure located to the east of the Vøring High (Fig. 6a). Combining seismic and core-log data, Ren et al. (1998) and Gernigon et al. (2003, 2004) pointed out that extension developed here during the Upper Cretaceous (e.g. Mid-Campanian to Maastrichtian), some 10 to 15 Myrs before the Paleocene-Eocene extension. Significantly, the major syn-sedimentary faults associated with this extension are of detachment types, and dip seawards (Gernigon et al., 2004). Most of them root at relatively shallow levels (c.a. 8 km beneath sea bottom), along an apparent decoupling level within the lower Cretaceous. Interestingly, this system seems to be correlated to deeper and apparently steeper brittle structures (still dipping oceanwards) which developed at the top of the dome-shaped “T-reflection” (Gernigon et al., 2003, 2004; Abdelmalak et al., 2017). This T-reflection corresponds to the top of a high velocity lower continental crust (HVLC) with velocities between 7.2 and 7.4 km.s-1 (Fig. 6b, Gernigon et al., 2004; Funck et al., 2016). This HVLC with small vertical velocity gradients, is common in NE-Atlantic and elsewhere and could represent the ductile middle crust (LC1 in Fig. 1) identified by Clerc et al. (2015) and Geoffroy et al. (2015), corresponding to a probable mafic magma intruded lower continental crust (White et al., 2008; Geoffroy et al., 2015). Beneath the Vøring and Møre margins, this HVLC largely extends seaward and continentward away from the upper section with SDRs. The HVLC is clearly in congruent association, in the upper crustal section, with evidences of mafic magma over-productivity, during the Tertiary. Seaward, a thick mafic crust of uncertain nature is located beneath the Norway and Lofoten basins (Funck et al., 2016) and, continentward, a pervasive Tertiary mafic sill system cross-cut the Cretaceous of the Vøring basin (Fig. 6a; Osmundsen & Ebbing, 2008). A comparable onshore exposed sill system is injected in the post-rift Cretaceous basin of NE-Greenland, overlying a high velocity lower continental crust (Schlindwein & Jokat, 1999).

The Vøring and Møre VPMs also developed outward from the previous rift and are similar to the previous case examples. However, the evolution of those two systems differs in two aspects: (1) the amount of pre-VPM crustal thinning and, (2) the large time-span between the major amagmatic extension and syn-magmatic stretching/thinning. At the scale of the Vøring and Møre sedimentary rifts, the westward dipping faults have probably consisted in the break-away of the major (and aborted) detachment fault systems reaching down to the brittle-ductile crust interface or down to the Moho (Osmundsen & Ebbing, 2008) during the Late Jurassic/Early Cretaceous extension. The basin was subsequently infilled during the thermal subsidence of the narrow Cretaceous rift and the VPMs are formed further seawards.

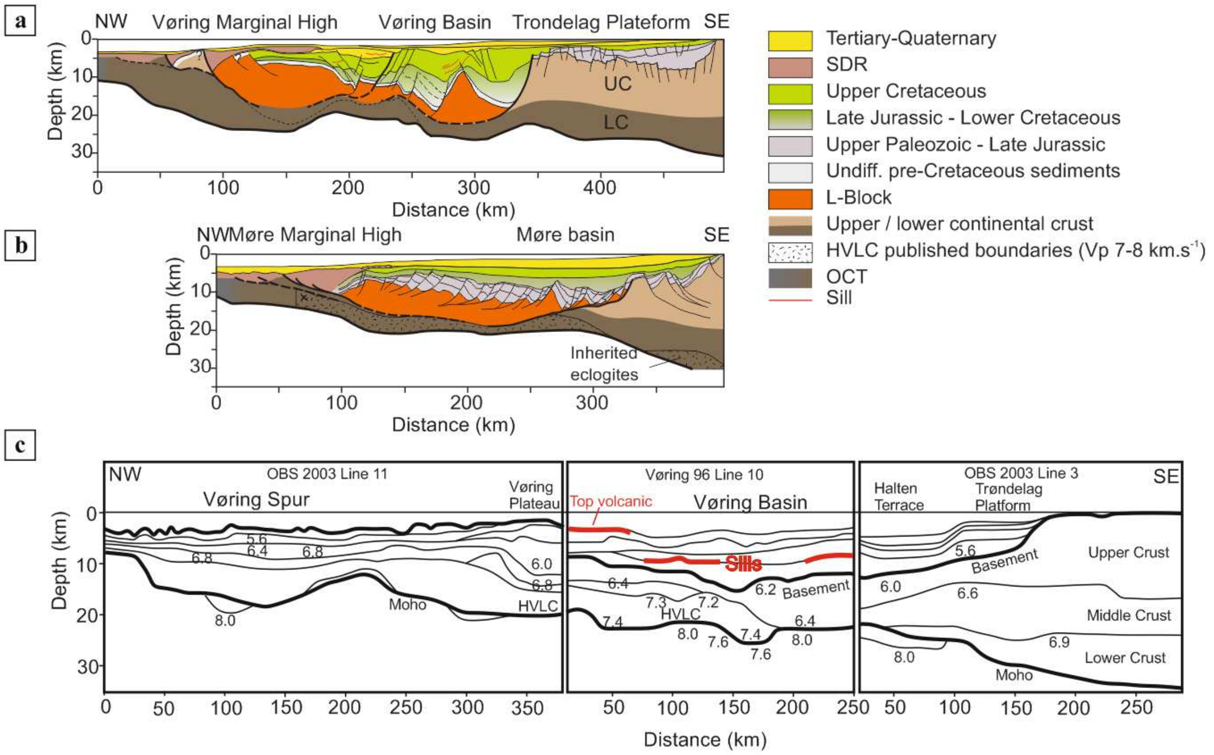

Figure 6. a,b. Schematic interpreted cross sections of the Vøring and Møre rifts and nearby VPMs. Interpretation of Møre Basin is from Theissen-Krah et al. (2017). Interpretation of the Vøring Basin is from Tsikalas et al. (2008) with updated interpretation of the Cretaceous and Tertiary according to Zastrozhnov et al., (2018). c. Vøring crustal scale interpretation from seismic refraction data (reproduced from Funck et al., 2016).

5. Discussion

VPMs development and consecutive plate breakup with syn-magmatic lithosphere thinning and stretching clearly occur without any previous amagmatic stretching and thinning. It is well documented in the South Atlantic (e.g. Blaich et al., 2013; Heine et al., 2013; Stica et al., 2014; Clerc et al., 2015) and in the W-Aden Gulf (Leroy et al., 2010).Diffuse amagmatic continental extension may locally exist before mantle melting (d'Acremont et al., 2005; Franke et al., 2006; Leroy et al., 2010; Watremez et al., 2011; Brune & Autin, 2013) but the narrow and ancient rifts were mostly associated with small lithosphere thinning and/or developed in trends highly oblique to the final VPMs.

We investigated in this paper four cases for which the rifted system combines the structure of an ancient and extinct sedimentary rifts sub-parallel to more distal and younger VPMs. In those cases, a bulk crustal asymmetry of conjugate VPMs may exist, in clear relation with pre-magmatic (or pre-breakup) across-strike gradient in crust thickness (Fig. 2b). This asymmetry is marked, notably, by unequal wideness of the crustal domains with outer SDRs, relatively to the locus of final oceanic breakup (Fig. 2b). The origin of this asymmetry is certainly rheological, in relation with across-strike variations in lower-crust thickness at the time of the syn-magmatic rupturing.

In both the W-Greenland and Laxmi cases, syn-sedimentary extension did not exceed 120% and there was apparently a small timespan between the last pulse of syn-sedimentary extension and the earlier stage of syn-volcanic extension. In those cases (Fig. 7a), a particular upper crustal structure developed over time with the individualization of a buoyant upper crustal block, hereafter called L-Block (L for Lateral), bounded by diachronic detachment faults with opposite dips, dying out, when observable, along the brittle crust/ductile crust transition zone. This whaleback shape upper crustal block is about 60-75 km and 120 km wide in the W-Greenland and Laxmi Basins (Fig. 4c, 5c).

The basin structure within the L-Block is probably very complex with former syn-sedimentary seaward dipping normal faults being rotated to the vertical or, even, to invert dip directions (Fig. 4c), being a geometric consequence of the seaward crustal flexure related to the inner SDR wedges. In both W-Greenland and Laxmi, the whaleback is post-tectonically eroded, suggesting, in addition to the general subaerial setting of lavas from inner SDRs, that this L-Block, as defined, was buoyant and uplifted throughout VPM development (Fig. 7a).

The L-Block as defined above should be distinguished from the H-Block introduced by Lavier & Manatschal (2006), which is associated with amagmatic to poorly magmatic systems. The H-Block develops during an initial stretching stage, where pure shear dominates lithosphere deformation (Lavier & Manatschal, 2006). It is interpreted as the common hangingwall of two synchronous high-angle faults dipping seawards. A change from lithospheric pure shear to simple shear is accompanied with the development of a major detachment fault developing from one of the initial conjugate fault bounding the H-Block (Lavier & Manatschal, 2006). This block becomes part of the narrow passive margin located at the upper plate of the major trans-lithospheric detachment fault which generally exhume the lithosphere mantle. Contrary to the H-Block, the L-Block forms in a diachronous way, in relation with a flip-flop of the extensional faults from the sedimentary to the volcanic stage. It also coincides with the concept of marginal plateau earlier defined by Lister et al. (1986).

In the Vøring and Møre cases (Fig. 6a), further south in the NE-Atlantic (Figs. 2a and 2b), significant extension occurred much before the conjugate VPM development. The Norway rifted margins are distinct from those of W-Greenland and Laxmi because of the time-span between major pre-magmatic extension and syn-magmatic breakup and the amount of pre-magmatic extension which is up to 500% (Fig. 6b; Tsikalas et al., 2008). However, we suggest that an L-Block remnant may still be distinguished in those systems although highly dilacerated due to hyperextension (Fig. 7b). The Late Cretaceous collapse-like extension associated with the Gjallar dome east of the Vøring High is highly interesting in this context. The transient dome-related Cretaceous fault system at the seaward edge of the Vøring Basin tends to individualize a younger and narrower 'secondary' L-Block individualized within the major block (Fig. 6a) with, finally, an overall structure similar to that proposed for the W-Greenland and Laxmi cases.

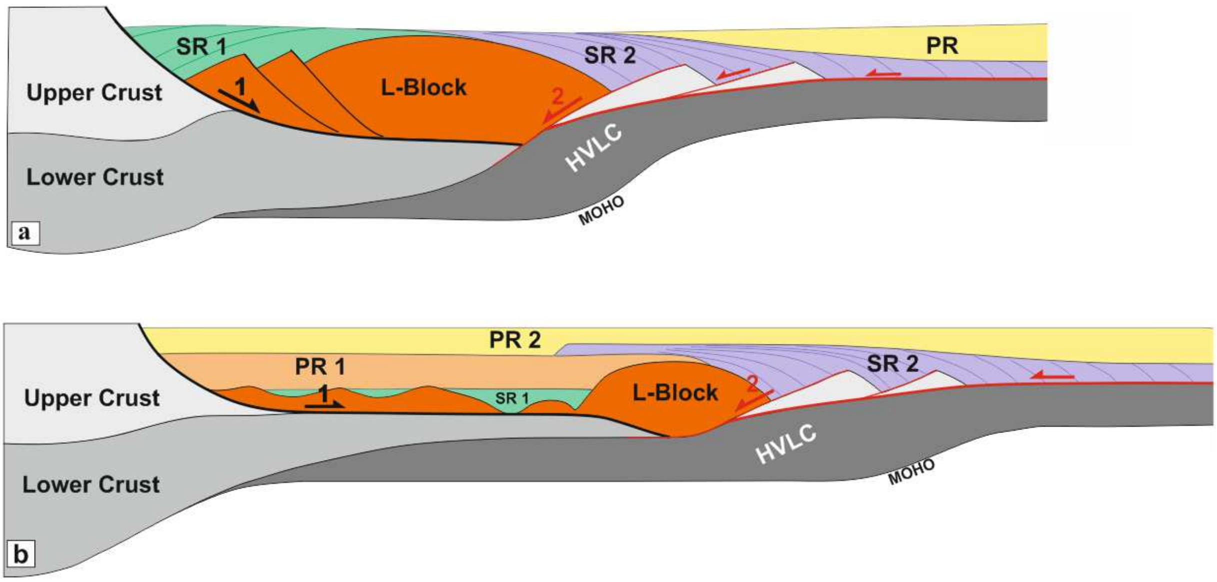

Figure 7. Ocean-continent transition when magmatic breakup follows non-magmatic rifting. a. moderate pre-magmatic extension and short time-span between the sedimentary stage and VPM formation. SR1: syn-rift sediments, SR2: syn-rift volcanics, PR: post-rift sediments; b. hyperextended amagmatic rift and long delay between SR1 and SR2. PR1 relates to post-SR1 and pre-SR2 post-rift sediments due to initial thermal subsidence and PR2 to post-

breakup (post-SR2) thermal subsidence. HVLC: high-velocity lower crust (injected and inherited lower continental crust).

In each of the studied cases, the final structure of the VPM depends on the amount and timing of syn-sedimentary extension regarding the magmatic breakup, due to distinctive tectonic processes from the sedimentary to the volcanic stages. When the time-span between syn-sedimentary and syn-volcanic extension is short (e.g. Laxmi Basin and W-Greenland, Fig. 4c and 5c), the L-Block, is not only narrow but remains buoyant (Figs. 4c, 5c and 7a) suggesting strong thermal gradients and no significant thermal re-equilibration.

When the time-span between syn-sedimentary extension and syn-magmatic reactivation is large (Vøring case, Fig. 6b), the L-Block is wider, dissected and suffers from thermal subsidence. Such amagmatic stretching/thinning is generally associated with moderate rates of extension (about 10-15 to 10-16 s-1) with, consecutively, moderate thermal gradients (Kusznir & Park, 1987) and significant lateral heat dissipation (Alvarez et al., 1984; England & Thompson, 1984). Whatever the amount of initial thinning, renewed extension along a pre-thinned lithosphere and after a long stage of thermal re-equilibration (Fig. 3) means reactivating a stronger lithosphere due to a low ratio between the inherited silicic crust to the total lithosphere thickness (e.g. Steckler & Ten Brink, 1986; Kusznir & Park, 1987). In such cases VPMs will tend to develop away from the initial and cooling rift axis. It is a common observation that the innermost SDRs at VPMs develop seaward of a thick ribbon of continental crust (i.e. more preserved from pre-magmatic extension) and are thus associated with a more deformable and buoyant continental lithosphere.

Thus, the Peron-Pinvidic et al. (2013) model appears to us to be in contradiction with the established fact that, in the NE Atlantic as elsewhere, the VPM-related mantle melting event is not confined to the COB but is generalized to the OCT and thus cannot be considered as a localized magma pulse at the end of hyperextension. The VPM-related magmatism in the NE Atlantic is part of a large igneous province (LIP), the products of which were emitted at very high rates during the Paleocene (> 1 km3/yr, i.e. exceeding the rate of magma production in the world oceanic ridge system; Eldholm & Grue, 1994). Both intrusions and sub-aerial lava extrusions are encountered through or over very large areas, including former Paleozoic and Mesozoic rift systems and unextended basement (e.g. non-rift related). There was a time-focusing of this regional magma production along the VPMs during the Eocene, when SDRs began to form during lithosphere necking. For comparison purpose, it should be noticed again that there is no evidence of demonstrated syn-tectonic volcanism of Jurassic age in the Alps margins or associated with the syn-rift sedimentary formations of the Iberia-Newfoundland margins, apart from very small volumes of alkaline volcanics (Jagoutz et al., 2007). However, in the pre-Tertiary NE-Atlantic realm, significant volcanism during the Late Carboniferous-Permian was documented (Neumann et al., 2004) as well as, occasionally during the Jurassic at the North Sea triple junction (Hendrie et al., 1993). More generally, the Peron-Pinvidic et al. (2013) model is purely crustal and not lithospheric. It involves very low rates of extension, as well as continuous (albeit migrating) stretching and thinning spanning over a 200 Myrs period (if one considers the Triassic to be the beginning of significant lithosphere extension in the NE-Atlantic).

6. Conclusion

a. There should be no straight forward comparison between VPMs and NVPMs margins. Both kinds of margins are associated with extreme lithosphere stretching and thinning, but at crustal scale NVPMs are isovolumic and highly asymmetric. In contrast, VPMs develop coeval with huge mantle melting and magmatic inflation. They consecutively increase in volume during syn-magmatic extension and SDRs formation (Geoffroy et al., 2015).The conjugated VPMs could have distinct wideness and crust thickness, depending on the effect of tectonic inheritance (this paper), but they do not show any tectonic asymmetry (continentward dipping detachments on both sides).

b. Continental breakup may involve several long and discontinuous periods of lithospheric thinning and stretching, especially in mobile inter-cratonic areas. The breakup and onset of oceanic-type lithosphere is often facilitated by syn-magmatic stretching and thinning, i.e. by the development of paired VPMs which develop away (seaward) from ancient rift systems.

c. One cannot understand or evaluate the processes of lithosphere extension at crustal scale only, especially regarding multistage extension. For simple rheological reasons, the amount of lithosphere tectonic thinning at each stage, and the time-span between each thinning episode, exert a first-order control on the localization of the breakup-related passive margins. Time-dependent thermal re-equilibration associated with crustal thinning makes the lithosphere stronger and unable to stretch again or to breakup. This naturally shifts any new deformation to areas where the crust is thicker.

d. Detailed structure of ocean-continent transitions illustrates the early development of upper crustal (L-Block) or whole crustal "blocks" (H-Block and C-Block) of distinct origins in association with synchronous or diachronous fault patterns. Our study defines the marginal plateaus or L-Blocks as being an early characteristic feature of composite VPMs ocean-continent transitions. Whatever their tectonic meaning, all of those blocks consist in less extended and stretched areas forming buoyant and shallow or subaerial plateaus at the time of lithosphere breakup and early oceanic spreading.

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